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MONOGRAFIE DI «NATURA BRESCIANA»

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DI BRESCIA

PERMIAN CONTINENTAL DEPOSITS OF EUROPE AND OTHER AREAS. REGIONAL REPORTS AND CORRELATIONS

Giuseppe CASSINIS (*Editor*)

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Front Cover Explanation

The Permian continental succession in the Val Trompia Basin (Central Southern Alps, Italy). It consists of calcalkaline acidic to intermediate volcanics (A, B, D, H) and alluvial to lacustrine (C, E, F, G) deposits, which are unconformably overlain by the Upper Permian fluvial redbeds of the Verrucano Lombardo (I).
Above, one of the first tetrapod footprints (interpreted as *Amphisauropus latus* by Ceoloni *et al.*, 1987), together with a tail drag. These were probably found in the upper part of member "C" near Malga Cuta (now called "Stabul Maggiore"), approximately two kilometres west of the basin. The discovery was made by curate Giovanni Bruni (1816-1880) (see photograph) on 23 September 1873.

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PREFACE

The papers assembled in this special volume by the Museum of Natural Sciences of Brescia represent the preliminary research on the subject "Late Palaeozoic stratigraphic and structural evolution in Alpine and Apennine sectors. Comparison with Sardinia and other areas of the western Mediterranean", which was co-financed by the Ministry of University and Scientific and Technological Research (MURST) in 1998. In addition to the aforementioned project, other regions, both in Europe and outside, have been included.

The present volume generally focuses only on the continental domains, contrary to the Brescia "Abstracts" which consist of a larger number of papers (60) that also cover marine areas and other topics.

These preliminary results will certainly be followed by others at the second meeting of the above research project, which will be held in Siena, in April/May 2001, when new ideas about the Permian evolution will be highlighted on a wider scale. Consequently, we want to encourage further investigations, *i.e.* related to the continental geology, which is still scarcely known. Moreover, the interpretations arising could indirectly stimulate the research activity of the "Continental Permian Working Group", founded some years ago by the executive board of the IUGS Subcommission on Permian Stratigraphy (SPS).

This foreword is now followed by the transcription of some speeches given by Prof. Paolo Corsini, the Mayor of Brescia, and Prof. Mario Vanossi, Director of the Earth Science Department and Head of the CNR "Alps Group" in Pavia, and Member of the Italian Geological Society Council.

OPENING SPEECHES

Ladies and Gentlemen,

As Mayor of Brescia, I have great pleasure in welcoming you to the Civic Museum of Natural Sciences with the aim of participating in this Congress on the "Continental Permian". As people say, this meeting deals with a research field which stimulates considerable interest among geologists, as that period lies between the Hercynian orogeny and the Alpine cycle, accompanied by paleontological, stratigraphical, petrographical, paleogeographical and geodynamic changes of worldwide importance.

From the first half of the 1900s, the Permian of the Brescian Pre-Alps already attracted the attention of Italian and foreign researchers, and Brescia was also the site of an unforgettable field trip throughout the Permian and the Permian-Triassic boundary of the central-eastern Southern Alps, in July 1986. As a consequence, Brescia and the Town Council are very pleased to receive you again in this Museum, so that your research can be debated, new evidence possibly discovered, and excellent results achieved.

Furthermore, I fully support the activity of this didactic and scientific centre, and the work carried out and still to be developed in other research sectors, along with a number of other organisations. In addition to make your stay in Brescia easier, in agreement with the Director of the Museum, Dr. Marco Tonon, the Curator of Earth Sciences of the Museum, Dr. Paolo Schirolli, and the Organising Committee of this meeting, we have also agreed to publish the proceedings in a special volume, of monographic type, in the local review «Natura Bresciana». I hope that this decision can be interpreted as a sign of our sincerity and sense of collaboration.

Finally, I want to thank everyone who has contributed to the organisation of this meeting, and all who wish to take part in this scientific initiative.

Buon lavoro a tutti!

Prof. Paolo Corsini
Mayor of Brescia

Good morning and welcome to everybody

The research group that I represent is currently known as the "Alps Group", although its complete definition would be "Research Group on the Geodynamics of the Alpine-Apennine System". Founded in the late 1950s, it is one of the oldest groups of the National Council of Research, uniting researchers from ten different universities of northern Italy, in order to share specific competencies and knowledge applied to some main projects: from the pre-Alpine setting, up to the present-day evolution of the chain and its foredeeps.

One of the principal projects is to investigate the Permian-Upper Carboniferous covers of the basements, and to acquire further knowledge about processes (magmatism, sedimentation, tectonics) during the post-collisional Variscan history, as well as timing and mutual relationships. This topic is not only very interesting, but also allows us to verify how and to what extent the pre-Alpine setting influenced location, geometries and evolution of the subsequent Mesozoic continental margins. Moreover, knowledge of the post-collisional evolution of the Variscan chain might provide models that can be used when comparing the younger Alpine chain.

Research on Late Paleozoic history has been carried out for more than 30 years by some of the universities in the Group, particularly the University of Pavia. So, when Prof. Giuseppe Cassinis submitted his programme for this meeting which involved not only the Southern Alps, but also the Sardinia sector, the proposal was greatly appreciated by the "Alps Group".

This Group is much indebted to the organisers of this meeting for the scientific coordination and the realisation of the two field-excursions in Sardinia and the Southern Alps, respectively.

We also wish to thank all the participants, and especially those who will present their work at this Conference. The collaboration of the Museum of Natural Sciences of Brescia is also gratefully acknowledged.

As to my wishes: discuss, fight – if necessary – and, above all, enjoy yourselves.

Don't worry if, sometimes, confusion seems to arise from discussions. As Friedrich Nietzsche, the German philosopher, stated, "one must have chaos inside himself to generate a dancing star". If so, may each one of you have his own brilliant dancing star!

Long life to Permian times and to everyone here. Thank you.

Mario Vanossi
Director of
the Earth Science Dept.,
University of Pavia

1. EUROPE

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LATE PALAEOZOIC-EARLY MESOZOIC PLATE BOUNDARY REORGANIZATION: COLLAPSE OF THE VARISCAN OROGEN AND OPENING OF NEOTETHYS

PETER A. ZIEGLER [1] and GÉRARD M. STAMPFLI [2]

Key words – Variscan orogen; Pangea; Apulia; plate reorganization; wrench tectonics; rifting; back-arc extension; Palaeotethys; Neotethys; Cimmerian orogeny; Carboniferous; Permian; Triassic sedimentation.

Abstract – Late Palaeozoic suturing of Laurussia and Gondwana was accompanied and followed by a major plate boundary reorganization that involved subduction progradation from the Hercynian suture in the interior of Pangea to its peripheries and detachment of the Cimmerian composite terrane from the non-collisional northern margin of Gondwana, entailing Permo-Triassic opening of Neotethys.
During the latest Carboniferous-Early Permian Alleghanian orogeny, a dextral translation of Africa relative to Europe gave rise to the development of a conjugate shear system that transected the Variscan fold belt and its northern foreland. Collapse of the Variscan orogen was accompanied by regional uplift, the subsidence of an array of transtensional and pull-apart basins and widespread magmatism that can be related to the detachment of subduction slabs and possibly mild plume activity. With the Mid-Permian consolidation of the Alleghanian orogen, tectonic and magmatic activity abated in the Variscan domain. Its Late Permian and Triassic evolution was dominated by thermal relaxation of the lithosphere, southward propagation of the Arctic-North Atlantic and westward propagation of the Tethys rift systems.
Back-arc rifting, controlled by roll-back of the Palaeotethys subduction zone, caused Permo-Triassic opening of the oceanic Meliata-Maliak and the Svanetia-Küre-Karakaya system of basins. Late Permian and Triassic opening of the East-Mediterranean branch of Neotethys was accompanied by progressive closure of the western parts of Palaeotethys, culminating in Middle and Late Triassic collision of the Greater Apulia terrane with the Pelagonia block that had been detached from Europe in conjunction with opening of the Meliata-Maliak basin. Similarly, Late Triassic collision of Cimmerian terranes with the eastern, Pontides parts of the Palaeotethys arc-trench system was accompanied by closure of the Svanetia-Küre-Karakaya back-arc basins during the Early Cimmerian orogeny.
In the Variscan domain, Late Palaeozoic-Early Mesozoic continental series can be grossly subdivided into a latest Carboniferous-Early Permian syn-tectonic and a Late Permian-Early Triassic post-tectonic cycle. The latter was diachronously interrupted by the onset of rifting and the Early Cimmerian orogeny. The Tethys rift system provided avenues for Late Permian and Triassic transgressions of the Tethys seas. The Norwegian-Greenland Sea rift system paved the way for the Late Permian Zechstein Sea transgression into Western and Central Europe. The Illawarra magnetic reversal and the end-Permian isotope anomalies provide a-biotic chronostratigraphic markers that take priority over biostratigraphic correlations.

Parole chiave – Orogene Varisico; Pangea; Apulia; riorganizzazione di zolle; tettonica trascorrente; *rifting*; Paleotetide; Neotetide; Orogenesi Cimmerica; Carbonifero; Permiano; sedimentazione triassica.

Riassunto – La formazione e la chiusura della sutura tardo-paleozoica tra Laurussia e Gondwana fu accompagnata e seguita da una riorganizzazione d'ordine maggiore dei limiti delle zolle, che coinvolse la progradazione della subduzione dalla sutura ercinica, posta all'interno della Pangea, verso le sue zone periferiche, nonché il distacco del *terrane* composito cimmerico dal margine settentrionale non collisionale del Gondwana, che comportò l'apertura permo-triassica della Neotetide.
Durante l'Orogenesi Appalachiana, tra la fine del Carbonifero e il Permiano inferiore, una traslazione sinistra dell'Africa rispetto all'Europa generò lo sviluppo di un sistema coniugato di tagli che segmentò la catena a pieghe varisica ed il suo avampaese settentrionale. Il collasso dell'Orogene Varisico fu accompagnato da un sollevamento regionale, dalla subsidenza di un insieme di bacini transtensivi e di *pull-apart*, e da un esteso magmatismo che può essere riferito al distacco di *slab* di subduzione e possibilmente a un'attività di pennacchi a temperatura moderata (*mild plumes*). Con il consolidamento nel Permiano medio dell'Orogenesi Appalachiana, l'attività tettonica e magmatica si affievolì nel dominio varisico. La sua evoluzione tardo-permiana e triassica fu dominata da un rilassamento termico della litosfera, una propagazione verso sud della zona artica-nord atlantica e una propagazione verso ovest dei sistemi di rift tetidei.
Un *rifting* di retro-arco, controllato dall'arretramento (*roll-back*) della zona di subduzione della Paleotetide, determinò l'apertura permo-triassica del sistema di bacini oceanici di Meliata-Maliak e di Svanetia-Küre-Karakaya.
L'apertura tardo-permiana e triassica del ramo mediterraneo orientale della Neotetide fu accompagnata da una progressiva chiusura dei settori occidentali della Paleotetide, culminante nella collisione medio- e tardo-triassica del *terrane* della *"Greater Apulia"* con il blocco della Pelagonia, che iniziò a distaccarsi dall'Europa in concomitanza con l'apertura del bacino di Meliata-Maliak. Analogamente, la collisione tardo-triassica dei *terranes* cimmerici con i settori della Pontide occidentale appartenenti al sistema arco-fossa della Paleotetide fu accompagnato,

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durante le fasi iniziali dell'orogenesi cimmerica, dalla chiusura dei bacini di retro-arco di Svanetia-Küre-Karakaya.
Nel dominio varisico, le serie continentali tardo-paleozoiche e mesozoiche inferiori possono essere grossolanamente suddivise in un ciclo sin-tettonico, intercorrente tra la fine del Carbonifero e il Permiano inferiore, e un ciclo post-tettonico, sviluppatosi tra il Permiano superiore e il Triassico inferiore. Quest'ultimo fu diacronicamente interrotto dall'inizio del *rifting* e dalle fasi iniziali dell'orogenesi cimmerica. Il sistema di *rift* della Tetide aprì vie alle trasgressioni tardo-permiane e triassiche dei mari della Tetide. Così, il sistema di *rift* dei Mari di Norvegia e della Groenlandia portò alla trasgressione tardo-permiana del Mare dello Zechstein nell'Europa occidentale e centrale. L'inversione magnetica nota come Illawarra e le anomalie isotopiche registrate alla fine del Permiano sono indicatori cronostratigrafici abiologici d'importanza prioritaria rispetto alle correlazioni biostratigrafiche.

INTRODUCTION

Late Carboniferous-Early Mesozoic times correspond to a period of global plate boundary reorganization that was presumably also accompanied by a reorganization of the deep mantle convection systems (Ziegler, 1993 a, b; Ziegler *et al.*, in press).

With the Late Carboniferous and Early Permian consolidation of the Variscan and Appalachian-Mauretanides-Ouachita-Marathon orogens, respectively, the Hercynian suturing process of Laurussia and Gondwana came to an end (Ziegler, 1989, 1990). However, during the Permian and Triassic, orogenic activity persisted in the Uralian system, along which Kazakhstan and Siberia were welded to the eastern margin of Laurussia (Zonenshain *et al.*, 1990; Matte, 1995; Nikishin *et al.*, 1996), as well as along the southern margin of Eurasia that was associated with the Palaeotethys subduction zone (Sengör *et al.*, 1984; Ricou, 1995). Locking of the Hercynian sutures in the centre of Pangea was accompanied and followed by accelerated orogenic activity along the American, Antarctic and Australian Panthalassa margins of Pangea (Proto-Cordillera: St. John, 1986; Ziegler, 1989, 1993 a; Visser & Praekelt, 1998). This reflects a first-order subduction progradation from the interior of Pangea to its peripheries. This important plate boundary reorganization was accompanied by a counterclockwise rotation of Pangea, amounting to 20° during the Permian and a further 17° during the Triassic around a pole located in the Gulf of Mexico (Ziegler, 1993 a).

Pangea apparently had an insulating effect on the deep mantle convection systems that were active during its Carboniferous and Early Permian suturing phases, causing the decay of old down-welling cells (Guillou & Jaupart, 1995). Moreover, accumulation of large amounts of subducted cool oceanic lithosphere near the core-mantle boundary apparently had a cooling effect on the outer core, causing changes in its convection pattern and the related geomagnetic field, as evident by the Late Carboniferous-Permian Reverse Superchrone (CPRS; "Kiaman Magnetic Interval"; Irving & Parry, 1963; Eide & Torsvik, 1996), that commenced during the Westphalian C (±310 Ma, Menning, 1995; Menning *et al.*, 1997) and terminated with the early Tatarian Illawarra reversal (265 Ma, Menning, 1995; Benek *et al.*, 1996; Menning & Jin, 1998). During the Late Permian and Triassic, the modern bi-polar deep mantle convection system began to develop, one branch of which welled up under the core of Pangea and now lies beneath Africa (Cadek *et al.*, 1995). This is compatible with the lower Tatarian resumption of frequent magnetic field reversals (Permo-Triassic Mixed Superchrone, PTMS, Menning, 1995) that 15 My later was followed by major mantle-plume activity at the Permo-Triassic transition (251 Ma; Courtillot *et al.*, 1999; Nikishin *et al.*, in press a). The mantle that welled up and radially flowed out beneath the core of Pangea exerted drag forces on the base of its lithosphere. Constructive interference of these drag forces with plate-boundary forces presumably contributed materially to the Mesozoic break-up of Pangea along its Panafrican, Caledonian and Hercynian sutures that was punctuated by repeated mantle-plume activity (Ziegler, 1990, 1993; Pavoni, 1993; Janssen *et al.*, 1995; Courtillot *et al.*, 1999; Ziegler *et al.*, in press).

CONSOLIDATION AND DEMISE OF THE VARISCAN OROGEN

Evolution of the Variscan orogen involved the step-wise accretion of Gondwana-derived terranes to the southern margin of Laurussia and ultimately the Late Devonian-Early Carboniferous collision of the northwestern margin of Africa with Iberia (Ziegler, 1989, 1990; Matte, 1991; Stampfli *et al.*, 1991; Stampfli, 1996, 2000; Tait *et al.*, 1997; Unrug *et al.*, 1999). During the Carboniferous main phases of the Variscan orogeny, the collision front between Gondwana and Laurussia propagated eastward and southwestward in conjunction with progressive closure of the Palaeotethys and Protoatlantic oceans. By Westphalian time, Gondwana had collided with the North American craton whereas to the east the Palaeotethys was still open (Fig. 1; Stampfli, 2000; Stampfli *et al.*, in press). Correspondingly, the western parts of the Variscan orogen were characterized by a Himalayan-type setting (continent-continent collision) whereas its eastern parts, that find their prolongation in the Scythian orogen, remained in an Andean-type setting (continent-ocean collision). Subduction of large volumes of oceanic and continental crust and sediments along a system of subduction zones associated with the Palaeotethys arc-trench system, the boundaries between the different Gondwana-derived terranes involved in the Variscan orogen, and Devonian-Early Carboniferous back-arc basins, accounted for a pervasive Late Devonian to Carboniferous syn-orogenic calc-alkaline I- and S-type intrusive magmatism (Ziegler, 1990; Matte, 1991; Neubauer & von Raumer, 1993; Bonin *et al.*, 1993; Dallmeyer *et al.*, 1995; von Raumer, 1998; Vigneresse, 1999).

The late Visean to Westphalian main phases of the Variscan orogeny involved major crustal shortening and subduction of commensurate amounts of crustal and mantle-lithospheric material; this was accompanied by the lateral escape of internal, relatively rigid blocks, such as the Aquitaine-Cantabrian, Armorican and Bohemian terranes, and the development of intramontane transtensional basins in which Namurian and Westphalian continental clastics, partly coal bearing, accumulated (*e.g.* Ancenis, Laval, Saar, Saale, Pilzen basins; Ziegler 1990; Matte, 1991; Dallmeyer *et al.*, 1995; Onken *et al.*, 1999). By end-Westphalian times, crustal shortening ceased in the western parts of the Variscan orogen, but persisted in the Appalachian-Mauretanides-Ouachita-Marathon orogen until mid-Permian times, controlling the Alleghanian orogeny. During the latest Carboniferous and Early Permian, the convergence of Gondwana and Laurussia changed from an oblique collision to a dextral translation, culminating by Mid-Permian times in a Pangea A2 continent assembly (a Pangea B assembly in not compatible with geological data and cannot account for the Permo-Carboniferous development of the Ouachita-Marathon orogen; Ziegler 1989, 1990; Matte, 1991; Stampfli *et al.*, in press).

During the Stephanian and Early Permian, the Palaeotethys spreading axis was obliquely subducted beneath the eastern parts of the Variscan and the Scythian orogens (Fig. 2; Stampfli, 1996, 2000; Stampfli *et al.*, in press). At the same time, a dextral translation of Africa relative to Europe, amounting to some 300 to 400 km, gave rise to the development of a conjugate shear system that transected the Variscan orogen as well as its northern foreland, partly terminating in pull-apart basins (*e.g.* Oslo graben). Main elements of this shear system are the dextral Teisseyre-Tornquist, Bay of Biscay, Gibraltar-Minas and Agadir fracture zones. The Gibraltar-Minas fracture zone marked to northern termination of the Alleghanian (Appalachian) orogen (Arthaud & Matte, 1977; Ziegler, 1988, 1990; Coward, 1993). Areas located between these four principal shear zones were transected by a conjugate system of subsidiary shears. At the same time, the Norwe-

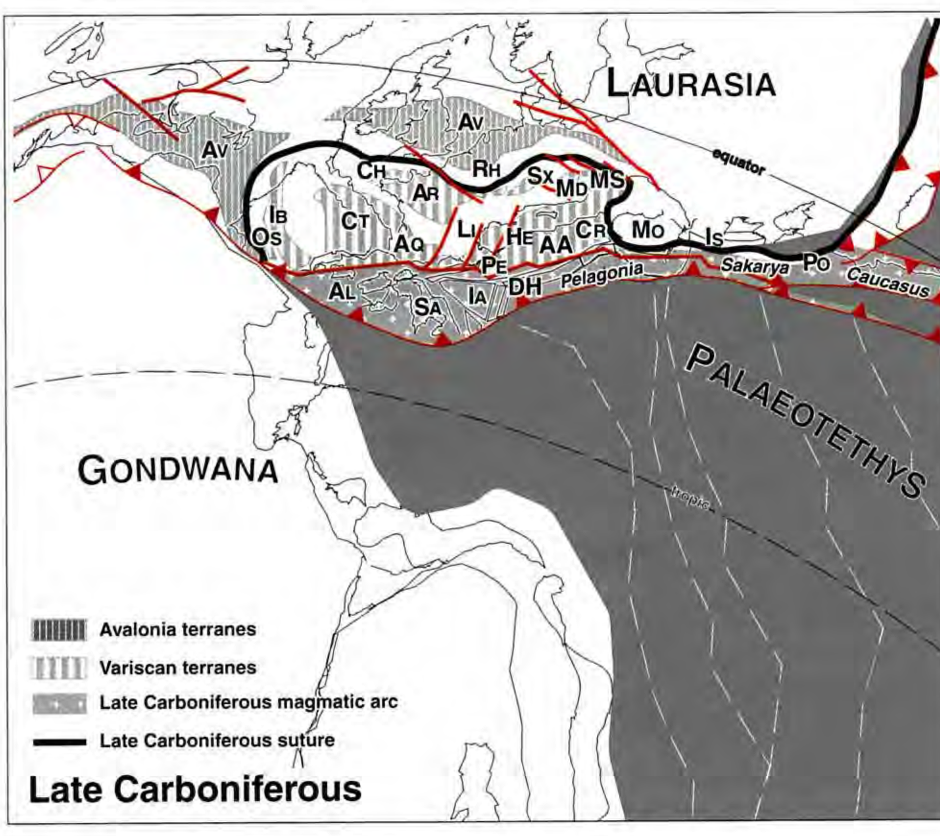

Fig. 1 – Late Carboniferous plate reconstruction, illustrating the Himalayan-type setting of the western parts of the Variscan orogen and the Andean-type setting of its eastern parts. Note subduction of the Palaeotethys sea-floor spreading axis beneath the Scythian orogen (Sakarya-Caucasus domain).

Abbreviations: intra-Variscan Gondwana-derived terranes that were accreted to the southern margin of Laurasia during Silurian to Carboniferous times (shown with different signatures). AA Austroalpine, AL Alboran, AQ Aquitaine, AR Armorica, AV Avalonia, CH Channel, CR Carpathian, CT Cantabria, DH Dinaric-Hellenic, HE Helvetic, IA Intra-Alpine, IB Iberian, IS Istanbul, LI Ligerian, MD Moldanubian, MO Moesia, MS Moravo-Silesia, OS Ossa-Morena, PE Penninic, PO Pontides, RH Rheno-Hercynian, SA South Alpine, SX Saxothuringia.

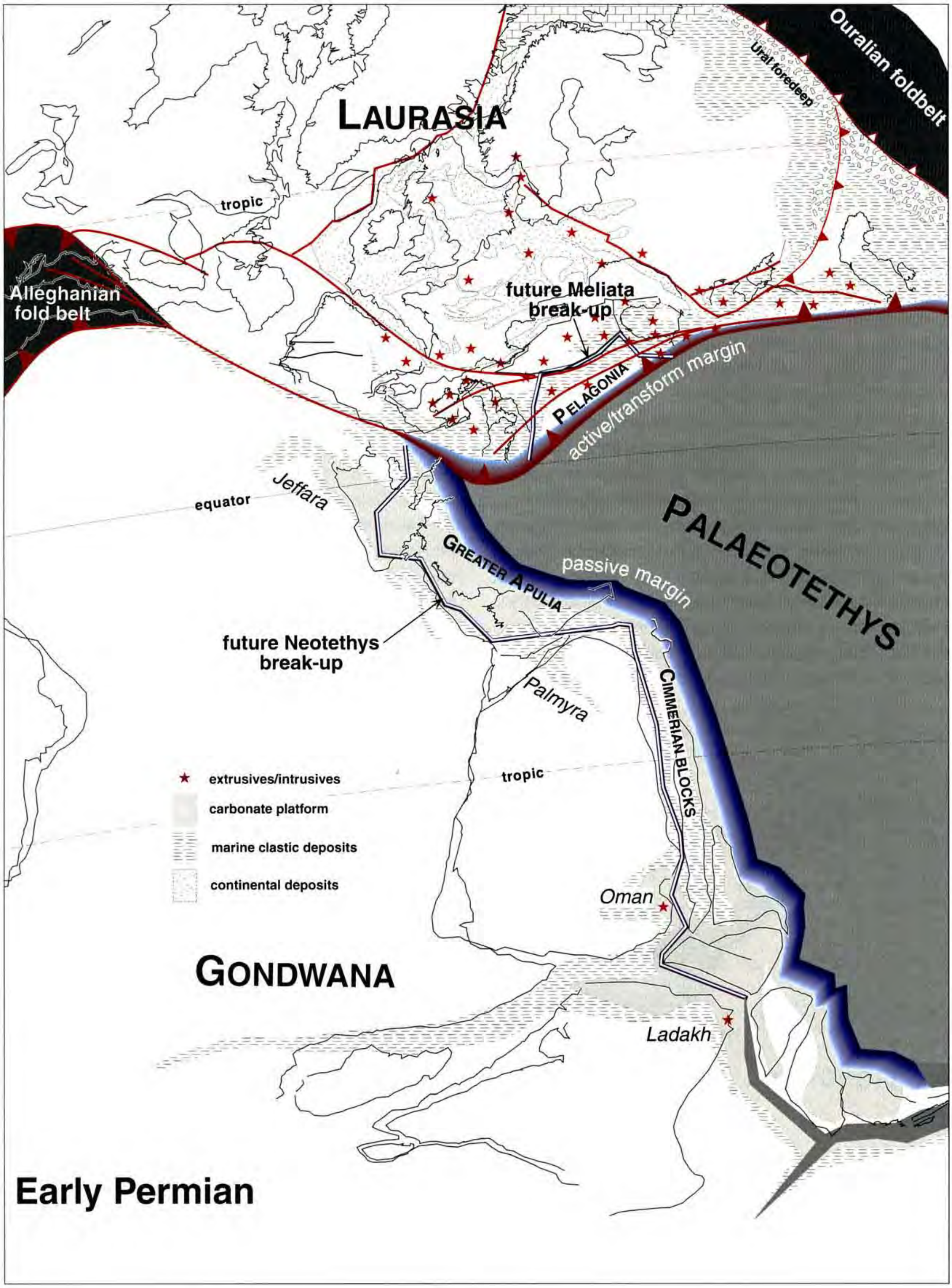

Fig. 2 – Early Permian plate reconstruction, showing trace of Permian rift systems along which the Cimmerian continental terranes were separated from Gondwana and the trace of the Meliata-Maliak back-arc rift system. Palaeogeography shown corresponds to Rotliegend times.

gian-Greenland Sea rift, that had come into evidence already during the Carboniferous main phases of the Variscan orogeny, possibly in response to compressional foreland splitting, was reactivated (Ziegler, 1989, 1990; Smythe *et al.*, 1995). In Western and Central Europe, the Stephanian-Autunian wrench-induced collapse of the rheologically weak, over-thickened crust of the Variscan orogen was accompanied by regional uplift, wide spread extrusive and intrusive magmatic activity that peaked during the Autunian, and the subsidence of an array of multi-directional transtensional trap-door and pull-apart basins in which predominantly continental clastics accumulated (Ziegler, 1990). However, Stephanian marine and transitional marine deposits occur in the wrench-induced basins of Cantabria (Martinez-García & Wagner, 1982) and the South Alpine-Mediterranean domain (Cassinis *et al.*, 1992; Cassinis, 1996). Moreover, also the Stephanian red beds of North Germany contain occasional marine intercalations (Hedemann *et al.*, 1984). On the other hand, throughout the Variscan domain, Early Permian (Autunian) deposits are developed in an entirely continental facies, reflecting its progressive regional uplift. The only exceptions are the South Alpine and Dinaric areas that fringed the evolving Meliata-Maliak back-arc ocean and the Sicilian area that was located on the Palaeotethys shelf (see below; Catalano *et al.*, 1991; Cassinis *et al.*, 1992; Schönlaub, 1993; Cassinis, 1996).

Microtectonic analyses in the Massif Central area indicate for Stephanian-Autunian times a progressive rotation of the principal horizontal compressional stress trajectories from N-S to E-W (Blès *et al.*, 1989). Basins that developed during these times show a complex, polyphase structural evolution, including a late phase of convergent wrench deformation (*e.g.* St. Etienne, Décazeville, Alès basins). Similar stress rotations probably governed the evolution of the wrench-induced Stephanian-Autunian basins in Central Europe and the Alpine domain, development of which generally terminating with a pulse of basin inversion and deep erosion (*e.g.* Saar, Saale and Boscovice basins: see Ziegler, 1990; Aiguilles Rouges basin: Pilloud, 1991; Badertscher & Burkhard, 1998).

Regarding dynamic controls on the post-orogenic collapse of orogens, it must be realized, that deviatoric tensional body forces inherent to the over-thickened lithosphere of an orogen can only come to bear after convergence has ceased and its subducted lithospheric slab(s) has (have) been detached (Fleitout & Froidevaux, 1982; Bott, 1990, 1993). Moreover, post-orogenic thermal re-equilibration of crustal root zones (cooling during rapid subduction of cold foreland lithosphere, followed by post-orogenic heating), involves retrograde metamorphism of eclogites into less dense granulites (Bousquet *et al.*, 1997; Le Pichon *et al.*, 1997). This causes further isostatic uplift of the orogen, thus enhancing its body forces. However, body forces inherent to abandoned orogens are, on their own, unlikely to be sufficiently large to drive apart major cratonic blocks, such as the constituents of a Pangea-type mega-continent (Ziegler, 1993 a). Therefore, it is likely that body forces inherent to the Variscan orogen played only a secondary role during its predominantly wrench-induced Permo-Carboniferous collapse that coincided with the Appalachian suturing phases of Pangea (Ziegler, 1989, 1990; Henk, 1999).

The model of the Cenozoic Basin-and-Range Province has been repeatedly invoked for the Stephanian-Autunian disintegration of the Variscan orogen (Lorenz & Nicholls, 1976, 1984; Jowett & Jarvis, 1984; Ménard & Molenar, 1988; Malavieille, 1993). Despite similarities in the tectonic setting of the Basin-and-Range Province and the Permo-Carboniferous basin system of Europe in terms of subduction of the East Pacific Rise (Verrall, 1989; Parsons, 1995) and the Palaeotethys spreading axis (Stampfli, 1996, 2000), respectively, there are major differences between these two provinces (Ziegler, 1990). The Basin-and-Range Province, which is superimposed on the Andean-type Cordillera, is dominated by a regularly spaced pattern of linear, sub-parallel, partly anastomozing half grabens and intervening horsts; furthermore, it is characterized by frequent core complexes, high strain rates and a large regional extension factor (ß=1.4-1.6; Bally & Snelson, 1980; Eaton, 1982; Coney, 1987; Hamilton, 1987; Oldow *et al.*, 1989; Wernicke, 1992; Parsons, 1995; Westaway, 1999). In contrast, the Permo-Carboniferous basins of Europe are superimposed on the Himalayan-type western parts of the Variscan orogen, as well as on its eastern parts that had remained in an Andean-type setting. These basins are multi-directional, have generally a complex architecture that resulted from syn- and post-depositional transtensional and transpressional deformation, and are closely associated with wrench faults, some of which extend well beyond the northern margin of the Variscan thrust belt (Ziegler, 1990; Bard, 1997; Cassinis *et al.*, 1992, 1997; Cassinis 1996). Moreover, Stephanian and Autunian transtensional and transpressional wrench tectonics gave only locally rise to the development of pull-apart basins and the uplift of core complexes (*e.g.* Massif Central: Malavielle *et al.*, 1990; Burg *et al.*, 1994; Black Forest: Eisbacher *et al.*, 1998). On a regional scale, these wrench tectonics were associated with relatively low crustal stretching factors, as evident, for instance, in the area of the Southern Permian Basin that is mainly located in the Variscan foreland but encroaches in its eastern parts on the Variscan fold-and-thrust belt (van Wees *et al.*, 2000). Nevertheless, and similar to the Basin-and-Range Province (Jones *et al.*, 1992; Parsons, 1995), also the Permo-Carboniferous wrench tectonics of Europe were accompanied by a widespread extrusive and intrusive, mantle-derived alkaline magmatism that shows evidence of strong crustal contamination (Ziegler, 1990; Bonin, 1990; Bonin *et al.*, 1993; Neumann *et al.*, 1995; Marx *et al.*, 1995; Benek *et al.*, 1996; Cortesogno *et al.*, 1998; Breitkreuz & Kennedy, 1999). Melt generation was probably related to localized divergent wrench-induced decompressional partial melting of the uppermost asthenosphere and the lithospheric thermal boundary layer, combined with upwelling of the asthenosphere in response to slab detachment. This was apparently coupled with a rise in the potential temperature of the asthenosphere, possibly related to the impingement of a not very active mantle plume on the base of the lithosphere. Supporting evidence comes from the isotopic signature of the most primitive melts (M. Wilson, pers. comm. 1998). Crustal-scale fractures provided avenues for magma ascent to the surface.

Synorogenic uplift and exhumation of the Variscan internides, up to formerly mid-crustal levels, commenced already during the late Visean and Namurian, partly in conjunction with strike-slip movements related to escape tectonics and the ensuing development of intramontane Namurian (*e.g.* Alpine Zone Houillière: Cortesogno *et al.*, 1993, 1998) and Westphalian neo-autochthonous continental basins (Ziegler, 1990; Henk, 1995, 1999). However, regional uplift of the entire orogen and its foreland began only after crustal shortening had ceased at the end of the Westphalian. Stephanian-Autunian uplift and erosional, as well as tectonic unroofing of the Variscan orogen, in many areas to formerly mid-crustal levels (Burg *et al.*, 1990; Vigneresse, 1999), can be related to a combination of wrench deformation, heating of crustal roots, involving eclogite to granulite transformation (Bousquet *et al.*, 1997; Le Pichon *et al.*, 1997), detachment of subduction slabs, and upwelling and partial melting of the asthenosphere, causing thermal attenuation of the mantle-lithosphere and magmatic inflation of the lithosphere. Mantle derived basic melts that ascended to the base of the crust underplated it, inducing crustal anatexis, fractional crystallisation and the intrusion of granitic to granodioritic-tonalitic melts into the crust (Cortesogno *et al.*, 1998; Breitkreuz & Kennedy, 1999). A unique opportunity to study these processes is offered by the Ivrea Zone where Early Permian mantle-derived (asthenospheric) mafic intrusions caused partial melting of metasediments at basal crustal levels, the ascent of granitic magmas to middle and upper crustal levels, and the development of volcanic activity in contemporaneous wrench-induced basin (Schmid, 1993; P. Brack, pers. comm. 1999). Both retrograde metamorphism of eclogitic roots and the interaction of mantle-derived basic melts with the felsic lower crust contributed to a re-equilibration of the Moho at depth ranges of 30 to 40 km and locally less. By Mid-Permian times, some 40 My after consolidation of the Variscan orogen, its former crustal roots had disappeared.

Permo-Carboniferous wrench tectonics and magmatic activity abated at the transition to the Late Permian, in tandem with the consolidation of the Appalachian orogen (Ziegler, 1989, 1990; Marx *et al.*, 1995).

In view of the above, we feel that kinematics underlying the development of the Cenozoic Basin-and-Range Province in the Cordilleran domain and of the Permo-Carboniferous wrench system transecting the Variscan orogen differ fundamentally to the end that models developed for one of these provinces cannot be indiscriminately extrapolated to the other. Yet, in both cases, ancillary processes inherent to the post-orogenic evolution of an orogen, such as gravitational instability of its over-thickened lithosphere (Dewey, 1988; Braun & Beaumont, 1989), steepening and detachment of subduction slabs, followed by upwelling of the asthenosphere and a mantle derived magmatism, causing thermal thinning of the lithosphere, an increase in heat flow and a commensurate lowering of crustal viscosities, probably contributed to the rapid disintegration of both mountain systems (Coney, 1987; Dilek & Moores, 1999). Subduction of active spreading axes presumably played an important role in slab-detachment beneath the Basin-and-Range and the eastern Variscan and Scythian domains. On the other hand, dextral transform motions between Africa and Europe probably contributed to the rapid detachment of subducted slabs associated with the Himalayan-type western and central parts of the Variscan orogen.

OPENING OF THE NEOTETHYS AND THE MELIATA-MALIAK OCEANS

Following subduction of the Palaeotethys sea-floor spreading axis beneath the eastern Variscan and Scythian orogens, increasingly older and mechanically stronger oceanic lithosphere was subducted. Correspondingly, increasingly larger slab-pull forces were exerted on the non-collisional northeastern margin of Gondwana. Moreover, it is likely that the mantle that welled up and radially flowed out beneath Africa exerted drag forces on the base of its lithosphere. Constructive interference of these mantle drag and slab-pull forces presumably controlled the development of a system of Late Carboniferous-Early Permian rifts along the northeastern peripheries of Africa and Arabia that culminated in the Mid-Permian detachment of the ribbon-like, composite continental Cimmerian terranes from Gondwana (Fig. 3). Continued northward subduction of Palaeotethys beneath the southern margin of Eurasia accounted for the gradual northward migration of these terranes and the Late Permian and Triassic opening of Neotethys (Sengör *et al.*, 1984; Robertson *et al.*, 1996; Stampfli, 1996, 2000; Stampfli *et al.*, 1991 and in press).

Moreover, steepening and roll-back of the Palaeotethys subduction zone, that dipped northwards beneath the eastern Variscan and Scythian orogens, is thought to have controlled the Late Permian and Triassic opening of the oceanic Meliata-Maliak (Kozur, 1991; Stampfli *et al.*, 1991), Crimea-Svanetia (Nikishin *et al.*, in press b), Karakaya (Okay & Mostler, 1994) and Küre (Ustaömer & Robertson, 1994, 1997) back-arc basins (Stampfli, 2000, Stampfli *et al.*, in press; Ziegler *et al.*, in press).

LATE PERMIAN AND TRIASSIC RIFTING AND LITHOSPHERIC RE-EQUILIBRATION

During the Late Permian, and particularly the Triassic, the Norwegian-Greenland Sea rift propagated southwards into the North Atlantic and ultimately the Central Atlantic domain. During the Triassic, the Tethys rift system, that can be related to the opening of the Meliata-Maliak and Neotethys oceans, propagated westwards and interfered

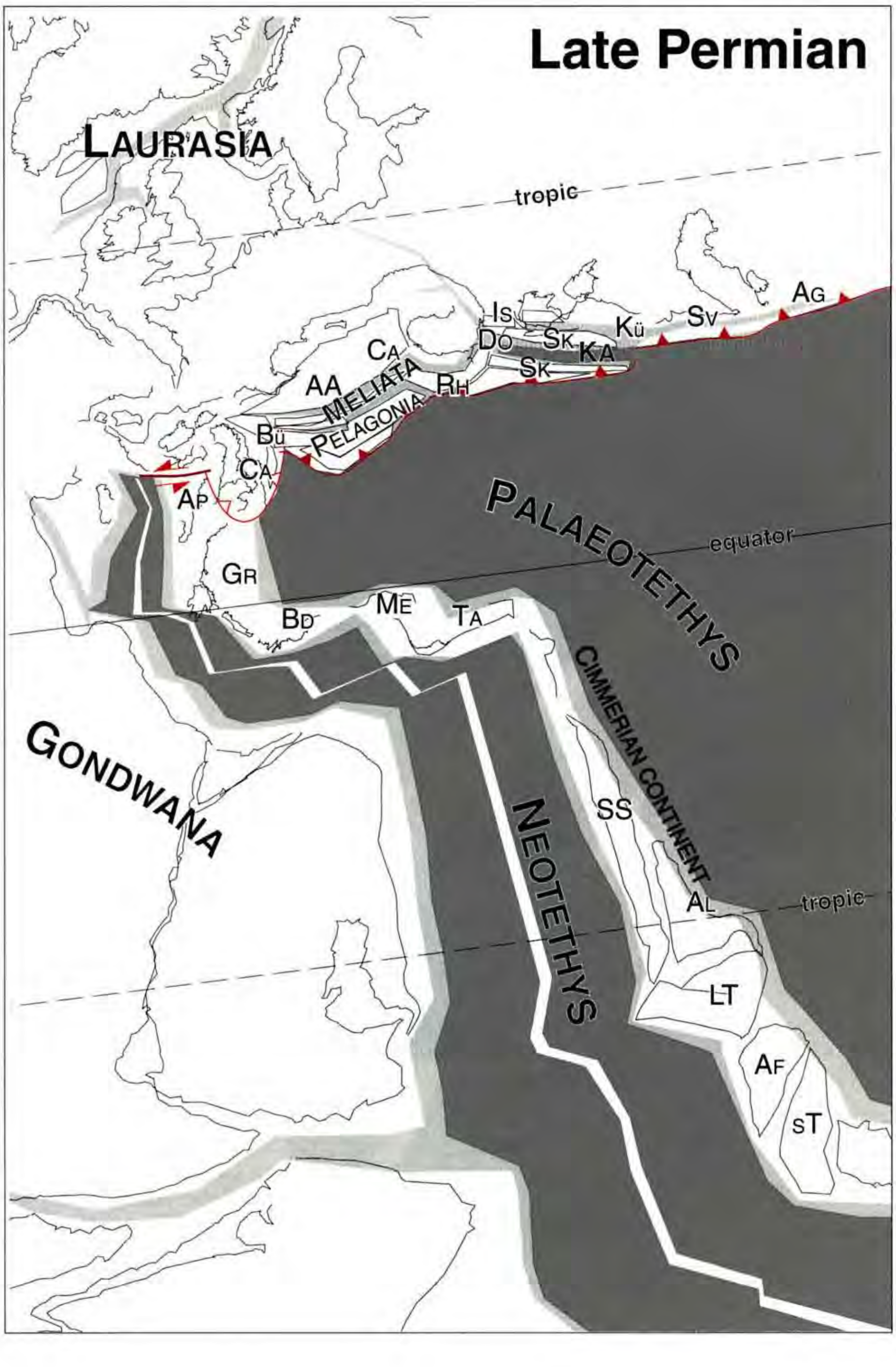

Fig. 3. – Late Permian plate reconstruction, showing opening of Neotethys, detachment of the Cimmerian terranes from the northern margin of Gondwana and opening of back-arc basins along the Eurasian active margin.

Cimmerian Terranes: AP Apulia, GR authochtonous Greece, BD Bey-Daglari, ME Menderes, TA Taurus, SS Sanandaj-Sirjan, AL Alborz, LT Lut-Tabas, AF Central Afghanistan, ST South Tibet.

Eurasian margin: BÜ Bükk, AA Austroalpine, CA Carpathians, RH Rhodope, DO Dobrogea, IS Istanbul, SK Sakarya, EP East Pontides, PL Pelagonia.

Marginal basins: AG Agh-Darband, CR Crimea, KA Karakaya, KÜ Küre, SV Svanetia.

with the Norwegian-Greenland Sea rift system in the Northwest African-Iberian-North Atlantic area. In the process of this, Western and Central Europe, including the Variscan domain, were transected by a complex, multidirectional graben system, some elements of which are superimposed on Permo-Carboniferous fractures (Ziegler, 1988, 1990; Stampfli *et al.*, 1991; Stampfli & Marchant, 1997). Whereas the Permo-Triassic Arctic-North Atlantic and the Triassic West and Central European rifts are essentially non-volcanic, the Late Permian and Triassic Tethyan rifts are characterized by considerable magmatic activity, related to partial melting of the lithospheric thermal boundary layer and upper asthenosphere in response to lithospheric extension (Pamic, 1984; Ziegler, 1988, 1990; Bonin *et al.*, 1987; Bonin, 1989, 1990).

Starting in late Early to early Late Permian times, areas that were not affected by rifting, began to subside in response to the decay of lithospheric thermal anomalies that

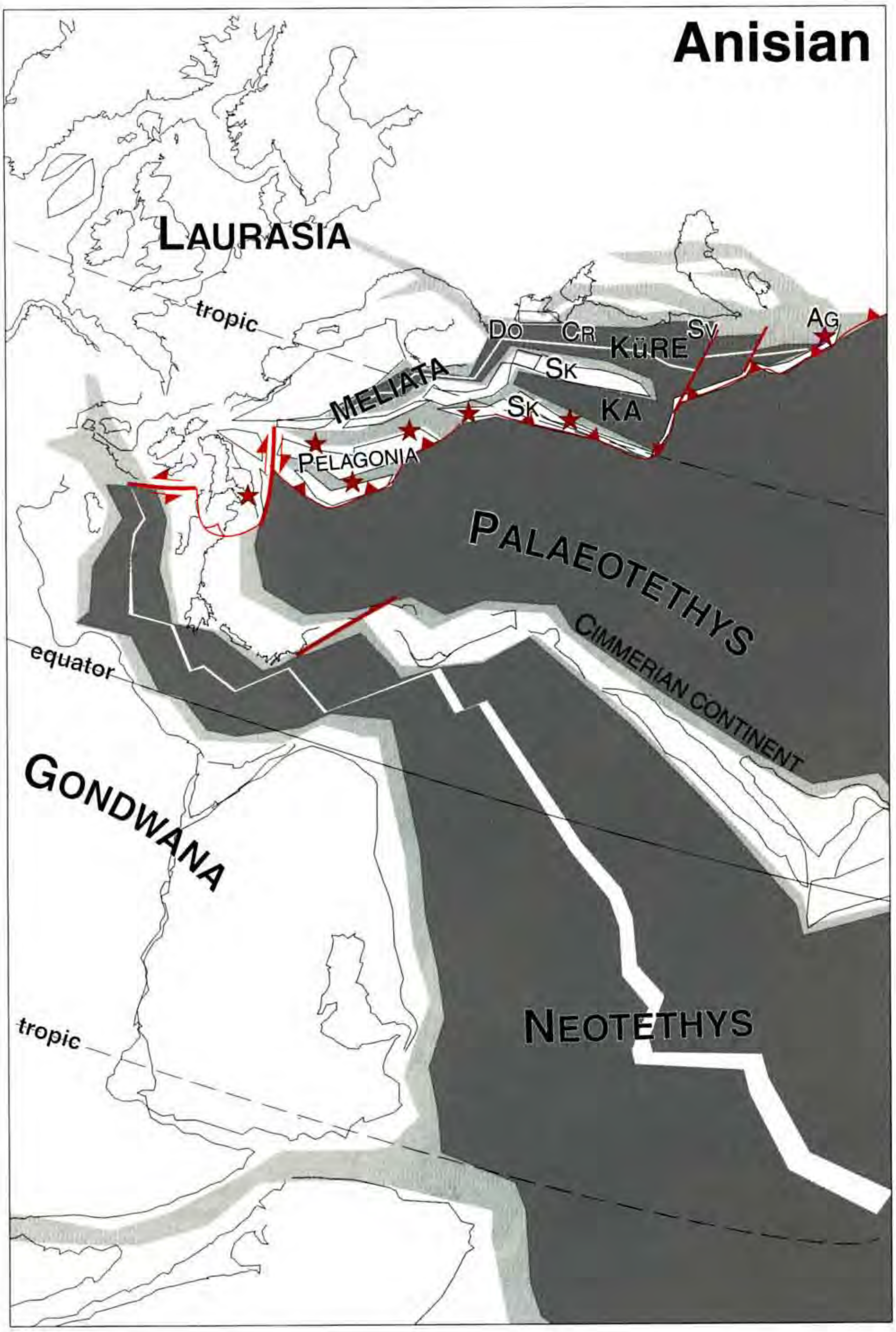

Fig. 4 – Anisian plate reconstruction, illustrating progressive closure of Palaeotethys, opening of the Neotethys and Meliata-Maliak oceans, and initial collision of the Apulia-Greek-Bey-Daglari block with the Pelagonian block and of Cimmerian blocks with the Pontides arc-trench system.

were introduced during the Permo-Carboniferous tectono-magmatic cycle. This is clearly evidenced by the evolution of the Northern and Southern Permian basins that are located in the northern foreland of the Variscan orogen and partly encroach on the latter. In these basins, the continental Upper Rotliegend clastics and the marine Zechstein carbonate-evaporite series accumulated under tectonically rather quiescent conditions (Ziegler, 1990; Plein, 1995; Koronovski, 1999; van Wees *et al.*, 2000). Although the Northern and Southern Permian basins were transected during the Triassic by the North Sea and the Polish-Danish rift systems, their thermal subsidence persisted at least up to the end of the Early Jurassic (Ziegler, 1990; van Wees *et al.*, 2000). Similarly, in the South Alpine domain, Early Permian continental syn-rift series, confined to a system of intramontane transtensional basins, were broadly overstepped by a Late Permian-Early Triassic post-rift sequence consisting of clastics and carbonates. However, this post-rift cycle terminated with the Middle Triassic onset of a new rifting cycle (Bertotti *et al.*, 1993; Cassinis *et al.*, 1997). On a regional scale, progressively larger areas of the deeply truncated Variscan orogen subsided thermally below the erosional base level and were transgressed by the eustatically rising Triassic and Jurassic seas (*e.g.* Paris Basin, South German Franconian Platform; Ziegler, 1990; Schumacher *et al.*, 1999).

CLOSURE OF PALAEOTETHYS AND EARLY CIMMERIAN OROGENY

Late Permian and Triassic opening of the oceanic Neotethys and the Meliata-Maliak back-arc basin had major repercussions on plate kinematics in the western Tethyan area. Progressive closure of the remnant Palaeotethys culminated in the Middle Triassic initial collision of the Cimmerian Apulian-Greek terrane with the Variscan deformed Pelagonia block that had been separated from Europe in conjunction with the opening of the Meliata-Maliak back-arc ocean (Fig. 4). Their suturing gave rise to the "Montenegrian" orogenic pulse of the Southern and Carnic Alps (Brandner, 1984; Castellarin *et*

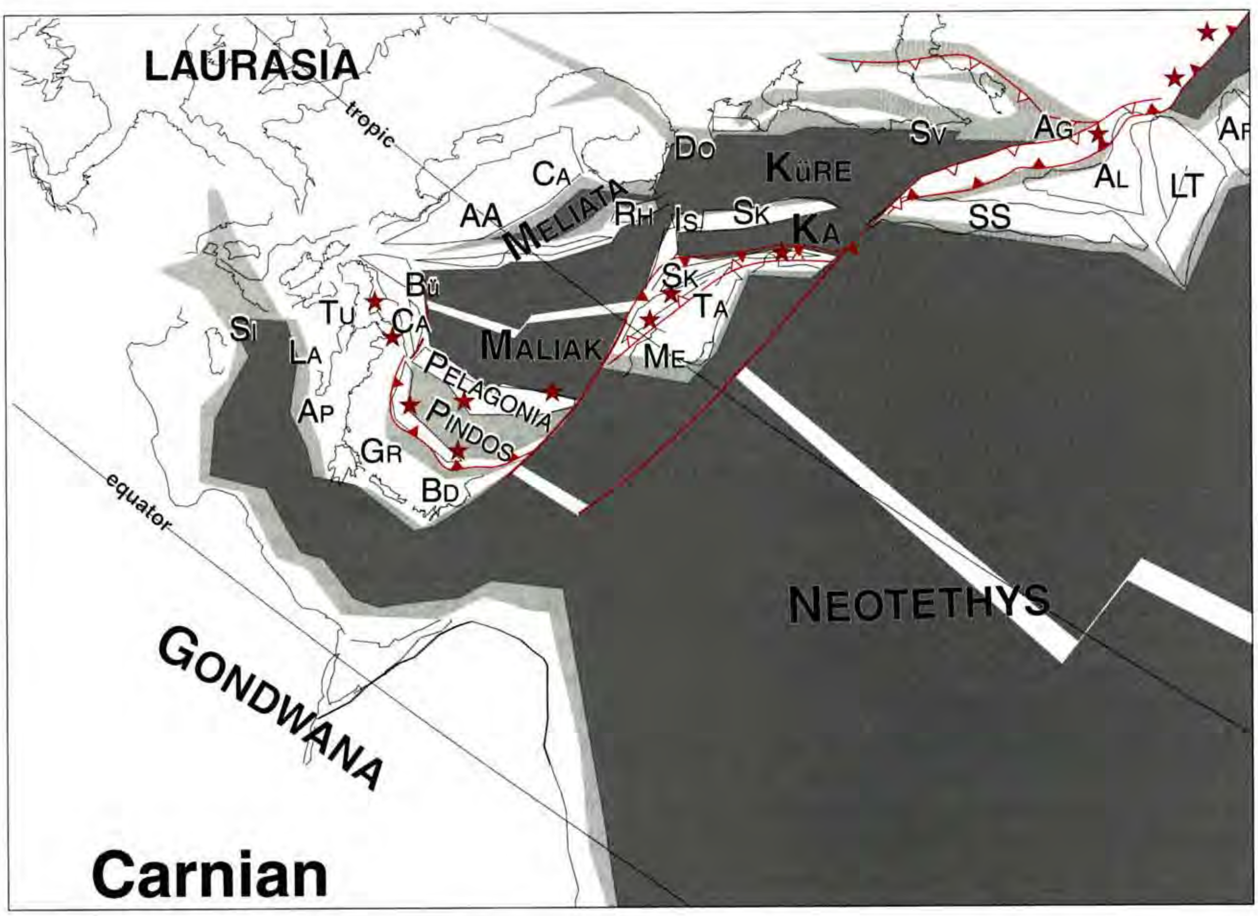

Fig. 5 – Carnian plate reconstruction, illustrating Early Cimmerian accretion of Cimmerian terranes to the Pelagonia block and the Pontides arc-trench system and resulting back-arc compression in the Black Sea domain. The Palaeotethys suture is shown by a broken barbed line. For *abbreviations* see Fig. 3; additional abbreviations: CA Carnic, LA Lagonegro, SI Sicanian, TU Tuscan.

al., 1988), Dinarides, Hellenides and Taurides (Stampfli *et al.*, 1998) that was accompanied by the development of a late Anisian to early Carnian subduction and slab detachment related calc-alkaline to shoshonitic magmatism (Bonadiman *et al.*, 1984; Castellarin *et al.*, 1988; Pe-Piper, 1998; Brack *et al.*, in press), that extended all along the Palaeotethys suture from northern Italy to western Turkey (Stampfli, 2000; Stampfli *et al.*, in press). Moreover, Late Triassic collision of Cimmerian terranes with the Pontides part of the Palaeotethys arc-trench system gave rise to the Cimmerian orogeny during which the Svanetia and Karakaya-Küre back-arc basins were closed (Fig. 5; Nikishin *et al.*, in press b; Stampfli *et al.*, in press).

LATE PALAEOZIC-EARLY MESOZOIC CONTINENTAL SEDIMENTATION

In conjunction with the counter-clockwise rotation of Pangea, the area of Western and Central Europe drifted during the latest Carboniferous out of equatorial latitudes into the northern trade wind belt (Ziegler, 1993 a). At the same time, post-orogenic uplift affected the Variscan domain and its forelands, causing a regional regression. Early Permian continental red beds and subordinate lacustrine deposits, partly overlaying Stephanian coal measures, accumulated under increasingly arid conditions in often isolated intra- and perimontane transtensional and pull-apart basins that evolved in response to Permo-Carboniferous wrench tectonics. Coeval volcanic activity played an important role (Ziegler, 1990). Following termination of the Early Permian tectonomagmatic activity, thermal contraction of the lithosphere governed the subsidence of the Northern and Southern Permian basins in the Variscan foreland; by the end of the Rotliegend these basins had subsided under landlocked conditions below the global sea level (van Wees *et al.*, 2000). Elsewhere, time equivalent strata accumulated in tectonically silled intramontane basins that gradually expanded in response to progressive thermal subsidence of the Variscan crust and the degradation of its palaeorelief. Rifting activity opened the Arctic sea way, through which the Late Permian Zechstein Sea invaded the Northern and Southern Permian basins, as well as a temporary Tethys sea way that extended during the Zechstein 1 cycle from the Svanetia basin, presumably via Dobrogea, into the eastern parts of the Southern Permian basin (Ziegler, 1988, 1990). The low stand in sea level at the Permo-Triassic boundary (Ross & Ross, 1988) induced a regional forced regression, causing in slowly subsiding basins the development of a regional unconformity (Cassinis, 1996). Continental conditions prevailed during the Early Triassic rise in sea level due to a clastic over-supply, compensating for continued thermal and/or rift-induced basin subsidence. This can be related to the development of an efficient drainage system through which erosion products were transported from Scandinavia and the remnant Variscan highlands into the continuously thermally subsiding and gradually expanding Northwest European basin, as well as into the evolving system of rifted basins. Rifting, related to opening of the Neotethys, Meliata-Maliak and Svanetia basins, provided avenues for the Late Permian and particularly the Middle Triassic transgression of the Tethys seas into the West Mediterranean and ultimately into the West and Central European domains, controlling the diachronous termination of continental sedimentation (Ziegler, 1988, 1990, Cassinis *et al.*, 1992; Cassinis, 1996; Marcoux & Baud, 1995; Stampfli *et al*, in press). Communications between the Tethys and Arctic seas were only established during the Early Jurassic (Ziegler, 1988, 1990).

The Late Carboniferous to Early Triassic sedimentary and volcanic series of the Variscan domain can be subdivided into several tectono-stratigraphic sequences that are, however, only grossly correlative. In essence, a latest Carboniferous-Early Permian syn-tectonic and a Late Permian-Early Triassic post-tectonic mega-sequence can be recognized (Cassinis *et al.*, 1992; Krainer, 1993).

Syn-Tectonic Cycle 1

On top of orogenically deformed and erosionally truncated Palaeozoic series, continental and partly marine sedimentation commenced variably during the Westphalian to Early Permian in wrench-induced transtensional and pull-apart basins. The highly variable geometry and internal architecture of these basins, partly related to an alternation of transtension and transpression giving rise to repeated breaks in sedimentation, reflect that many of them evolved under changing stress fields. Such stress field changes may be of a regional as well as a more local nature, with the latter being related to the interaction of a mosaic of crustal blocks delimited by wrench faults under increasing strain (Blès *et al.*, 1989; Burg *et al.*, 1990). Syn-depositional volcanic activity commenced during the Stephanian, reached a peak during the Early Permian and abated gradually during the Late Permian. The frequently observed break in sedimentation at the end of the Early Permian, often associated with variable degrees of basin inversion, reflects a last pulse of wrench deformation that accompanied the terminal phase of the Appalachian orogeny.

Post-Tectonic Cycle 2

During the Late Permian, large areas commenced to subside under a tectonically quiescent regime in response to thermal contraction of the lithosphere, thus defining the second tectono-stratigraphic mega-sequence that extends into Triassic and Jurassic. Generally, this "post-tectonic" sequence broadly overstepped the cycle 1 "syn-tectonic" basins. However, this post-tectonic cycle is not everywhere well expressed as the onset of new rifting activity, related to opening of the Neotethys and the Meliata-Maliak back-arc basin commenced diachronously during Late Permian to Middle Triassic times (Ziegler *et al.*, in press). Similarly, the onset of rifting activity related to southward propagation of the Arctic-North Atlantic rift system is diachronous and ranges from Late Permian to Middle Triassic (Ziegler, 1988, 1990). Furthermore, Middle and Late Triassic compressional tectonics, related to the collision of Cimmerian terranes with the Palaeotethys subduction system, account for the interruption of the post-tectonic cycle 2 in the South and Carnic Alpine, Dinarides, Hellenides, Taurides and Black Sea domains (Nikishin *et al.*, in press b; Stampfli *et al.*, in press).

Correlation of Late Permian-Early Triassic continental and marine strata has to contend with major uncertainties, both at regional and global scales, due to the climatic zonation of continental biota (both latitude and elevation controlled) and the distinct provinciality of shallow marine faunas (Kozur, 1998). This provides for uncertainties in geohistory analyses of continental basins, as well as in palaeogeographic/palaeotectonic-reconstructions.

In this respect, two a-biotic, globally valid markers can play a crucial role in assessing the validity of purely biostratigraphic correlations. The first marker is the Illawarra magnetic reversal (265 Ma) that marks the end of the CPRS and the onset of the PTMS (Menning, 1995). Occurrence of this reversal in Late Permian lower Tatarian strata of Eastern Europe and in Middle Permian Guadalupian ones of North America (Menning & Jin, 1998), illustrates the problematic nature of standard biostratigraphic correlations. The second, albeit less sharp marker is provided by the $\delta^{18}O$, $\delta^{13}C$, $\delta^{34}S$ and $^{87}Sr/^{86}Sr$ isotope anomalies which straddle the Permian-Triassic boundary that is associated with major biotic extinction events (251 Ma; Erwin, 1993; Stanley & Yang, 1994; Faure *et al.*, 1995; Kozur, 1998; Yin & Tong, 1998; Atudorei, 1999). This marker reflects a period of some 1 My during which several severe global environmental crises occurred. Their cause is seen in the massive extrusion of mantle plume-related trap basalts (Siberian Tunguska Province, Emeishan traps of southwestern China; Courtillot *et al.*, 1999; Nikishin *et al.*, in press a) and a voluminous subduction-related volcanism along the Proto-Cordillera and Proto-Altaids (Faure *et al.*, 1995), that gave rise to an increase in CO_2 pressure, an aerosol-induced reduction in solar irradiation and resulting temperature decreases ("volcanic winter" scenario, Kozur, 1998).

Where ever possible, both of these markers should be used to chronologically constrain biostratigraphic zonations and to firm up time scales applied in quantitative subsidence analyses.

PERMO-TRIASSIC MARINE SEDIMENTS AND TECTONICS OF THE WEST-TETHYAN DOMAIN

In the South-European domain we have to deal with several marine basins of Tethyan origin that have undergone different fates during Permo-Triassic times (Stampfli, 2000; Stampfli *et al.*, in press). These are a) the Palaeotethys Basin, which began to opened during the Silurian and was finally closed from west to east during the Permo-Triassic, b) the Neotethys Basin, which began to opened during the late Early Permian in conjunction with the detachment of the Cimmerian composite terrane from Gondwana, and c) a system of back-arc basins, located to the north of the Palaeotethys subduction zone, that began to open during the Late Permian and that were variably closed during the Late Triassic and Jurassic. The former southern Palaeotethys passive margin forms an integral part of the Cimmerian terrane. In our model we consider the East-Mediterranean and Ionian Sea basins as forming part of Neotethys. The Apulian domain is regarded as the westernmost part of the Cimmerian composite terrane (Figs 2 to 5).

Along the northern, active Palaeotethys margin, accretionary prisms and fore-arc sequences developed during Carboniferous until Permian or Triassic times, depending on the timing of closure of the respective Palaeotethys segment. Older elements of the Palaeotethys suture are possibly preserved in the Chios Island (Papanikolaou & Sideris, 1983; Baud *et al.*, 1990; Stampfli *et al.*, 1991) and Karaburun peninsula (Kozur, 1997 a), where pelagic Silurian to Late Carboniferous blocks have been described in a Carboniferous matrix, locally enriched in chrome-spinel (Statteger, 1983). Younger, more external Permo-Triassic fore-arc or foredeep clastic sequences are known from the Hellenides (*e.g.* Liri Flysch, Arna schists, Tyros beds; De Bono, 1999) and the Dinarides (Kozur, 1999). Northward, these flysch-type sequences can be linked to the deep-water Kungurian to Roadian flysch found just south of the Periadriatic line in the Carnic Alps within the Trogkofel clastic sequence (*e.g.* Kozur & Mostler, 1992). In time, the Hellenic-Dinarides flysch basins were replaced by the episutural basins of the Pindos-Budva domain (Fleury, 1980; Gorican, 1993; Degnan & Robertson, 1998). This domain continued to subside from Early Triassic (*e.g.* Vardousia sequence; Ardaens, 1978) until Oligocene times (Richter & Müller, 1993 a, b; Richter *et al.*, 1993). In this area, the Early Triassic is represented by pelagic and volcaniclastic deposits; these are overlain by the Middle Triassic Pindos flysch (Priolithos formation; Degnan & Robertson, 1998), the petrography of which plots in the field of recycled orogens and, thus, characterizes the Early Cimmerian event in these regions. Upwards, the Pindos flysch passes into the pelagic Late Triassic-Early Jurassic Drimos series.

Associated with the northern, active Palaeotethys margin, Permian sediments occur also in a system of back-arc basins. Although these Permian sediments are dominated by continental clastics, deposition of marine carbonates commenced locally already during the late Early Permian (*e.g.* Hydra Island for the Meliata-Maliak rift; Stampfli, 2000; Stampfli *et al.*, in press; De Bono *et al.*, 1999). Roll-back of the Palaeotethys subduction slab certainly commenced during the late Early Permian and induced "back-arc" rifting along the entire northern Palaeotethys margin. East of the palaeo-Apulian promontory, rifting culminated in Late Permian and Triassic sea-floor spreading in the Meliata-Maliak Basin (Figs 2 and 3; Kozur, 1991). In conjunction with Late Permian gentle docking of the Apulian terrane against the western-most segment of the Palaeotethys arc-trench system, opening of the Meliata-Maliak back-arc basin aborted in the South Alpine domain, whereas to the East it continued to open and linked up with the Crimea-Svanetia-Küre Basin (Nikishin *et al.*, in press b) and the Dobrogea (*e.g.* Spathian MORB pillow lava of N-Dobrogea Niculitel formation; Cioflica *et al.*, 1980; Seghedi *et al.*, 1990; Nicolae & Seghedi, 1996). To the SE, the Karakaya Basin, that persisted until Early to Middle Triassic times, can be interpreted as a Marianna-type (intra-oceanic) back-arc basin that was located southward adjacent to the Sakarya micro-continent.

All of these areas are characterised by the presence of Early to Middle Triassic pelagic series, often grading upwards into Late Triassic flysch or molasse-type deposits, and the absence of post-Bashkirian to Late Permian pelagic deposits, with the exception of latest Permian pelagic limestones in the Karakaya complex (Kozur, 1997 b). Late Permian to Middle Triassic back-arc basins occur also further to the East in the Pontides (Agvanis-Tokat Basin; Okay & Sahintürk, 1997), the Caucasus (Nikishin *et al.*, 1997, in press b), NE Iran (Baud & Stampfli, 1989), northern Afghanistan (Boulin, 1988) and in the Pamirs (Khain, 1994; Leven, 1995). East of the Moesian promontory, and due to the Late Triassic collision of Cimmerian terranes with the Eurasian margin (*e.g.* Stampfli *et al.*, 1991; Alavi *et al.*, 1997), these back-arc basins were closed during the Early Cimmerian orogeny (Fig. 5).

The pre-Mid-Permian northern passive margin of Gondwana, facing Palaeotethys, is characterized by continuous Devonian to Early-Middle Triassic platform sediments. With the Mid-Permian separation of the Cimmerian composite terranes, this passive margin prism was detached from Gondwana and, in conjunction with progressive closure of Palaeotethys, ultimately collided with its northern active margin (Figs 2 and 3). In Crete, the Early Permian open marine to pelagic series of the (par)autochthonous Mani unit (König & Kuss, 1980) may represent the oldest exposed elements of the southern Palaeotethys passive margin that had persisted until the Early Triassic. The angular unconformity, that separates the Mani unit from the overlying Late Triassic carbonates, marks the Early Cimmerian diastrophism. A Late Carboniferous (Bashkirian) to Early Triassic pelagic sequence is also observed in the Phyllite-Quartzite unit of Crete (Krahl *et al.*, 1983, 1986; Krahl, 1992); it corresponds to the accretionary prism of Palaeotethys (Stampfli *et al.*, 1995; De Bono, 1999) and represents either a distal pelagic facies of the southern Palaeotethys margin or part of the sedimentary cover of the oceanic Palaeotethys Basin.

The carbonate platform sequence of the Taurus nappes (Gutnic *et al.*, 1979; Demirtasli, 1984) probably represents from Late Permian times onward the post-rift sequence that was deposited on the former rift shoulder and/or in a rim basin flanking the Permo-Triassic East-Mediterranean segment of Neotethys. More internal Taurus sequences (*e.g.* Altiner *et al.*, 1980; Demirel & Kozlu, 1998) show strong affinities with the Alborz sequences (NE Iran) that are defined as the reference Palaeotethys-type margin (Stampfli *et al.*, 1991).

South of these areas, the palaeogeographic origin of the Antalya nappes, involving a deformed passive margin sequence, is still controversial (Dumont, 1976; Robertson & Dixon, 1984; Poisson, 1984; Marcoux *et al.*, 1989; Stampfli *et al.*, 1991; Robertson, 1993; Robertson *et al.*, 1996). Similarities between the Antalya nappes and the Mamonia nappe complex of Cyprus suggest, however, that these units are more likely related to the northern Neotethys margin, rather than to tectonically translated more internal domains. In this context, it should be noted that the Maliak-Pindos back-arc ocean and the East-Mediterranean Neotethys opened simultaneously. Therefore, in terms of syn-rift and post-rift subsidence, similar geological histories can be expected in both areas. Thus, a Maliak-Pindos origin of the Antalya nappes cannot be ruled out.

Similar to the southern Neotethys margin, as defined in Oman (Pillevuit, 1993; Pillevuit *et al.*, 1997), the Mediterranean-Neotethyan series extend westwards as far as Sicily. In the Sosio area of western Sicily, Early Permian to Early Triassic marine sediments are represented by pelagic microfauna (Catalano *et al.*, 1988; Catalano *et al.*, 1991) of the clastic Sicanian Basin sequence that also contains Permian fusulinid limestone olistoliths (Skinner & Wilde, 1966). The oldest fauna in the Sicanian Basin is a late Early Permian pelagic bathyal and shallow water microfauna (Catalano *et al.*, 1992; Vachard *et al.*, 1999), the Pacific affinity (Kozur, 1990) of which suggests that this basin was connected to Palaeotethys (Catalano *et al.*, 1995). However, at this time the Neotethys-East Mediterranean basin had not yet open. Similar deep-water Kungurian to Roadian flysch deposits occur also in the Carnic Alps (*e.g.*

Kozur & Mostler, 1992; Kozur, 1999). Together with the Late Palaeozoic Tuscan sequences, located between these two areas, these occurrences could be regarded as forming part of a remnant Early Permian foredeep basin that was associated with the Palaeotethys accretionary prism.

In contrast, Late Permian Hallstatt-type pelagic limestone, similar to those found in Timor and Oman where they sometimes rest directly on MORB (Niko *et al.*, 1996), are also reported from the Sosio complex (Kozur, 1995), strongly suggesting that by Late Permian times the area was connected to the Neotethys.

The Sicanian and Lagonegro sequences of southern continental Italy (Ciarapica *et al.*, 1990 a) were probably deposited in the same basin. Subsidence of the Lagonegro Basin commenced during the Late Permian and persisted during the Triassic, as indicated by the transition from an Early Triassic outer platform facies to Middle Triassic basinal sequences (Miconet, 1988). These sequences are separated by Middle Anisian slope deposits that contain olistostroms and olistolithes, some of which are derived from an Upper Permian carbonate platform (Ciarapica *et al.*, 1990 a, b). This Anisian basin deepening event could be regarded as being related to the final suturing of the Apulian promontory to Variscan Europe (Figs 4 and 5; Stampfli & Mosar, 1999). Pelagic sedimentation dominated the Lagonegro Basin from Middle Triassic until Oligocene times, and thus precludes during that time any important extensional or compressional deformation phases.

CONCLUSIONS

The latest Carboniferous to Triassic stratigraphic and associated magmatic record of the Variscan domain reflects fundamental changes in its megatectonic setting. These changes are related to the late Westphalian termination of syn-orogenic crustal shortening, followed by the Stephanian-Autunian wrench-induced collapse of the Variscan orogen that, at the onset of the Late Permian, gave way to regional thermal subsidence of the lithosphere. However, this post-tectonic cycle of basin subsidence terminated diachronously with the onset of a new rifting cycle that governed the break-up of Pangea along its Palaeozoic and Late Precambrian sutures. Major elements of this break-up system were the southward propagating Arctic-North Atlantic and the westward propagating Tethys rift systems. These linked up in the North Atlantic domain and propagated southward into the Central Atlantic and Gulf of Mexico. Evolution of the Tethys rift system reflects repeated plate kinematic changes in the western Tethys realm. These entailed the Late Permian and Triassic opening of Neotethys and a system of back-arc basins, followed by the closure of remnant Palaeotethys and of some of the Permo-Triassic back-arc basins during the Early Cimmerian orogeny.

The latest Carboniferous-Early Triassic evolution of the Variscan domain mirrors the fundamental plate boundary reorganization that accompanied and followed the terminal suturing phases of Pangea and that underlies its Mesozoic break-up. Changes in the outer core convection system underlay the development of the CPRS and the resumption of frequent magnetic reversals with the early Tatarian Illawarra reversal that, 15 My later, was followed by major plume-related volcanism at the Permo-Triassic boundary. Late Permian and Triassic gradual development of the modern bi-polar mantle convection system provided for regional uplift of the central parts of Pangea and contributed significantly to the Mesozoic break-up of Pangea.

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1.1. ITALY

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EARLY PERMIAN PALAEOFAULTS AT THE WESTERN BOUNDARY OF THE COLLIO BASIN (VALSASSINA, LOMBARDY)

DARIO SCIUNNACH[1]

Key-words – Permian; Orobic Alps; palaeofaulting; red beds; petrofacies.

Abstract – Detailed field mapping at the 1:10,000 scale in the western Orobic Anticline revealed that the boundary between two intrusive bodies in the Southalpine Basement corresponds to an Early Permian palaeofault.
The distribution of the overlying Cisuralian volcanics and clastics (volcanic member of the Collio Formation; Ponteranica Conglomerate, here reported for the first time west of the Biandino Valley), in turn unconformably sealed by upper Guadalupian? to Lopingian red beds (Verrucano Lombardo), might indicate tectonic subsidence of the hanging-wall of another Early Permian normal fault, marking the western boundary of the Collio basin, where coarse-grained alluvial clastics and, further to the east, sandy and muddy sediments were deposited.

Parole chiave – Permiano; Alpi Orobiche; paleofaglie; strati rossi; petrofacies.

Riassunto – Un rilievo geologico di dettaglio in scala 1:10,000 al margine occidentale dell'Anticlinale Orobica ha evidenziato che il limite tra due corpi intrusivi del Basamento Sudalpino corrisponde in realtà ad una paleofaglia del Permiano Inferiore. La distribuzione dei soprastanti depositi vulcanici e terrigeni cisuraliani (membro vulcanico della Formazione di Collio; Conglomerato del Ponteranica, qui descritto per la prima volta ad ovest della Val Biandino), a loro volta ricoperti in discordanza da depositi alluvionali del Guadalupiano superiore?-Lopingiano (Verrucano Lombardo), potrebbe indicare la subsidenza tettonica del tetto strutturale di una seconda faglia normale del Permiano Inferiore che segna il confine occidentale del bacino del Collio, in cui si deposero sedimenti clastici grossolani e, muovendo verso est, arenaceo-pelitici.

INTRODUCTION AND PREVIOUS STUDIES

The Italian program of geological mapping at the 1:50,000 scale (CARG: Catenacci, 1995), attended to for the territory of Lombardy by Regione Lombardia, can be seen as an opportunity to revise both the stratigraphic succession and the tectonic framework of the central Southern Alps. The study area of the present case history is located at the northwestern corner of sheet 076 "Lecco", corresponding to the western end of the Orobic Anticline (Fig. 1). There, a volcanic to terrigenous succession of Permian to Anisian age overlies the Southalpine crystalline basement, here consisting of low- to medium-grade Variscan paraderivates intruded by early post-Variscan plutons.

The study area was considered in the comprehensive monographs of Crommelin (1932), Merla (1933) and De Sitter & De Sitter-Koomans (1949). A thorough stratigraphic revision of the Upper Carboniferous to Triassic succession just east of the study area was provided by Casati & Gnaccolini (1967); the post-Variscan intrusive complex was studied in detail by Pasquaré (1967) and Thöni *et al.* (1992, *cum bibl.*). The relationships between the Variscan basement and its Permian cover were described in Casati (1968) and Gaetani *et al.* (1987).

GEOLOGICAL FRAMEWORK

Crystalline basement

• *Metamorphic host rocks.* Melanocratic micaschists and paragneisses, locally yielding relics of andalusite (Fig. 2A), kyanite, garnet and staurolite and displaying widespread quartz rods, pass to foliated quartzites (Morbegno Gneiss, "Gneiss minuti a biotite" *Auct.*); prominent contact effects by the adjacent plutons are documented by extensive growth of randomly-oriented phyllosilicates (biotite, chlorite) and cordierite. The analysis of quartz rods failed to reveal compositional or textural features, such as relics of stable lithic grains, alignments of ultrastable heavy minerals, or syntaxial quartzose overgrowths, hinting at a primary sedimentary origin; thus, quartz has to be considered as entirely neoblastic.

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Two main deformation phases are recognised: the first marks the development of a widespread S1 foliation, on which a S2 crenulation cleavage is superposed; locally, tight crenulation causes isoclinal folding of the S1 surface and disruption of the transposed hinges. A late metamorphic overprint is recorded by mild – although widespread – kinking of the pre-existing fabrics (S3).

• *Post-variscan plutons.* A large system of sills and laccoliths, mostly consisting of finely crystalline, biotite-rich quartz-diorites (Val Biandino Granodiorite), passes westwards to coarser biotite-rich granodiorites, in which widespread poikilitic quartz encloses plagioclase phenocrysts ("Cortabbio diorite"; Fig. 2B); a large mass of leucocratic hypabyssal granites, displaying large K-feldspar porphyrocrysts in a granophyric microcrystalline groundmass ("Vle Biagio Granite" of De Sitter & De Sitter-Koomans, 1949; Fig. 2C), is confined to the Bindo-Prato San Pietro area.

Lower Permian volcanic-sedimentary succession

• *Collio Formation.* The paroxysmal effusion of intermediate to acidic ignimbrites (benmoreites to rhyolites), yielding zircon ages at 287 (Cadel, 1986) and 283 My (Schaltegger & Brack, 1999) which correspond to the Sakmarian-Artinskian in the time scale of Menning (1995), was followed by episodic eruptions of bimodal products (mugearites-andesites to dacites-rhyolites; preliminary unpublished data by the Author), alternating to stages of volcanic quiescence during which, east of the study area, clastic sedimentation took place.

The contact between the Orobic crystalline basement and the basal Volcanic Member *Auct.* of the Collio Formation (*breviter* Collio volcanics in this paper) is invariably tectonic, and underlined by a blanket, up to a few metres-thick, of black cataclasites (Casati & Gnaccolini, 1967; Casati, 1968) yielding boron mineralisations (Zhang *et al.*, 1994). This tectonic basal contact hampers a precise assessment of the thickness of the Lower Permian volcanics, whereas thickness estimated for the overlying Ponteranica Conglomerate, here bracketed between the Collio volcanics and the Verrucano Lombardo, are more accurate.

The hypothesis according to which the Valsassina intrusions would represent the magma chambers of the Collio volcanics (Merla, 1933; Cadel, 1986), although rea-

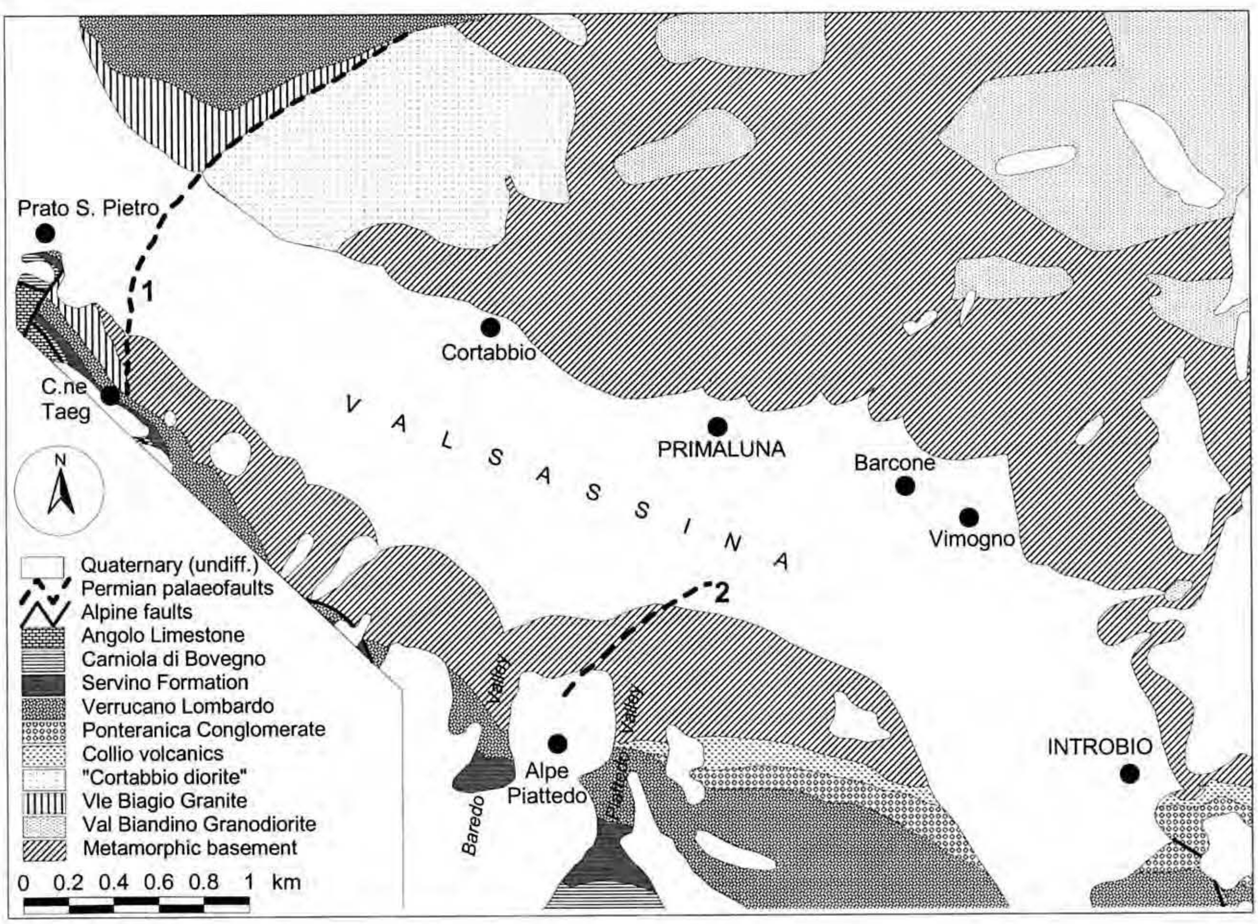

Fig. 1 – Geological sketch map of the study area. Numbers 1, 2 refer to the palaeofaults described in text. The northeastern corner of the study area was mapped in collaboration with G.B. Siletto.

sonable, is difficult to demonstrate because the outcrops lack evidence of a physical continuity between the plutons and the volcanics.

• *Ponteranica Conglomerate.* Coarse-grained clastics become predominant in the upper Collio Fm. Poorly-sorted cobble conglomerates, with angular to subangular volcanic and metamorphic clasts by far prevailing over quartz pebbles, sharply overlie the Collio volcanics from Alpe Piattedo to the Acquaduro Valley (Introbio), where they are eventually cut out by the Valtorta Fault. The few coarse-grained sandstones intercalated to these conglomeratic red beds (Fig. 2E) yielded detrital modes (Fig. 3) clustering in

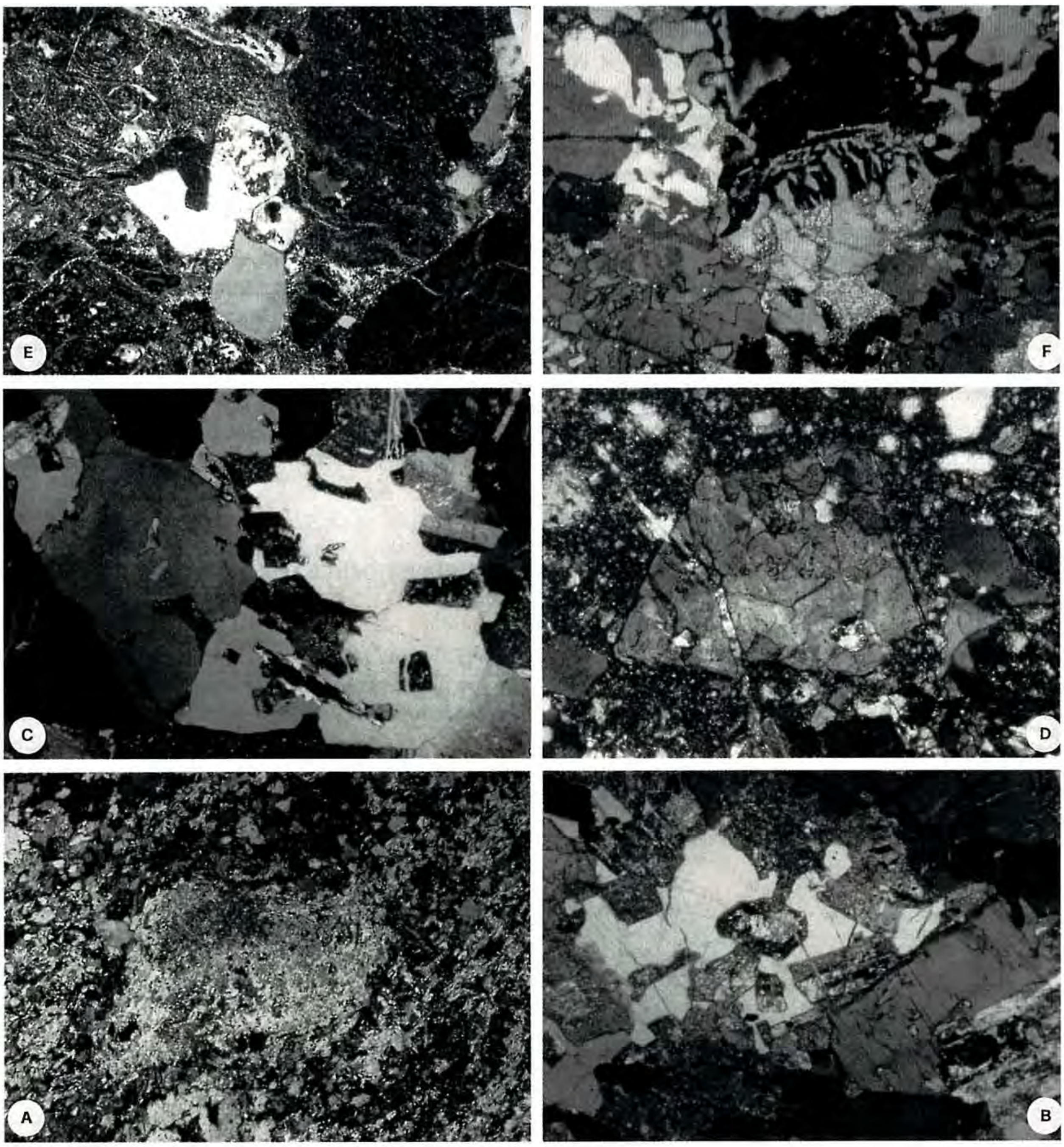

Fig. 2 – Representative photomicrographs for the studied rock types. A. Rotated andalusite? porphyroblast in a phyllonitic sericite paragneiss (metamorphic basement), Cortabbio, 11x. B. Biotite-rich granodiorite ("Cortabbio diorite") with plagioclase phenocrysts implicated by poikilitic quartz, Cortabbio, 17x. C. Leucocratic granite with granophyric implications ("Vle Biagio Granite"), Bindo, 17x. D. Tourmaline-rich cataclasite, Cortabbio/Cortenova boundary fault, 67x. E. Very coarse-grained volcanic arenite with volcanic lithics displaying biotite and embayed quartz phenocrysts, as well as spherulitic structures (Ponteranica Conglomerate), Pizzo dei Tre Signori area, 11x. F. Granophyre pebble in the Verrucano Lombardo conglomerate facies, Introbio, 13x. All photos are in cross-polarised light.

the Lower Permian petrofacies P0 (Q = 10±5, F = 20±8, L = 70±8; C/Q = .17±.13; P/F = .82±.09; V/L = .99±.01), which characterises the sedimentary member of the Collio Formation (Sciunnach *et al.*, 1999). Compositional identity has to be regarded as a reliable indication of physical continuity of clastic bodies at basin scale; the described red beds, displaying lower textural maturity and mineralogical stability with respect to the overlying Verrucano Lombardo, can be thus correlated with the Ponteranica Conglomerate and interpreted as its western end. The Ponteranica Conglomerate, absent on basement highs – due to either non-deposition or erosion – and up to 150 m-thick in the study area, was deposited in footwall-sourced fan-deltas, bordering tectonically-controlled lacustrine basins; thickness locally exceeds 600 m (Casati & Gnaccolini, 1967), as a result of pronounced tectonic subsidence. Regional correlation with the section dated by Schaltegger & Brack (1999) constrains the age of deposition to the Early Permian, possibly to the Artinskian.

Upper Permian to Anisian sedimentary succession.

• *Verrucano Lombardo.* Wine red conglomerates, pebbly cross-bedded sandstones and siltstones, arranged in decametric fining-upward cyclothems, rest either unconformably on the Collio clastics (Casati & Gnaccolini, 1967) or non-conformably on the Early Permian basement highs (Garzanti & Sciunnach, 1997), documenting widespread subsidence of a pre-existing basin-and-swell palaeotopography (Gaetani *et al.*, 1987; Cassinis *et al.*, 1988). The Verrucano Lombardo was largely deposited in a vast braidplain; rounding of volcanic pebbles and sharp variations in thickness (increasing from less tha 150 to over 400 m, west to east, in the Introbio area) are consistent with transport, under high topographic gradients, of detritus derived from sources located some tens of kilometres to the west. Actually, rare pebbles with granophyric structure (Fig. 2F) could be tentatively restored to a source rock coeval with the Cuasso al Monte granophyre (Buletti, 1985), although their local concentration might even suggest provenance from nearby sources. Age is constrained as mostly Late Permian (latest Guadalupian? to Lopingian) due to stratigraphic position and regional correlation with the roughly coeval Val Gardena Sandstone.

• *Servino to Bellano Formations.* Quartzose clastics passing upwards to poorly exposed and strongly deformed marly dolostones and silty marls (Servino Formation) are overlain by badly tectonised, evaporitic vuggy dolostones (Carniola di Bovegno). These transitional units, deposited after transgression of a shallow epicontinental sea on the Verrucano Lombardo braidplain, form a continuous cataclastic belt at the tectonic contact between the Orobic Anticline and the Northern Grigna thrust sheet. In the latter, the Carniola di Bovegno is overlain by the coarse-grained clastics of the Anisian Bellano Formation, deposited in fan-delta setting, and laterally by the silty member of the Angolo Limestone (Gaetani *et al.*, 1987; Sciunnach *et al.*, 1996).

OUTCROP EVIDENCE FOR EARLY PERMIAN PALAEOFAULTS

Fault 1 (Figs 1, 4) is exposed along the talweg of a small stream (Fig. 5) marking the administrative boundary between the towns of Primaluna and Cortenova. It is underlined by black cataclasites and separates the Vle Biagio Granite (hanging-wall) from the "Cortabbio diorite" (footwall); the latter includes slivers of the metamorphic host rocks (Fig. 4). Both units are sealed by a thick sheet of red beds (Verrucano Lom-

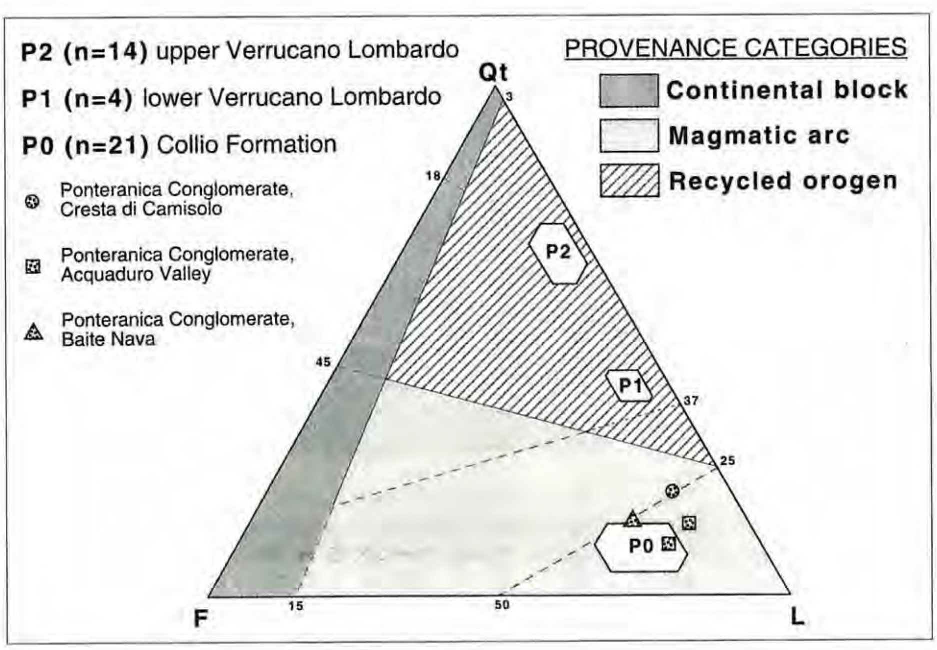

Fig. 3 – QtFL plot of the Ponteranica Conglomerate sandstones compared to the petrofacies recognised in the Permian succession of the Orobic Alps by Sciunnach *et al.* (1996, 1999). Polygons are one standard deviation each side of the mean.

bardo) that appears to be unfaulted, and just gently buckled by the large-scale, open folding of the Orobic Anticline. It is noteworthy that the same outcrop pattern as in Fig. 1 had been already described in Crommelin (1932), where the rock units were mapped correctly, but their mutual geometrical relationships were not interpreted in terms of prealpine faulting; fault 1 is also pictured in Schönborn (1992, fig. 17) as a pre-Verrucano fault, but without any comment in the text. On the path leading from Prato San Pietro to C.ne Taeg, a non-conformable stratigraphic contact of the Verrucano Lombardo on the "Vle Biagio Granite" is observed; the topmost 2÷3 metres of

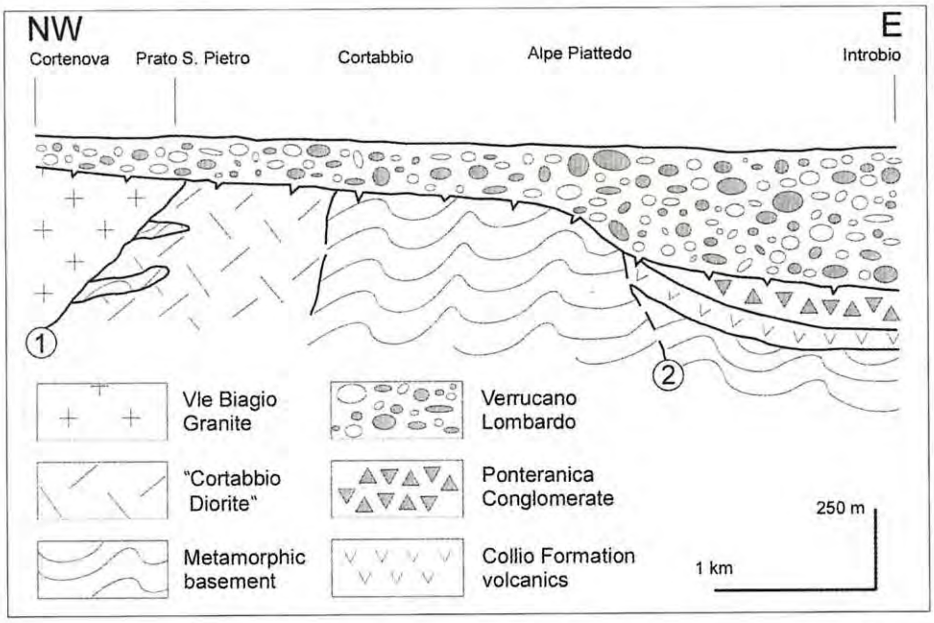

Fig. 4 – Simplified stratigraphic scheme for the studied succession (vertical exaggeration 2x). The upper datum plane is given by the inferred Permian-Triassic boundary.

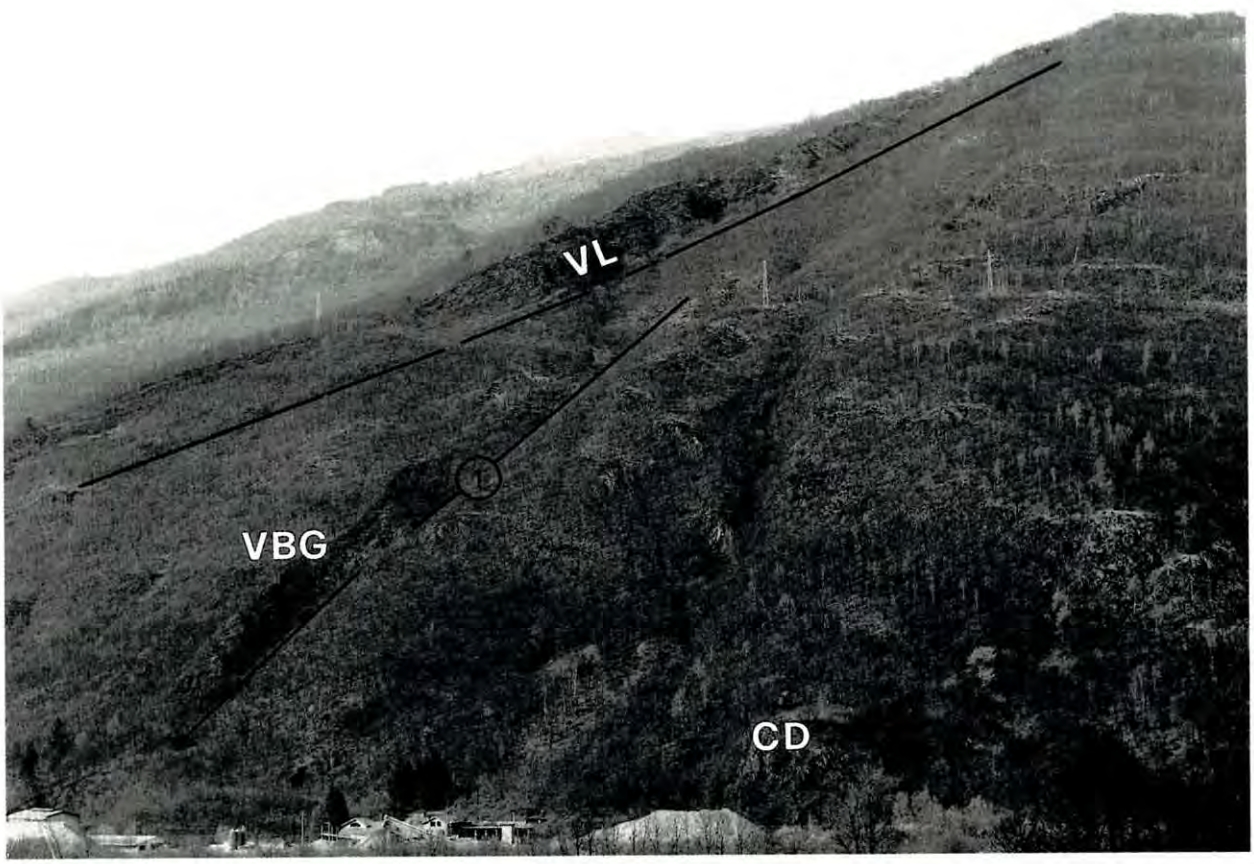

Fig. 5 – Palaeofault 1 in outcrop along the northwestern slope of Valsassina. CD = "Cortabbio diorite", VBG = "Vle Biagio Granite", VL = Verrucano Lombardo.

the granite below the non-conformity are reddened and deeply weathered, probably as a result of prolonged subaerial exposure during the Permian.

The black cataclasites yielded abundant tourmaline (Fig. 2D), occurring as euhedral crystals up to 3 mm in length. According to WDS analysis (Tab. 1), tourmaline can be classified as a solid solution of dravite and schorl and related to Ca-poor metapelites, metapsammites and quartz-tourmaline rocks according to Henry & Guidotti (1985). In the Orobic Alps, glassy tourmalinites at the tectonic contact between basement and volcanic cover have been interpreted in terms of contact by boron-rich hydrothermal fluids with aluminous host rocks, that might correspond to the Collio sediments (Zhang *et al.*, 1994) as well as to the associated volcanics.

Around Alpe Piattedo (Introbio), previously unrecognised outcrops of Ponteranica Conglomerate are restored to the western end of a clastic wedge that pinches out towards a structural high to the west and is sealed by a continuous blanket of Verrucano Lombardo red beds; thickness of the latter strongly and abruptly increases from northwestern Valsassina to the Introbio area, probably as a result of a pre-existing basin-and-swell palaeotopography (Gaetani *et al.*, 1987) that is inferred to have been controlled by a normal fault (2 in Figs 1, 4). It is reasonable to assume that the deposition of the Ponteranica conglomerate wedge was triggered by the tectonic subsidence of the eastern hanging-wall.

Oxides, weigth %		Element stoichiometry		
B_2O_3	10.56	B	3.000*	
SiO_2	36.69	Si	6.035	
Al_2O_3	30.55	Al	5.921	
TiO_2	1.14	Ti	0.141	
FeO	8.01	Fe	1.102	2.983
MnO	0.05	Mn	0.007	
MgO	7.07	Mg	1.733	
CaO	0.36	Ca	0.063	
Na_2O	2.25	Na	0.717	0.784
K_2O	0.02	K	0.004	
Total	96.70	Structural formula on the basis of 24.5 + 4.5 oxygens		

Tab. 1 - WDS microanalysis on tourmaline from the cataclasite underlining fault 1. FeO is meant to include possible Fe_2O_3; the weight of B_2O_3 is calculated from the theoretical stoichiometric value (*).

CONCLUSIONS

• *Cortabbio/Cortenova boundary fault (1).* Age of this pre-alpine fault is fairly well constrained. The proposed geometrical scheme (Fig. 4) clearly shows that faulting post-dates the emplacement of the plutons and pre-dates the deposition of the Verrucano Lombardo: the faulting event is thus safely constrained to the Early Permian. In fact, the post-Hercynian plutons of the western Orobic Alps have been mostly dated as Late Carboniferous to earliest Permian (Thöni *et al.*, 1992; Siletto *et al.*, 1993 *cum bibl.*), whereas the Verrucano Lombardo is usually ascribed to the Upper Permian due to its stratigraphic position; however, its base might even reach locally into the Guadalupian (mid-Permian) as suggested by the occurrence of the Kiaman/Illawarra paleomagnetic reversal in the roughly coeval Val Gardena Sandstones (Dachroth, 1976; Mauritsch & Becke, 1983). In the Early Permian, the most important phase of extensional tectonics in the central Southern Alps is related to the development of the rapidly-subsiding Collio basin, that can be ascribed to the Autunian in the Orobic Anticline (Casati & Gnaccolini, 1967) and can be dated all over its outcrop area as largely Sakmarian to Artinskian according to the zircon ages by Cadel (1986) and Schaltegger & Brack (1999; time scale after Menning, 1995). Fault geometry and widespread granophyric structures in the Vle Biagio Granite (hanging-wall) seem to indicate a normal displacement of the latter with respect to the deeper-seated "Cortabbio diorite". Downthrow of the northwestern hanging-wall would imply the opening of a trough-shaped basin in an area lacking Lower Permian sediments; this might indicate that the fault was associated with an early stage of extensional collapse of the Hercynian orogen, possibly influenced by positive buoyancy of sectors of the range intruded by plutons (Val Biandino complex) with respect to the adjacent, intrusion-free areas (Taceno district, Collio basin). Moreover, the fault forms an angle of about 45° with the base of the Verrucano Lombardo; normal faults are commonly inclined by at least 60° due to mechanical convenience criteria, so it is likely that 1) either the whole considered sector of the Southalpine basement was tilted by at least 15° prior to deposition of the Verrucano Lombardo, 2) or the fault was lystric and Permian erosion, coupled with isostatic compensation and uplift of the Hercynian orogen, cut deep into the Southalpine Basement where the fault plane was asymptotically tending to horizontal.

• *Alpe Piattedo fault (2).* Coarse-grained clastics of Early Permian age, testifying to proximal alluvial fan facies, are superposed to the Collio volcanics as westwards as Alpe Piattedo. The western boundary of the area where the Lower Permian volcanics are preserved was interpreted in Gaetani *et al.* (1987) as a normal fault, accommodating the sharp increase in thickness of the Verrucano Lombardo from nothwestern Valsassina to the Introbio area. Such an increase, however, is partly accounted for by the occurrence of the Ponteranica Conglomerate; the resulting stratigraphic scheme also allows to better constrain the age of normal faulting, which seemingly shortly followed the volcanic activity and triggered deposition of the volcanic detritus eroded from the footwall in the adjacent hanging-wall troughs. The inferred Alpe Piattedo palaeofault seemingly marks the western boundary of the articulated Early Permian Collio Basin; its NE strike matches the regional trend of basin boundary faults in the Novazza-Val Vedello district (Cadel, 1986 *cum bibl.*).

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TETRAPOD FOOTPRINTS FROM THE LOWER PERMIAN OF WESTERN OROBIC BASIN (N. ITALY)

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Key words – footprints; Permian; Southern Alps; stratigraphy.

Abstract – Large numbers of tetrapod footprints have recently been discovered within sediments pertaining to the Early Permian Collio Fm. in the Western Orobic Basin (between the upper Varrone and upper Gerola valleys, Lombardy, Italy).
They can be ascribed to the following taxa: *Amphisauropus latus* Haubold, 1970, *A. imminutus* Haubold, 1970, *Dromopus lacertoides* (Geinitz, 1861) and *Varanopus curvidactylus* (Moodie, 1929).
This low-diversity ichnoassociation is coeval with those of Central Europe and North America and can also be correlated with that yielded from the lower portion of the Brescian Collio Fm. The geological setting as well as the stratigraphic position of the track-bearing sequence is outlined below, together with a discussion of some hypotheses for the presence of this low-diversity ichnoassociation.

Parole chiave – impronte; Permiano; Sudalpino; Stratigrafia.

Riassunto – Un gran numero di impronte di tetrapodi è stato rinvenuto recentemente in terreni appartenenti alla sommità stratigrafica della Formazione di Collio (Permiano inferiore) nel bacino Orobico occidentale (tra l'Alta Val Varrone e l'Alta Val Gerola, Lombardia, Italia). Si tratta delle seguenti icnospecie: *Amphisauropus latus* Haubold, 1970, *A. imminutus* Haubold, 1970, *Dromopus lacertoides* (Geinitz, 1861) e *Varanopus curvidactylus* (Moodie, 1929). Questa icnoassociazione, caratterizzata da una bassa diversità, è coeva con le associazioni del Permiano inferiore dell'Europa Centrale e del Nord America e può inoltre essere correlata con quella rinvenuta nella parte inferiore della Formazione di Collio, nell'omonimo bacino triumplino. Oltre all'inquadramento geologico e stratigrafico della successione di terreni che ha fornito tali impronte, vengono di seguito discusse varie ipotesi sulle possibili cause di tale ristretta icnoassociazione.

INTRODUCTION AND GEOLOGICAL SETTING

Late Palaeozoic tetrapod footprints from sediments cropping out both in the Lower Permian of the central Southern Alps and in the Upper Permian of the Dolomites have been known since the late 19th century (see Conti *et al.*, 1991, for a historical review).

Recently, numerous tetrapod footprints have been found in the Permian deposits of the Western Orobic Prealps (upper Gerola-Inferno and Varrone valleys, Lombardy, Italy) (Fig. 1).

The footprints come from the uppermost levels of the local Collio Fm. and have been ascribed to *Amphisauropus latus* Haubold, 1970, *A. imminutus* Haubold, 1970, *Dromopus lacertoides* (Geinitz, 1861) and *Varanopus curvidactylus* (Moodie, 1929) (Plate 1 and Fig. 3).

This association is closely comparable to a similar one recorded in the lower portion of the Collio Fm. cropping out within the Collio Basin in the Brescia region, and to the similar and coeval Early Permian association of Central Europe and North America.

The first to report the presence of tetrapod footprints, but on the eastern side of the Orobic Basin (eastern Brembana Valley) was Dozy (1935), who ascribed these forms to *Anhomoicnium orobicum* and *Onychicnium escheri*. The first of them was later attributed to an extramorphological imprint of *Batrachichnus salamandroides* (Haubold, 1996), while the other taxon was considered as *incertae sedis* (?*Actibates*) (Haubold, 1971).

Many years after, Casati & Gnaccolini (1967) and Casati & Forcella (1988) reported scarce findings of some tetrapod footprints in the area close to Rifugio Falc and Pizzo Varrone.

Concerning the geology and stratigraphy of the area, the first detailed studies were undertaken by Porro (1931, 1932); subsequently, De Sitter & De Sitter-Koomans (1949 and references therein) published a wide geological map of the Bergamo Alps, putting together the contributions and

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surveys of many other authors. For a more up-to-date stratigraphic framework of the Western Orobic Basin (including a 1:25,000 geological map), we must refer to the work of Casati & Gnaccolini (1967). The stratigraphy of the whole area is also reported by the same authors in the Illustrative Notes of Sheet 18 "Sondrio", 1:100,000 Geological Map of Italy (Bonsignore *et al.*, 1971).

As in the other Late Paleozoic basins of the Southern Alps, in the Orobic Basin two main tectonosedimentary cycles can be recognised: the first one, spanning from the Late Carboniferous(?) to the Early Permian, is mainly represented by a volcanosedimentary lithosome (*i.e.* the Collio Fm. and heteropic formations) deposited in intramontane lacustrine-to-alluvial troughs; the second one, Late Permian in age and separated by a regional unconformity, is represented by a coarse-to-fine alluvial blanket, named the Verrucano Lombardo-Arenaria di Val Gardena (to the west or east of the Adige Valley respectively), which has filled the previous depressions and spread over wide areas.

STRATIGRAPHY

Northeast of Pizzo Varrone, where the richest ichnofossil-bearing strata are located, a 130 m thick stratigraphic section was described in the Collio Fm (up to several hundreds of metres in the area).

It is represented by grey-to-reddish shales and arenites (referred to as the "Scisti di Carona", so-called by De Sitter & De Sitter-Koomans, 1949) with medium to coarse pebbly sandstone and microconglomerate layers interfingered in the lower part. These latter could represent the distal portions of coarse detritic alluvial fan deposits named "Conglomerato di Ponteranica" (Casati & Gnaccolini, 1965).

As a whole, the section (Fig. 2) is very well bedded in centimetre to decimetre-scale layers, mostly representing minor sedimentary cycles with frequent scours or convex erosional bases and internal fining-upwards sequences. Sedimentary structures such as planar and cross-laminations, as well as trough cross-bedding, can commonly be observed; on the bedding surfaces, climbing and wave-ripple marks, raindrop casts, mudcracks and burrows are also frequent. This part of the formation, from which the ichnofossils come, can be related to a flat alluvial plain cut by small ephemeral streams and shallow lakes, pertaining to a semi-arid environment.

The presence within arenitic layers of imbricated black clay-chips of intraformational origin suggests frequent changes in water energy and abrupt increases in erosional power.

In the aforementioned section, macrofloral fragments (*Walchia* sp.), purple-red silicified wood stems of centimetre to decimetre-scale, freshwater coelenterates (hydromeduse), stromatolitic mounds and algal oncolites were also recognised by the authors. These forms are associated with yellow ochre carbonate and ferroan dolomite beds and concretions. It is also worth remembering the recent discovery, close to this site, of significant three-dimensional macrofloral remains of the conifer *Cassinisia orobica* (Kerp *et al.*, 1996).

In the described section, the boundary between the Collio Fm. and the overlying Verrucano Lombardo Fm. is

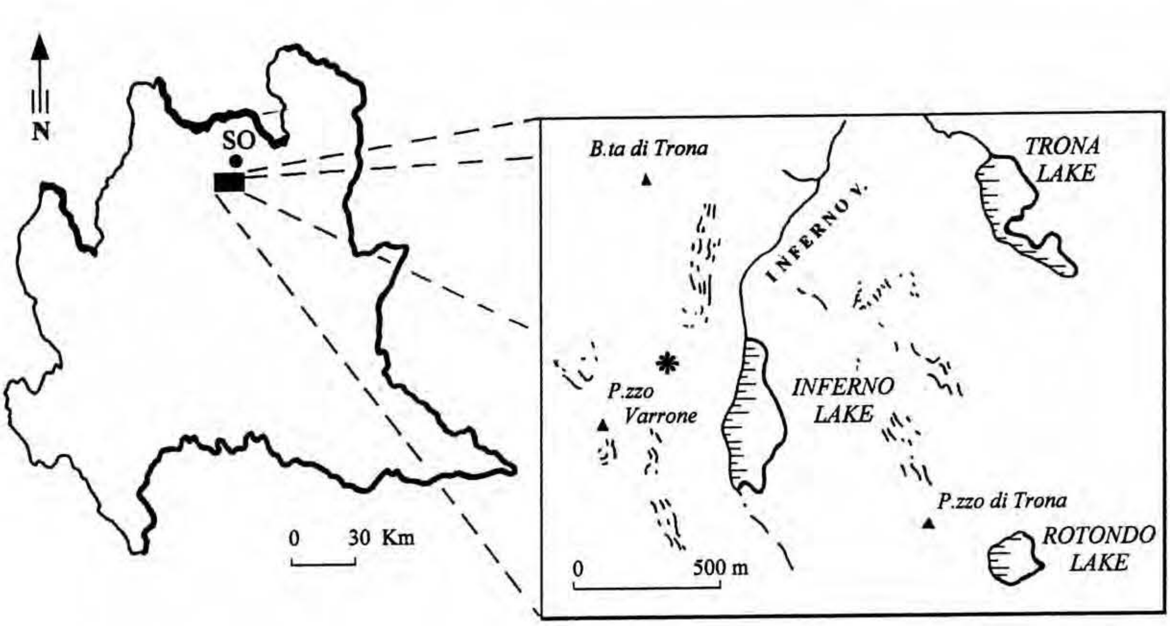

Fig. 1 – Location of the investigated area.

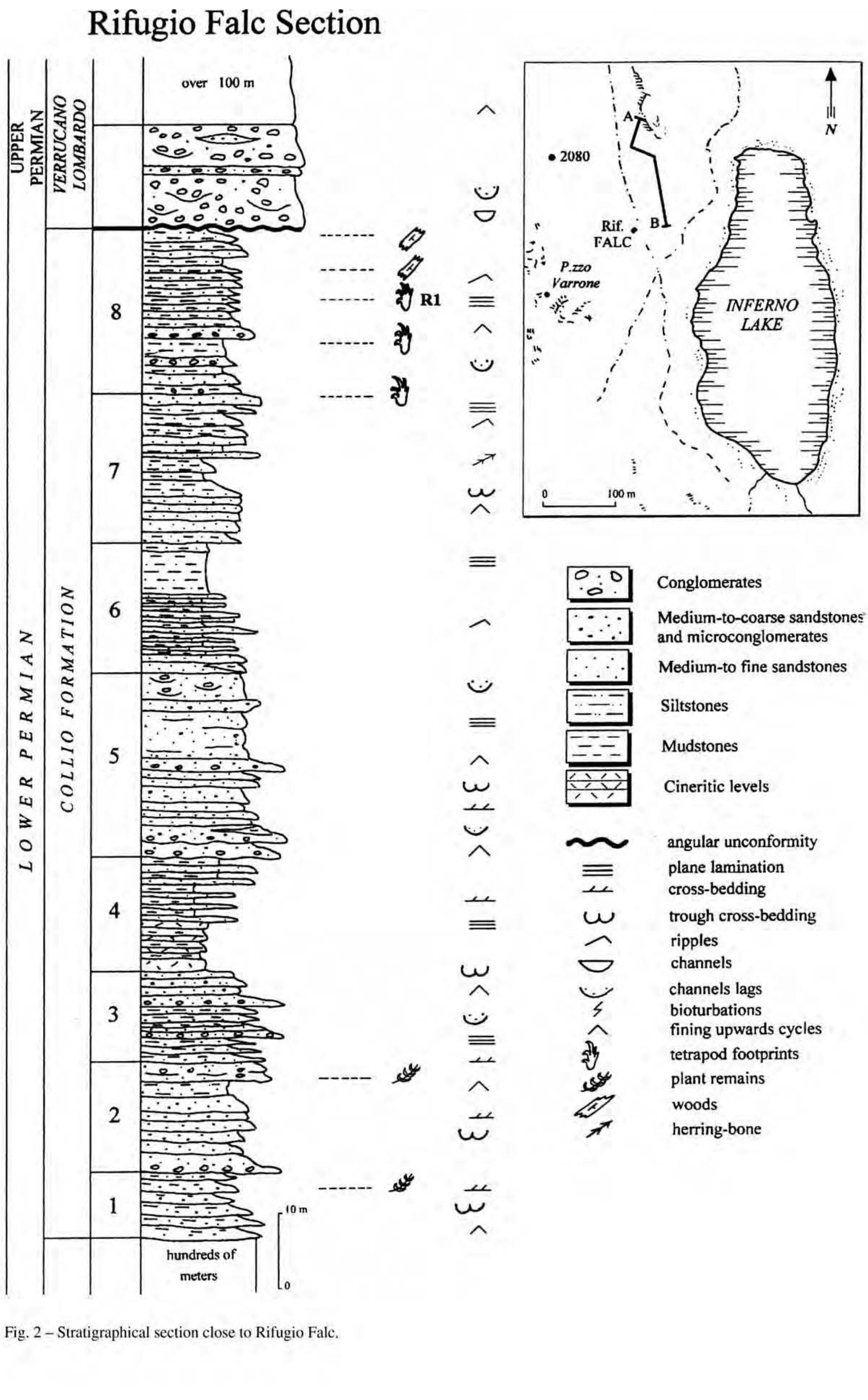

Fig. 2 – Stratigraphical section close to Rifugio Falc.

marked by an angular unconformity (locally more than 20°). The reddish deposits of the latter formation are represented by pebble-to-cobble conglomerates, sandstones and subordinate siltstones of a braided river environment. These deposits, cropping out in the study area with a thickness of nearly 100 m, mark the inception of the Late Permian major sedimentary cycle, which has been recognised over the whole southern Alps region. The stratigraphic log was measured from a few metres above the unconformity (between Verrucano and Collio Fm.) up to q. 2150 on the western shore of the Inferno Lake. From the base it has been subdivided into eight different lithostratigraphical units.

ICHNOLOGY

As is usual for Early Permian tetrapod footprint associations, the Gerola Valley ichnofauna is represented by a few taxa (Plate 1 and Fig. 3). These are more frequently recognised within the corresponding North American and central European Permian associations (Haubold, 1996; Schult, 1995). The relatively low diversity does not permit firm conclusions, but nevertheless the presence of an ichnoassociation of at least four taxa allows us to distinguish it with respect to the *Dromopus didactylus* monotypical ichnoassociation which characterises the highest portion of the Lower Permian in the Collio Fm. type area (Conti *et al.*, 1997). Conversely, the ichnofauna from the Gerola Valley is rather similar to that recognised by Ceoloni *et al.* (1987) in the lower portion of the Collio Fm. in the Trompia Valley (see also the up-to-date data in Conti *et al.*, 2000).

With respect to this last ichnoassociation, the main differences are the lack of *Batrachichnus* and *Ichniotherium* and the already-debated presence of *Varanopus curvidactylus* (Plate 1) in place of *Camunipes cassinisi*. The only noticeable difference between the low diversity of the aforementioned Orobic Basin fauna and the Lower Permian European ones is the lack in the former basin of *Batrachichnus, Ichniotherium, Limnopus* and *Dimetropus*. These last two ichnogenera are also lacking within the association of the Brescia region, and at this time could be considered as forms restricted to the more northern regions of Laurasia.

BIO- AND CHRONOSTRATIGRAPHICAL CONSIDERATIONS

The generally low number of ichnotaxa from Permian sediments has previously been interpreted in two different ways. The usual interpretation considers this a result of an actual Early Permian 'low diversity'; however, this interpretation presents some problems, it being difficult to imagine a long-lasting fauna so reduced in number. A second interpretation considers the low number of ichnotaxa a result of a 'taxonomical compression', related to an assumed very low power of resolution for footprints. In this hypothesis each ichnogenus would represent a bone-based family of reptiles (Lucas, 1998).

We could suggest a third hypothesis that can be defined as 'deposition time compression'. In this case the low number of ichnotaxa could be related to a very short time interval in which all the Lower Permian track-bearing sediments were laid down. This last hypothesis is supported by recent geochronometric data (Cassinis *et al.*, 1999, and in press), by the recognition of the increasing importance and areal effect of stratigraphic gaps present within most of the

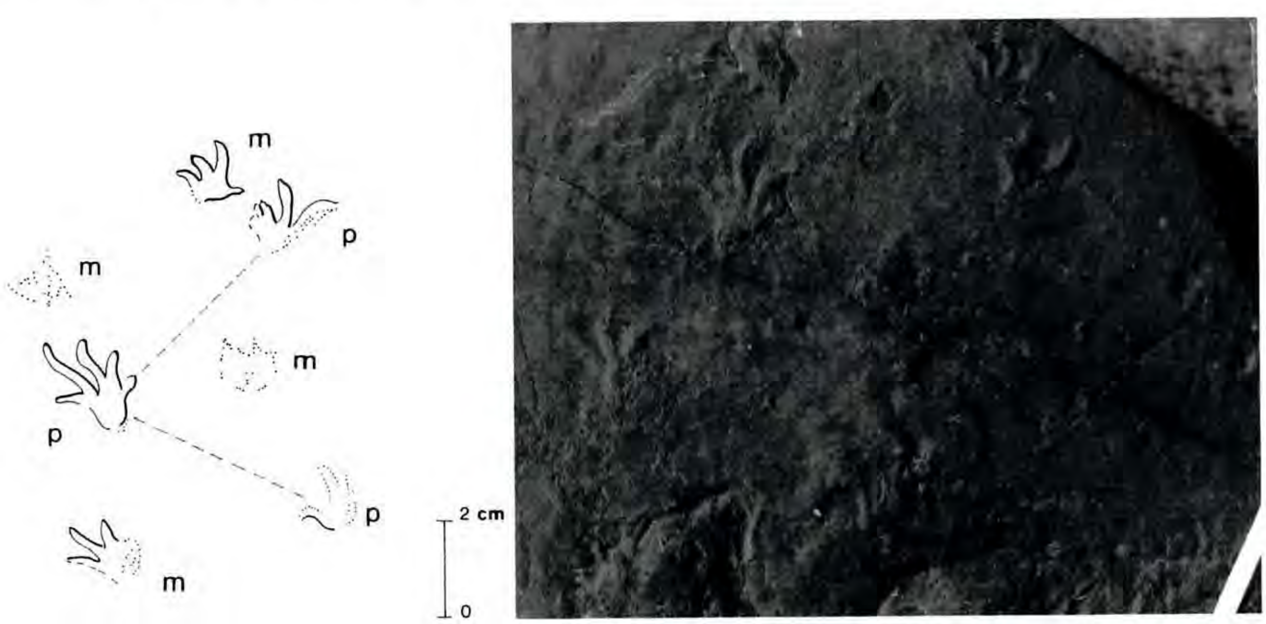

Fig. 3 - Slab with tracks of *Amphisauropus imminutus* Haubold, 1970.

Permian basins, and by the downward compression of the Rotliegend (following the calibration of the Tambach Fm. by Sumida *et al.*, 1996).

Conti *et al.* (1997) used the central Alps Lower Permian ichnofauna, systematically studied mainly on a complete section sampled bed by bed in the Collio Fm., to establish a faunal unit (Collio FU) and the associated faunal age (Collio FA). This last was in its turn subdivided into two faunal subages (Rabejac FsA and Tregiovo FsA)(Conti *et al.*, 1997; Cassinis *et al.*, 2000). This was a first attempt at establishing a sequence of Permian biochronological units. These can usefully replace the less meaningful use of biozones or associations (Boy & Fichter, 1982; Holub & Kozur, 1981), or the subdivision of the continental deposits by means of

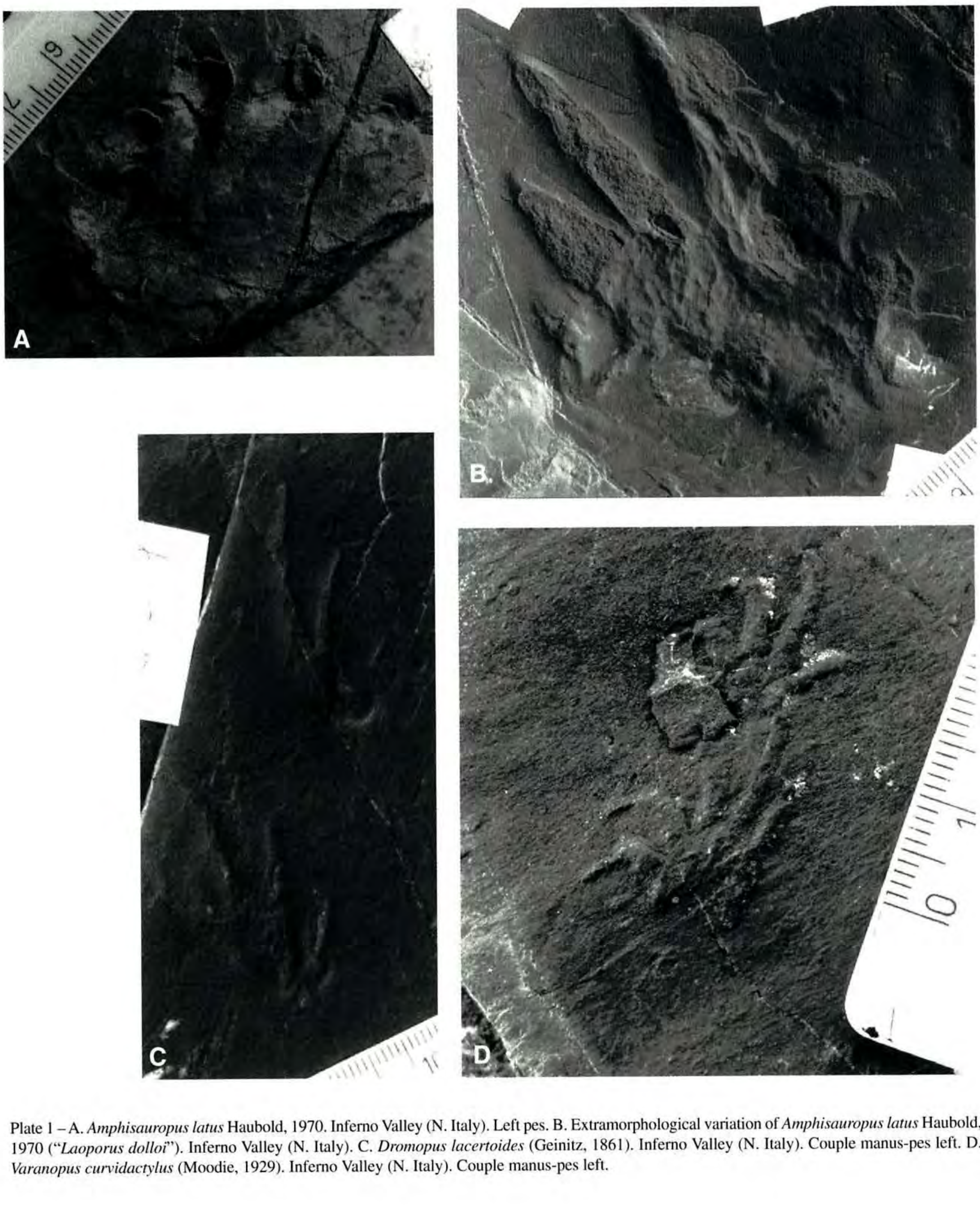

Plate 1 – A. *Amphisauropus latus* Haubold, 1970. Inferno Valley (N. Italy). Left pes. B. Extramorphological variation of *Amphisauropus latus* Haubold, 1970 ("*Laoporus dolloi*"). Inferno Valley (N. Italy). C. *Dromopus lacertoides* (Geinitz, 1861). Inferno Valley (N. Italy). Couple manus-pes left. D. *Varanopus curvidactylus* (Moodie, 1929). Inferno Valley (N. Italy). Couple manus-pes left.

marine stages, or, in the worst case, the so-called 'continental stages' of the European geological tradition.

The new finds allow the use of the Rabejac FsA, ascribing to the outcrops of the upper Val Varrone and Gerola valleys an age corresponding to the depositional age of the lower portion of the Collio Fm. If this is confirmed, the unconformity between the Collio Fm. and the second-cycle Verrucano Lombardo deposits could correspond to a larger gap in this area than in the Collio Basin, including also the time interval in which sediments ascribed to the Tregiovo FsA were laid down. This hypothesis also fits well with the marginal paleogeographical position of the section in relation to the more depocentral locations of other sections.

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THE LOWER PERMIAN IN THE OROBIC ANTICLINES (LOMBARDY SOUTHERN ALPS): CRITERIA FOR FIELD MAPPING TOWARDS A STRATIGRAPHIC REVISION OF THE COLLIO FORMATION

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Key words – Permian; Southern Alps; geological mapping; lithofacies, stratigraphy.

Abstract – Detailed field mapping at the 1:10,000 scale in the Orobic and Cedegolo Anticlines requires a balanced approach to the stratigraphic subdivision of the Collio Formation.
A scheme of mapping lithofacies for this complex unit is proposed and tested in the three Orobic Anticlines, aiming at a stratigraphic revision of the Lower Permian in the central Southern Alps that, however, needs further detailed mapping and analytical work.

Parole chiave – Permiano; Alpi Meridionali; rilevamento geologico; litofacies; stratigrafia.

Riassunto – Il rilevamento geologico di dettaglio alla scala 1:10,000 nelle Anticlinali Orobica e di Cedegolo necessita di un approccio bilanciato alla suddivisione stratigrafica della Formazione di Collio. Uno schema di litofacies mappabili per questa complessa unità viene qui proposto e verificato nelle tre Anticlinali Orobiche, nell'ottica di una revisione stratigrafica del Permiano Inferiore nel Sudalpino centrale che, tuttavia, richiederà ulteriore lavoro analitico e di terreno.

INTRODUCTION

In the framework of the Italian project of geological mapping at the 1:50,000 scale (CARG; Catenacci, 1995), two areas of about 75 and 50 km^2 have been mapped at the 1:10,000 scale in the Orobic and Cedegolo Anticlines (Lombardy Southern Alps; Fig. 1). There, one of the most problematic geological objects to cope with is the Lower Permian Collio Fm., a complex continental succession of volcanic and clastic rocks, locally exceeding 1000 m in thickness, largely deposited in a transtensional tectonic setting (Massari, 1988). After the original description by Gümbel (1880), who restricted the use of this term to the dark plant-bearing shales and sandstones of Autunian age in Val Trompia, the name "Collio Formation" has been extended even in the type area to embrace the whole volcanic and volcaniclastic succession bracketed between the crystalline basement (± the aporphyric Basal Conglomerate) and the Verrucano Lombardo, thus enclosing also the volcanic plateau underlying the typical "Collio" sediments (Cassinis, 1966). Strong lateral variability of facies and thickness documents syntectonic deposition in hangingwall basins accomodating the clastic sediments from footwall source areas.

The stratigraphic reference framework in the study areas subdivides the Collio Fm. simply into a "volcanic" lower member and a "sedimentary" upper member (Casati & Gnaccolini, 1967; Dozy, 1933); on the other hand, an accurate study carried out in the Trabuchello-Cabianca Anticline (Cadel *et al.*, 1996) proposes to subdivide the "volcanic" lower member into 18 lithozones and the "sedimentary" upper member into 6 facies, that are admittedly difficult to correlate outside their type sections and form a stratigraphic framework too detailed for the purposes of regional mapping. In the attempt at balancing these different approaches into a coherent lithostratigraphy that can be employed all over the study areas, a preliminary facies mapping has been carried out. The complex problems of stratigraphic nomenclature related to the Collio Fm. are beyond the scope of this paper, and the lithostratigraphic rank of this unit, although needing reconsideration, will not be questioned.

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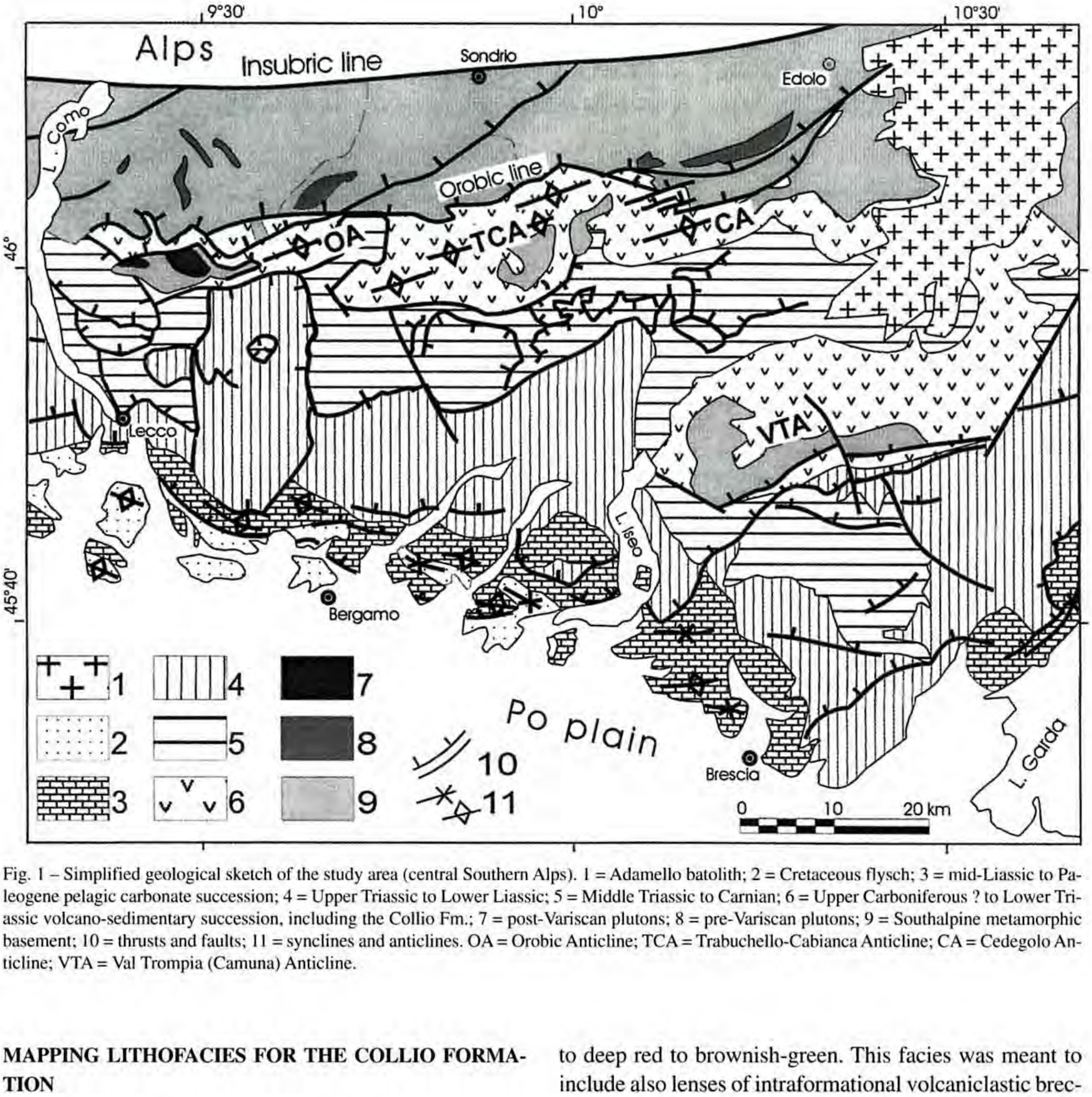


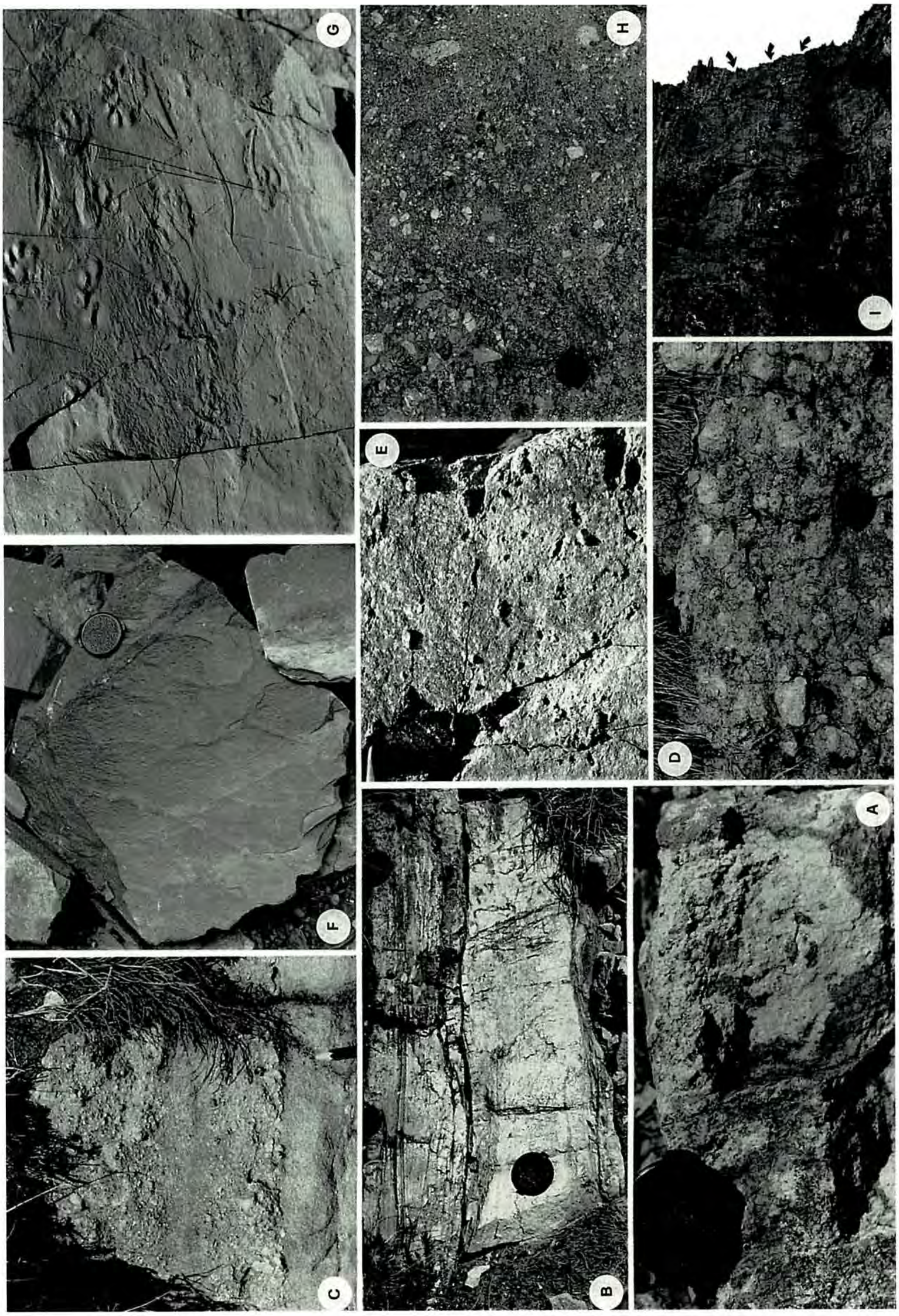
Fig. 1 – Simplified geological sketch of the study area (central Southern Alps). 1 = Adamello batolith; 2 = Cretaceous flysch; 3 = mid-Liassic to Paleogene pelagic carbonate succession; 4 = Upper Triassic to Lower Liassic; 5 = Middle Triassic to Carnian; 6 = Upper Carboniferous ? to Lower Triassic volcano-sedimentary succession, including the Collio Fm.; 7 = post-Variscan plutons; 8 = pre-Variscan plutons; 9 = Southalpine metamorphic basement; 10 = thrusts and faults; 11 = synclines and anticlines. OA = Orobic Anticline; TCA = Trabuchello-Cabianca Anticline; CA = Cedegolo Anticline; VTA = Val Trompia (Camuna) Anticline.

MAPPING LITHOFACIES FOR THE COLLIO FORMATION

In our proposal, the lithofacies to be systematically mapped in the field (Figs 2, 3) form a relatively narrow range and are listed below. They are somehow oversimplified with respect to the complexity of facies characters observed in the field, as they were established to allow easy recognition even in small outcrops. The lithofacies are listed in a rough stratigraphic order from base to top, but, owing to the episodic character of volcanism and to the mobility of the depositional environment, exceptions, repetitions and gaps may be locally present (Fig. 4).

V_1 – Tabular welded tuffs and lapillistones (locally preserving *fiammae* and thus correctly classified as ignimbrites) of benmoreitic to rhyolitic composition, with variable amount of porphyrocrysts and degree of welding, up to massive porphyries, ranging in colour from whitish to deep red to brownish-green. This facies was meant to include also lenses of intraformational volcaniclastic breccias consisting of 100% angular volcanic clasts welded in an aphanitic matrix. A rough bedding is locally recognised, underlined by thin tuffaceous interlayers, but in

Fig. 2 – Representative outcrops of facies described in the three Orobic Anticlines. A = welded lapillistone in amalgamated beds (small arrow at a bedding plane) with variable degree of porphyricity and welding (V1, Orobic Anticline); B = laminated tuff with preserved depositional structures (V1, Orobic Anticline); C = normal to inverse grading (FU then CU) in a lapilli tuff with lithic and pomiceous volcanic grains (V2, Orobic Anticline); D = intraformational volcanic breccia (V2, Orobic Anticline); E = volcaniclastic paraconglomerate (S1, Orobic Anticline; hammer head at the top left corner for scale); F = mud-cracks in grey mudrocks (S3, Trabuchello-Cabianca Anticline); G = tetrapod footprints and tail marks preserved in fine-grained sandstone (S2, Trabuchello-Cabianca Anticline; field of the photograph approximately 1 m; walking direction to the right); H = Ponteranica Conglomerate pebble to cobble breccia (S4, Orobic Anticline); I = ferroan carbonate beds alternating to bedded sandstones (S5, Cedegolo Anticline).


G
H
I
E
D
F
A
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B

general is pervasively overprinted by the Alpine cleavage.

V_2 – Massive to bedded breccias and arenites ("epiclastics") with poorly-sorted, angular clasts exclusively volcanic in origin embedded in a salt-and-pepper sandy matrix; abundant plagioclase is recognised in the arenites.

V_3 – Massive lava flows, deep green to almost black in colour, of mugearitic to "andesitic" composition, scattered at various stratigraphic levels and fed by porphyrite dykes. Black femics and saussurritized pale green plagioclase are locally observed; aphyric textures are, however, more common.

V_4 – Volcanic breccias, locally interpreted as agglomerates, to cinerites with poorly-sorted angular to volcanic clasts and rounded to euhedral quartz clasts. Volcanic

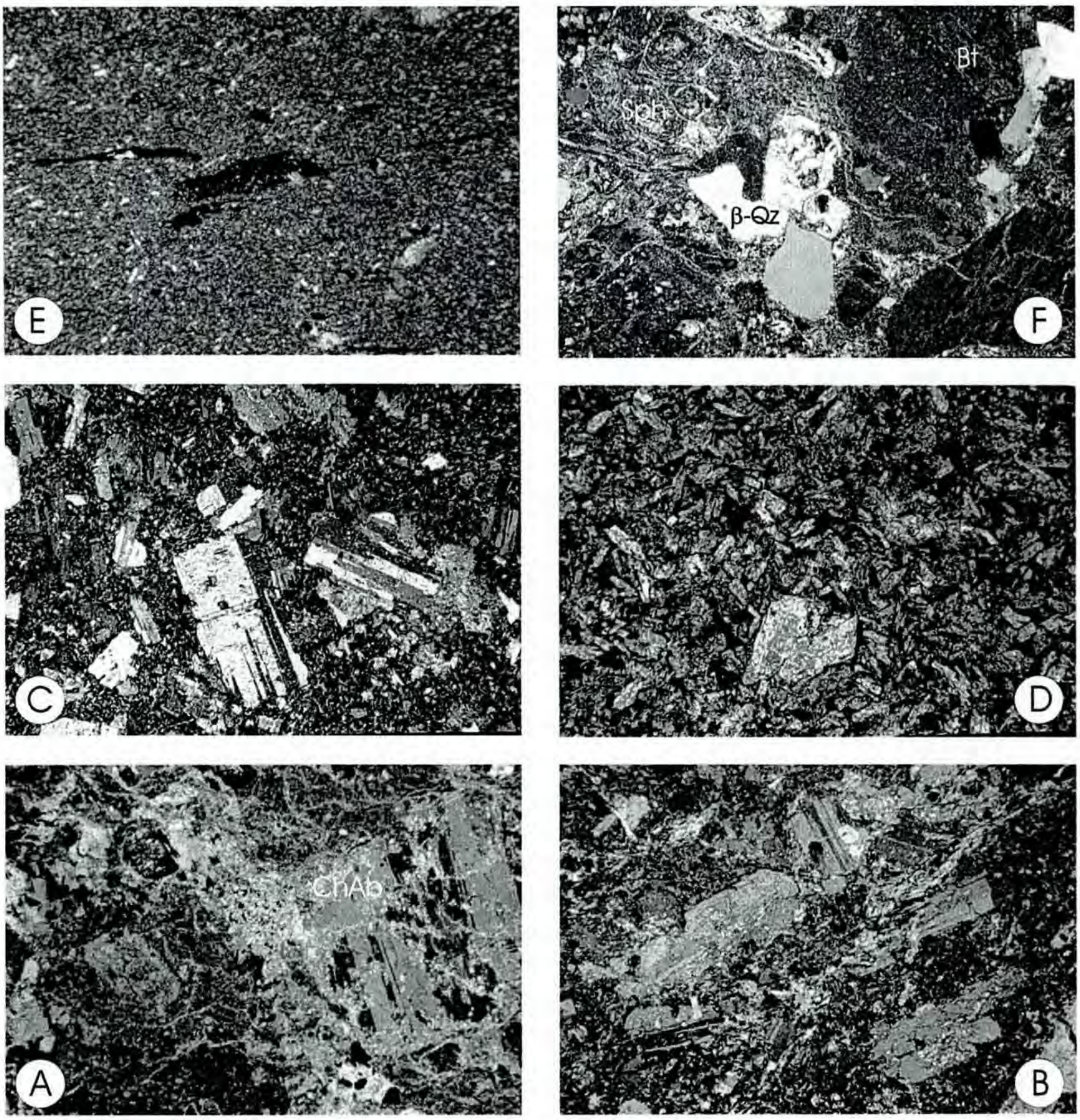

Fig. 3 – Representative photomicrographs of facies described in the Orobic Anticline. A, B = weathered intermediate ignimbrites (benmoreites; V1, Pradini Valley section), displaying feldspar phenocrysts locally transformed into chessboard-albite (ChAb); C = fresh trachybasalt? lava flow (V3, Caprile area); D = sericitised mugearite lava flow (V3, Pradini Valley section) with a plagioclase lathwork encasing large calcite-chlorite pseudomorphs after femic phenocrysts (amphiboles ?); E = black shale (S3, Pradini Valley section) including imbricated carbon-rich chips; F = very coarse-grained, moderately sorted volcanic arenite (S4, Pradini Valley section) rich in volcanic lithics displaying spherulitic structures (Sph) and containing embayed quartz (β-Qz) as well as leached biotite (Bt) phenocrysts. All photos are in cross-polarised light, magnification 14x, except F (9x).

clasts are usually flattened due to the Alpine deformation. Cinerites form lenses interbedded in the breccias.

S_1 – Medium- to coarse-grained, grey to pink volcanic arenites, displaying abundant fresh detrital feldspar and scattered white mica flakes when examined with the hand lens; white mica flakes are locally concentrated on the bedding planes in the finer-grained fraction. Layers are up to several tens of cm-thick, mostly homogeneous in grain size but locally displaying a rough grading, either normal or inverse. Subordinate intercalations of pelites, pyroclastics and conglomerates occur.

S_2 – Prevailing fine- to medium-grained, dark grey volcanic arenites, in layers a few cm to less than 20 cm-thick, fining upwards to black pelites rich in white mica flakes (sand/mud ratio > 1). A variety of sedimentary structures (load casts, commonly asymmetrical but locally symmetrical ripple-marks, mud-cracks ...) is displayed.

S_3 – Grey to black siltstones and shales with subordinate intercalations of fine-grained sandstone (sand/mud ratio >1), locally transformed into flagstones and slates by anchizonal conditions reached during the Alpine Orogeny.

S_4 – Pebble to cobble conglomerates, alternating to coarse-grained pebbly sandstones, deposited in proximal alluvial fans. In the Pizzo dei Tre Signori area this facies has been formally introduced as a distinct unit, occupying a precise stratigraphic position (Ponteranica Conglomerate; Casati & Gnaccolini, 1967), whereas in the Trabuchello-Cabianca Anticline the term "Monte Aga Conglomerate" has been recently introduced (Cadel *et al.*, 1996). Distinct conglomerate bodies characterised by peculiar clast composition should be mapped separately.

S_5 – Coarse-grained arenites to conglomerates with Fe-carbonate interlayers up to 50 cm-thick, that might represent lacustrine carbonates and evaporites.

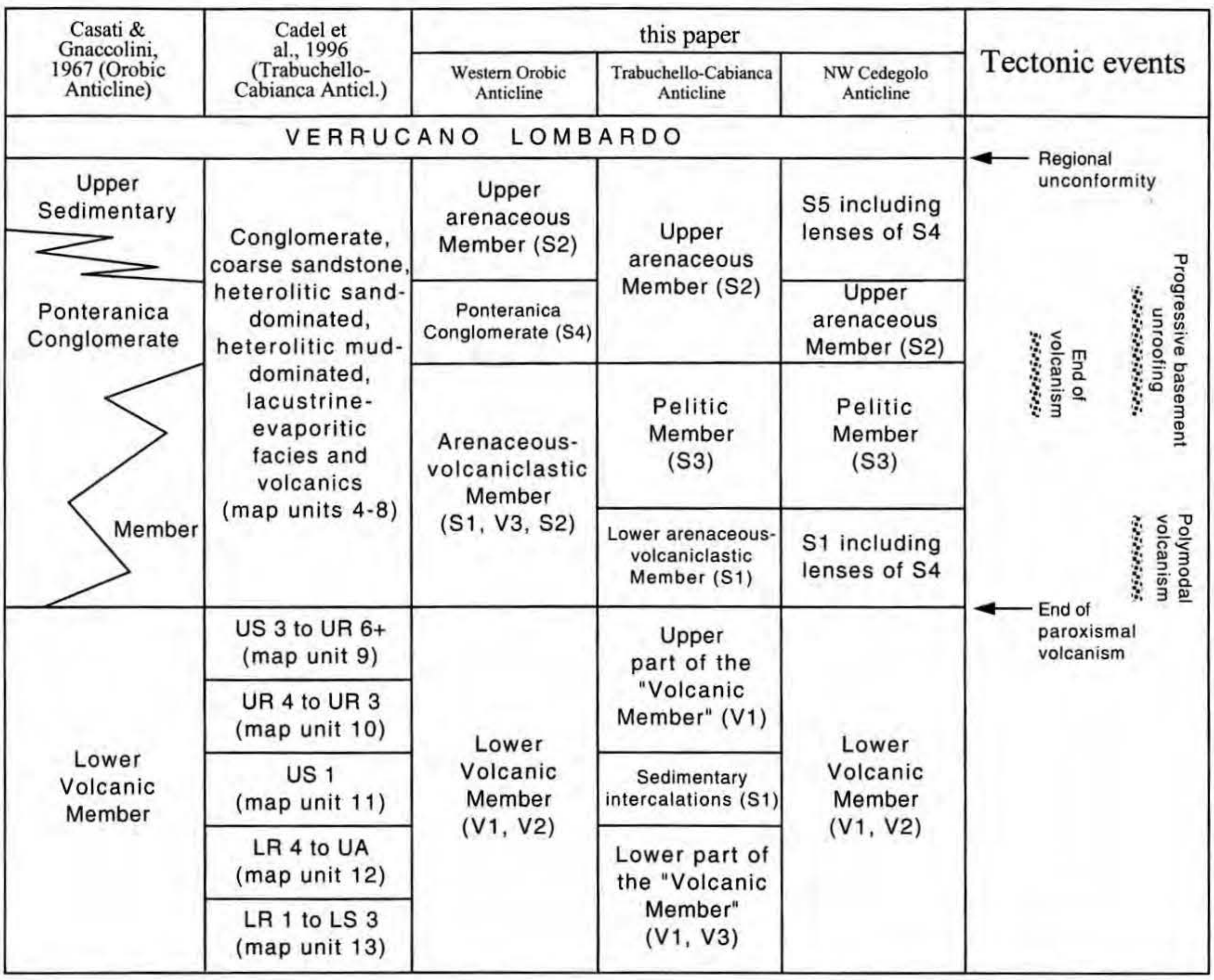

Fig. 4 – Simplified scheme depicting the vertical organisation of the mapping lithofacies proposed for the Collio Fm. in the present paper, compared with the reference frameworks available from the literature. Indications for interpreting the proposed lithofacies as a result of tectonic and magmatic events are also provided. Correlation of mapping lithofacies across adjacent anticlines is a mere graphic device; in fact, the assumption of a physical continuity among the Early Permian basins presently exhumed in the three Orobic Anticlines has to be regarded as undemonstrated, and stratigraphic correlation would require much better age constraints than available.

PIZZO DEI TRE SIGNORI AREA

Although severe faulting and folding of the Collio Fm., as well as poor exposure of its sedimentary upper part, commonly hamper a detailed measurement of this unit from base to top, the sandier facies exposed in the Orobic Anticline allowed description of two stratigraphic sections (Casati & Gnaccolini, 1967). In the reference Pradini Valley section, nearly 1000 m-thick, facies V_1, V_2, S_1, V_3, S_2 and S_4 (in first appearance order) are represented: V_1 ignimbrites are essentally intermediate Na-alkaline products (benmoreites), and V_3 mugearite flow, 20 m-thick, is found 92 m above the oldest sediments. Petrographic analysis on sandstones sampled both in the Orobic and Trabuchello-Cabianca Anticlines revealed that, despite marked lateral variability of facies and wide grain size range, composition of the Collio sandstones is remarkably uniform over an area extending for more than 35 km west to east and cluster into a single petrofacies (P0 of Sciunnach *et al.*, 1999). No significant stratigraphic trends were recognised, consistent with the short time span recently indicated for the deposition of the Collio Formation in the Val Trompia Basin (Schaltegger & Brack, 1999).

MONTE CABIANCA-PIZZO DEL DIAVOLO DI TENDA AREA

A few transverses, from the northern slopes of Monte Cabianca to the Lago Rotondo and up to the Monte Aga - Pizzo del Diavolo divide, allowed us to group the lithozones and facies of Cadel *et al.* (1996) into the following mapping units:

1. Lower part of the "Volcanic member" of the Collio Fm., corresponding to lithozones LR1÷LR5 + UA of Cadel *et al.* (1996) and to facies V_1, V_3 of the present paper.
2. Sedimentary intercalations in the "Volcanic member": prevailing volcanic arenites, commonly in fining-upwards beds with basal conglomerate lags. This interval corresponds to lithozone US1 of Cadel *et al.* (1996) and to facies S_1 of the present paper.
3. Upper part of the "Volcanic member": lapilli welded tuffs (ignimbrites), more or less lithified, with frequent epiclastic intercalations. This interval represents lithozones UR1÷UR6 of Cadel *et al.* (1996) and facies V_1 of the present paper.
4. Lower arenaceous-volcaniclastic member: prevailing coarse-grained arenites, alternating to epiclastics and tuffs displaying a more pervasive cleavage. This member is poor in sedimentary structures and, in the study area, exhibits an overall higher strain with respect to the upper arenaceous-volcaniclastic member. It largely corresponds to the "thick-bedded coarse sandstone facies" of Cadel *et al.* (1996) and to facies S_1 of the present paper.
5. pelitic member ("heterolitic mud-dominated facies"; Cassinis *et al.*, 1986): prevailing dark pelites that preserve sedimentary structures with thin intercalations of volcanic ashes. This member commonly displays a pervasive slaty cleavage related to transpositive folding; corresponds to facies S_3 of the present paper.
6. upper arenaceous-volcaniclastic member ("heterolitic sand-dominated facies"; Cassinis *et al.*, 1986): prevailing grey to greenish sandstones, largely medium to fine-grained, rich in sedimentary structures. Corresponds to facies S_2 of the present paper.

Sandstones and pelites pass laterally to heteropic conglomerates ("conglomerate and pebbly sandstone facies"; Cassinis *et al.*, 1986; corresponding to facies S_4 of the present paper), the composition of which substantially reflects the overall composition of sandstones, once the grain size-induced variations are taken into account.

CEDEGOLO ANTICLINE

The development of Alpine folds, thrusts and pervasive cleavage makes stratigraphic reconstruction of the Collio succession quite difficult (Albini *et al.*, 1994). Three main tectonic units can be distinguished at map scale, each of them displaying a peculiar sedimentary succession (see also the companion paper by Forcella & Siletto, this volume).

In the northern and structurally higher tectonic unit, the following lithofacies can be observed (from bottom to top): volcanics (V_1), coarse heterolitic conglomerates with arenaceous matrix (S_4), prevailing fine- to medium-grained dark grey arenites (S_1), prevailing black pelites with subordinate arenites (S_3), grey to greenish arenites with tabular and cross-lamination (S_2) and coarse-grained conglomerate bodies bounded by synsedimentary faults (S_4).

In the intermediate tectonic unit, massive to roughly-bedded ignimbrites, highly variable in thickness and degree of welding (V_1), are followed upsection by volcanic arenites with variable thickness (S_1), by siltstones and shales (S_3), prevailing fine- to medium-grained arenites (S_2), and eventually by coarse-grained arenites with Fe-carbonates interlayers (S_5).

In the southern tectonic unit, ignimbrites and welded tuff (V_1) are followed upsections by "epiclastics" (V_2) and agglomerates to cinerites (V_4), heteropic with S_2, S_3 and S_5. In both the intermediate and southern units (P.zo Recastello, M. Cimone and M. Gleno) boudinaged levels of ferroan carbonates (facies S_5) are intercalated in facies S_2 and S_3.

CONCLUSIONS

Marked lateral variability of facies, thickness, geochemistry of volcanic products and subsidence patterns in the different sectors of the wide area in which the Collio Formation was deposited indicates lithofacies mapping as the most viable tool to unravel stratigraphic and geometric relationships among the Collio lithosomes. Organisation of mapping lithofacies into a coherent stratigraphic framework, eventually leading to a stratigraphic revision of the Collio Fm., appears to be still premature at the moment and requires further detailed field work, as well as better age constraints.

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STRUCTURE AND STRATIGRAPHY OF THE PERMO-CARBONIFEROUS COVER AND VARISCAN METAMORPHIC BASEMENT IN THE NORTHERN SERIO VALLEY (OROBIC ALPS, SOUTHERN ALPS - ITALY): RECOGNITION OF PERMIAN FAULTS

FRANCO FORCELLA [1] and GIAN BARTOLOMEO SILETTO [2]

Key words – Alpine tectonics; Permian extensional tectonics; Southern Alps; Lombardia.

Abstract – An area belonging to the South-Alpine domain of the Lombardic Orobic Alps (Italy) is considered, in particular the eastern sector of the Anticlines belt that represents a major structural partition of the chain. It is formed by a polydeformed (D_1-D_2)Variscan metamorphic basement unconformably covered by Late Paleozoic sedimentary and volcanic sequences.
Alpine compressional events are related to southward-verging thrusting phases (D_3). The regional thrust geometry is outlined by large cuspate synforms of sedimentary rocks, locally boudinaged along cataclastic zones. A D_3 deformation phase produced a crenulation cleavage in basement rocks and a pervasive cleavage in the sedimentary cover, with axial planar to meso- and megascopic folds, which often makes stratigraphic reconstruction difficult. Moreover, the spatial correlation of individual thrust planes is often hampered by variously intersecting faults which transfer or interrupt the individual thrust sheets.
None the less, three main units can be distinguished on the map, displaying sedimentary successions that change in thickness and lithofacies in relation to the main Alpine thrusts, which therefore represent inverted Permian normal faults.

Parole chiave – tettonica alpina; tettonica estensionale permiana; Alpi Meridionali; Lombardia.

Riassunto – In questa nota viene descritto l'assetto stratigrafico e tettonico di un tratto del settore nord-orientale della catena sudalpina lombarda. Esso fa parte dell'unità strutturale denominata, con terminologia semplicistica, "Anticlinali Orobiche".
Queste sono costituite da un basamento cristallino varisico rappresentato da varie litofacies e una storia metamorfica evolutasi dalla facies anfibolitica a quella scisti verdi e da una copertura sedimentaria terrigena e vulcanica.
Tale copertura, di età Carbonifero sup.(?)-Permiano, si è deposta in ambiente continentale subaereo e/o fluvio-lacustre e mostra evidenti variazioni sia nello spessore che nelle litofacies in corrispondenza delle principali superfici di sovrascorrimento d'età alpina.
Questo fatto induce a pensare che esse rappresentino l'inversione di faglie normali prodotte dalla tettonica transtensiva tardo-varisica, responsabile, alla scala locale, della formazione dei bacini ospitanti le suddette successioni stratigrafiche. Nell'area sono state distinte tre principali unità tettoniche generate dalla fase compressionale alpina.
Esse sono grossolanamente allineate in senso WSW-ENE, caratterizzate da (i) piegamenti poliarmonici sud-vergenti a diversa scala, intersecati da clivaggio di piano assiale (D3) regionalmente persistente, e (ii) ritagliate da *thrust* minori e sistemi di faglie successive.
Una fase tettonica più recente (tardo- e post-alpina) è rappresentata da sistemi di faglie estensionali ad elevata inclinazione o da piani di taglio pellicolari che sezionano pinnacoli rocciosi in precaria staticità. Nel suo insieme, il tratto di catena esaminato mostra l'impronta di un diffuso collasso gravitazionale.

INTRODUCTION

In the northern part of the southern Orobic Alps, in a roughly EW-trending belt, a Carboniferous(?) to Lower Permian terrigenous-volcanic succession lies unconformably on the Variscan metamorphic basement. The general stratigraphic succession, starting with the "Basal Conglomerate" (Upper Carboniferous? – Lower Permian), is mainly represented by the volcaniclastic - lacustrine Collio Fm. (Lower Permian), suggesting as a whole a mobile, transtensive depositional environment, followed by the clastic Verrucano Lombardo (Upper Permian) and the mixed clastic-carbonate Servino (Scythian – Lower Anisian). The belt is presently structured in a series of *en*

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échelon ENE-trending anticlines (De Sitter, 1963; De Sitter & De Sitter-Koomans, 1949, and following literature) which roughly correspond to former Permian basins inverted during the Alpine orogeny. The junctions between the anticlines are areas of intense and complicated deformation. In this paper we investigate the region of the junction between the Trabuchello-Cabianca anticline (to the W) and the Cedegolo anticline (to the E) in the upper Serio Valley (Fig. 1). Here, three main tectonostratigraphic units, separated by inverted extensional faults, can be recognised, each characterised by a specific sedimentary succession and geometric setting. The lithofacies classification used in the following is that proposed by Forcella *et al.* (this volume). In particular, in this paper and in Fig. 2, both the volcanic, volcaniclastic and the sedimentary facies of Lower Permian age are categorised into the Collio Fm. following the mapping tradition in the Orobic (or Brembano) Collio basin.

THE NORTHERN TECTONOSTRATIGRAPHIC UNIT

In a huge area of the northern unit (to the E), micaschists, gneisses and minor amphibolites crop out, in which a polyphase Variscan metamorphic evolution is recorded (Albini *et al.*, 1994; Spalla *et al.*, 1999), from amphibolite facies (D1 relict folds and schistosity) to greenschist facies (D2 meso- and macro-scale folds and pervasive foliation). To the west of the Malgina Valley, the basement is capped by the Permian volcanic and clastic succession (Fig. 2). The stratigraphic succession normally starts with pelitic to arenaceous purplish beds, often burrowed and rich in detrital micas. This lithofacies appears at various levels in the Lower Permian succession of the Southern Alps (*e.g.* "Pietra Simona" member of the Dosso dei Galli Fm.). Coarse conglomerates, mainly composed of quartz pebbles in an arenaceous matrix, with a rough bedding, over-

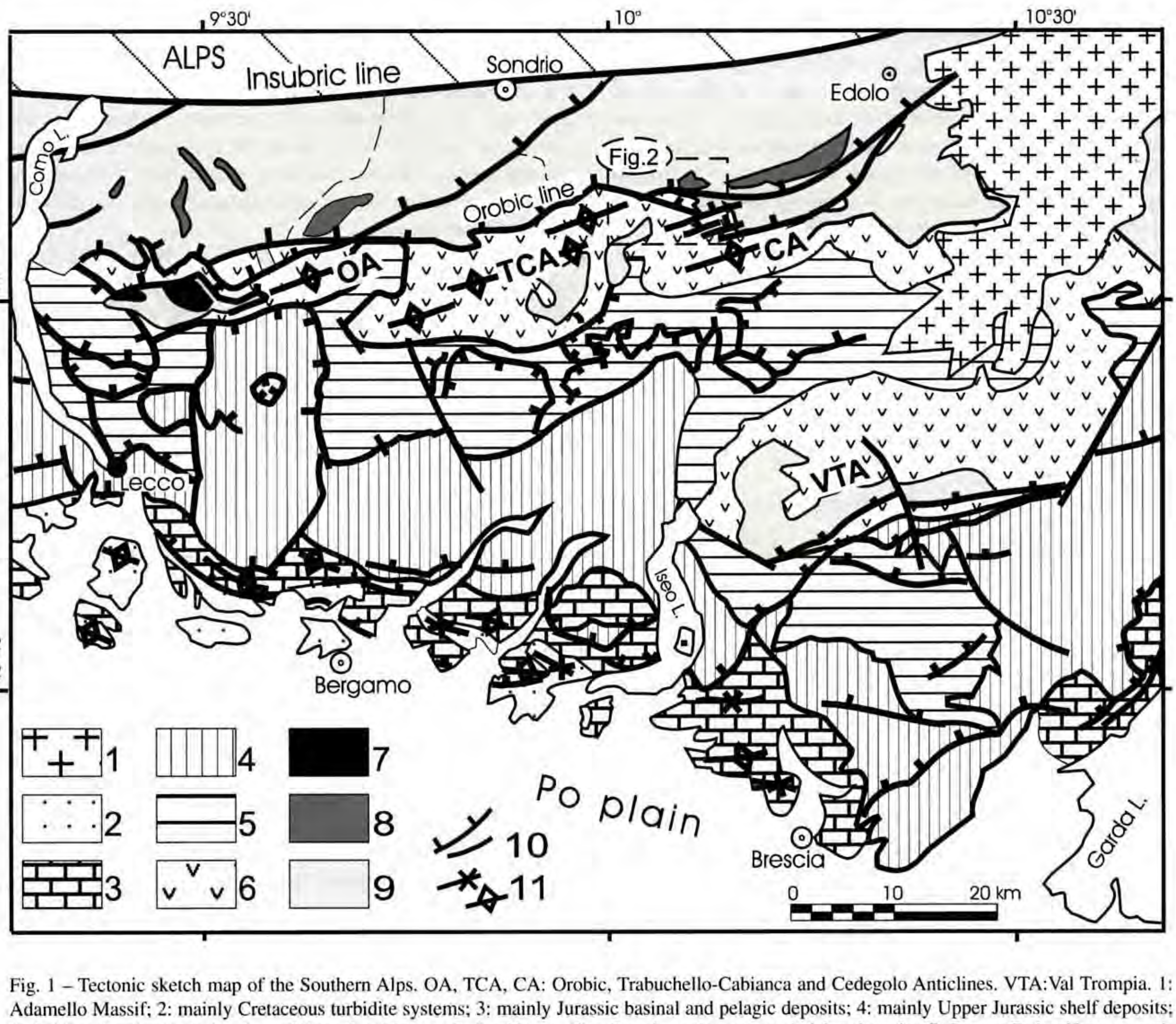

Fig. 1 – Tectonic sketch map of the Southern Alps. OA, TCA, CA: Orobic, Trabuchello-Cabianca and Cedegolo Anticlines. VTA:Val Trompia. 1: Adamello Massif; 2: mainly Cretaceous turbidite systems; 3: mainly Jurassic basinal and pelagic deposits; 4: mainly Upper Jurassic shelf deposits; 5: Anisian-Carnian deposits; 6: undifferentiated Upper Carboniferous (?) – Permian – lowermost Anisian deposits; 7: Permo-Carboniferous granitoids; 8: Variscan metagranitoids; 9: Variscan metamorphic rocks; 10: main thrusts and faults; 11: axial surface trends of main folds.

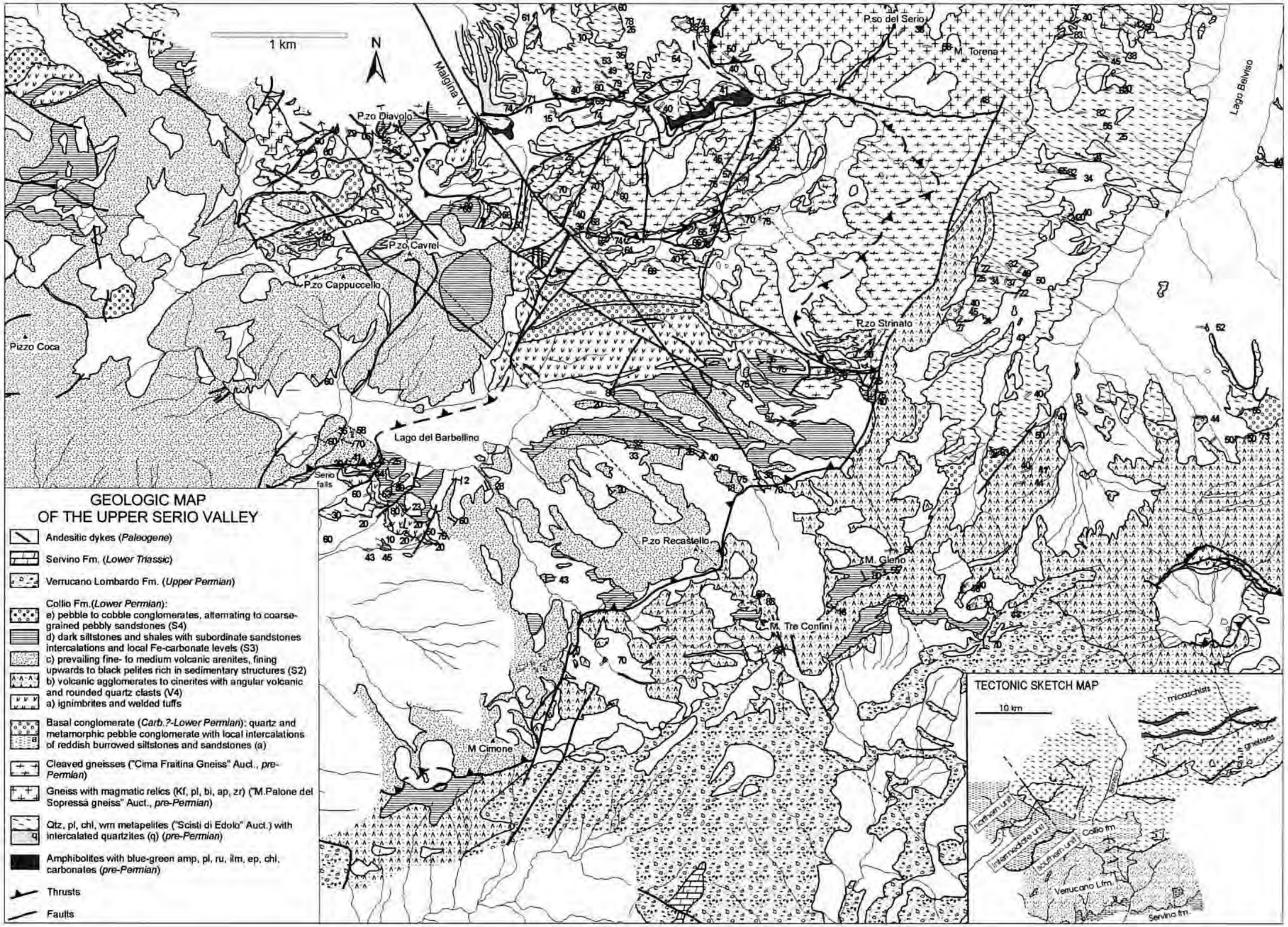

Fig. 2 – Geological map of the Upper Serio Valley.

lie or are interfingered with the purplish pelites. In spite of the local paucity of metamorphic pebbles, this unit is considered to represent the "Basal Conglomerate" (Dozy, 1933; Dozy, 1935). Upwards, the conglomerates are capped by a thin volcanic and volcaniclastic layer, which makes the transition to a several hundred metres thick arenaceous sequence (lithofacies S2). Toward the top the sandstones become finer and the beds thinner, and parallel and cross laminations, mudcracks and raindrop casts are common. This arenaceous-pelitic succession forms the highest peaks of the Orobic Alps. Locally, conglomeratic wedges, rich in quartz, metamorphic and volcanic pebbles to cobbles (lithofacies S4), are connected to synsedimentary fault scarps along the SE wall of Pizzo di Coca. These conglomerates resemble both in composition and stratigraphic location the Dosso dei Galli (Assereto & Casati, 1965) or Ponteranica formations (Casati & Gnaccolini, 1965), widely cropping out more to the east (Val Trompia) and to the west (Orobic anticline), but never reported in this sector of the Orobic Alps.

Alpine compressional events, related to southward-vergent thrusting phases, often make stratigraphic reconstruction difficult. The Alpine deformation has produced a crenulation cleavage in basement rocks and pervasive cleavage in the sedimentary cover, axial planar to EW-trending meso- and megascopic folds (D_3), regional thrust surfaces outlined by large cuspate synforms of sedimentary rocks, locally boudinaged in pieces along cataclastic zones, and sets of shear planes, gently dipping to the north, which bear qz-slickenfibres indicating a top-to-the-SSE translation (D_4). Along any shear plane the dislocation is only a few centimetres to a few metres, but the shear induces a sigmoidal distortion of the D_3 cleavage and, in finegrained layers of Collio and Verrucano Lombardo formations, a few mm to cm-spaced D_4 crenulation cleavage.

The general plunge of D_1 to D_3 structures towards the west is not sufficient to explain the lack of sedimentary cover to the east, which can be better understood by considering the effects of Permian NE-SW-trending extensional faults. These produced a structural depression to the NW (present-day coordinates) in which a thick volcanic and clastic succession accumulated. In particular, the gently westward-dipping tectonic surface exposed near the base of the vertical cliff of the Serio River waterfall and further to the west along the northern side of the Serio Valley, brings into direct tectonic superposition the Collio sedimentary sequence and the crystalline basement; this geometry and the occurrence of kinematic indicators suggesting a normal sense of movement (top to the W), preserved near the Serio waterfall, suggests the extensional kinematis of a pre-Alpine master fault. The kinematics along the extensional fault system was inverted during the Alpine compressional phase, and some segments are now represented by top-to-SSE oblique faults, with different geometries along strike.

East of the Malgina Valley, the northern tectonostratigraphic unit can be subdivided into some thrust slices underlain by local networks of Fe-carbonate veins and cataclastic zones up to tens of metres thick, which can be followed for hundreds to thousands of metres. West of the Malgina Valley, a tantalising belt of fault-bounded outcrops align near the base of the northern unit (Fig. 2); they represent slices of the formerly faulted margin that were subsequently dragged and/or tilted southward during the Alpine compression. In the sedimentary cover, west of the Malgina Valley, a set of steeply dipping, south-verging D_3 folds progressively lower the stratigraphic units to the south and follow, at the surface, the steep structural attitude of the

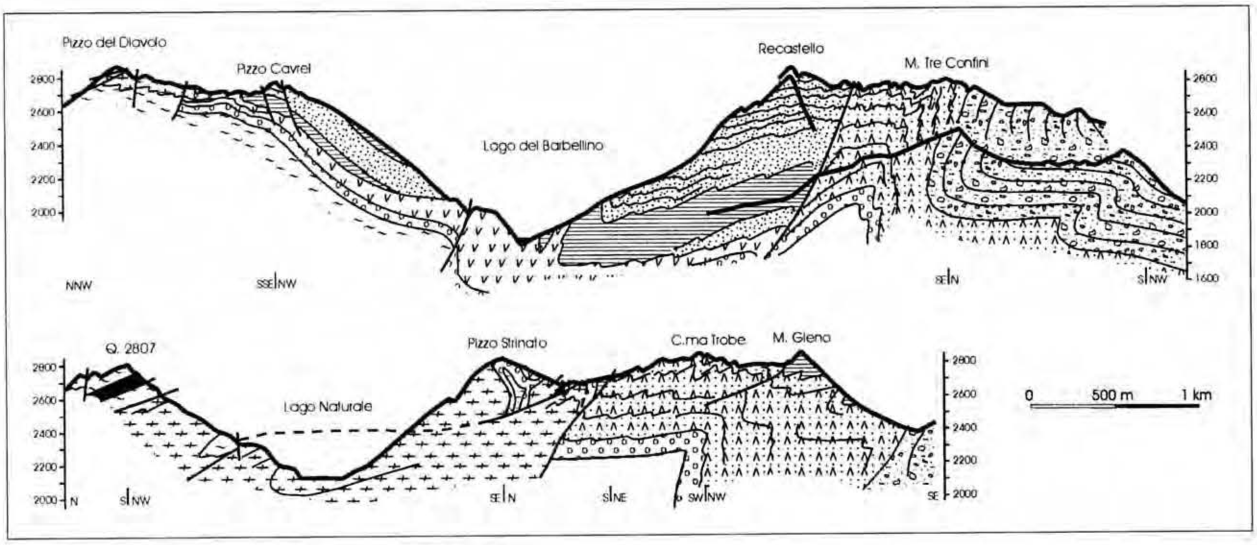

Fig. 3 – Schematic cross sections. Symbols are as in Fig. 2.

basement. The fold belt is best developed in the arenaceous-pelitic facies that form the southern slopes of Pizzo Cavrel and Pizzo Cappuccello. D_3 folds are cut by andesitic dykes related to the Adamello Tertiary intrusion, representing the main time constraint for the Alpine deformation.

Superficial shear planes, steeply dipping to the S and SW, dissect the Alpine structure, sometimes dragging the D_3-related cleavage into anomalous attitudes, and favouring deep-seated gravity-driven deformation and surficial landslides. As a whole, the entire ridge shows evidence of diffuse gravitational collapse.

THE INTERMEDIATE TECTONOSTRATIGRAPHIC UNIT

In the intermediate unit the metamorphic basement is unconformably overlain by a sedimentary succession consisting of milky quartz, metamorphic and volcanic pebbly conglomerate interbedded with purplish-red, often burrowed siltstone ("Basal Conglomerate"), very similar to that of the northern unit, followed by rhyolitic ignimbrites ("lower volcanics"). Volcaniclastic layers interbedded with grey sandstones and shales turning into well-bedded dark-grey siltstones and shales (facies S3) constitute the "lower Collio" succession. Well bedded dark greyish-greenish shales and sandstones, locally interbedded with conglomerates form the "upper Collio" succession (facies S2).

From a structural point of view this unit is characterised by a large-scale synformal-antiformal pair verging to the south, linked by a gently dipping to the north intermediate limb. Asymmetric parasitic folds, up to tens of metres in amplitude, further deform in a systematic way the limbs of the first-order folds (Fig. 3). In the northern upright limb of the main synform, large-scale "lenses" of Basal Conglomerate and Collio Fm. pelitic facies (S3) are intercalated in the intermediate volcanic and volcaniclastic layers; we interpret these "lenses" as the outcropping cores of parasitic cuspate folds that raise from the top of the Basal Conglomerate or hang from the bottom of the Collio Fm. pelitic facies. We cannot, however, exclude the presence of stratigraphic intercalations between the volcanic and sedimentary facies of the Collio Fm. The gently dipping intermediate limb crops out in the northern slope of Pizzo Recastello; it is mainly formed by the arenaceous facies deformed by a train of mesoscopic S-verging folds. In proximity to the Pizzo Recastello thrust surface, it bends into a tight antiformal structure with a vertical to overturned limb. The thrust surface fades away to the west along the western slope of Monte Cimone; to the east it merges into a NNE-SSW-trending fault that possibly acts as a lateral ramp for the intermediate thrust unit. Local complications of the structure are widespread; as an example, the upper slope of Pizzo Strinato on tight disharmonic folds and faulted folds are the consequence of a local thrust surface that intersects the ridge between the Belviso and the Serio valleys. The thrust can be followed for a few hundred metres down the western slope of the peak, from which its continuation can only be supposed to continue within the crystalline basement (dashed line on the map).

THE SOUTHERN TECTONOSTRATIGRAPHIC UNIT

The sedimentary succession of the southern thrust unit mainly consists of volcaniclastics of various grain-sizes (from coarse pebbly conglomerates to very finegrained cinerites, lithofacies V4) with locally interbedded qz-rich conglomerates and agglomerates, in places arenite-matrix supported. Grey to black siltstones and shales (facies S3; Mt. Tre Confini, Mt. Gleno) and coarse-grained arenites to conglomerates with Fe-carbonate interlayers of possible lacustrine origin (facies S5) are locally interbedded in the volcaniclastic succession, which is directly overlain by the Upper Permian Verrucano Lombardo reddish conglomerates and sandstones, and therefore ascribed to the "upper Collio". This succession could reasonably represent a marginal part of the basin, deepening toward the NW.

This structural unit is characterised by a train of south-verging mega- to mesoscopic folds, also involving the usually brittle Verrucano Lombardo Fm. A locally very pervasive D_3 cleavage flattens the volcanic pebbles of the conglomeratic layers into ellipsoidal shapes. The unit plunges to the south towards the Dezzo Valley, forming a wide antiform bending (Cedegolo Anticline *p.p.*) whose toe wedges into the carbonatic belt cropping out to the south.

CONCLUSIONS

The studied area is considered an example of the polyphase tectonics that characterise the belt of Orobic Anticlines, where the Upper Carboniferous (?) to Upper Permian stratigraphic succession of the Lombardian southern Alps extensively crops out. The late- to post-Permian extensional phase is suggested by Alpine thrust surfaces that follow abrupt discontinuities in thickness and facies of the stratigraphic record; the changes point to a northward deepening of the original basin with a step-like geometry. With reference to the northern main thrust surface, the original extensional kinematics are further suggested by a "younger over older" thrust (Serio River waterfall, and further west, outside the mapped area) and by alignments of tectonic slices, variously imbricated and/or tilted (between the Malgina Valley and the tail of Barbellino Lake), which we interpret as former horses of a main extensional boundary fault that was inverted and tilted during the Alpine compression. Minor listric normal faults dipping to the east (antithetic?) bound conglomeratic wedges comparable to the Upper Permian conglomerates; the conglomerates are still preserved with their original geometry along the eastern slopes of the Coca and Redorta peaks.

The Alpine compressional phase(s) gave rise to the main structural grain of the area, expressed by the fold-thrust sheets dealt with previously. The main phase is older than the age of the transecting andesite dykes (radiometric ages of 50-64 Ma according to Zanchi *et. al.*, 1990 and Fantoni *et al.*, 1997). A younger phase is expressed by D_4 shear planes, pervasive at the map scale, expressed both in finegrained sediment and in coarse volcaniclastic conglomerates; these structures have been described by Albini *et al.* (1994). No definite upper-age constraint is available for this event.

The late- to post-Alpine extensional phase is documented by high-angle N-S-trending faults, well expressed along the main mountain ridges, and by map-scale, near-vertical shear planes that distort the trend of the regional Alpine folds and cleavage surfaces and create huge rock pinnacles and impending rock masses ready to slide down the slopes below. Huge boulders caused by a rock-slide are now scattered in the nearby Maslana village (Valbondione) and evidence of deep-seated gravitational collapse is documented on the northern slope of Mt. Toazzo ridge (located immediately SW of Mt. Cimone).

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GEOLOGICAL SETTING OF THE BRESCIAN ALPS, WITH PARTICULAR REFERENCE TO THE PERMIAN OUTCROPS: AN OVERVIEW

PAOLO SCHIROLLI [1]

Key words – Brescia Province; N-S geological profile; Stratigraphy; Tectonics; Southern Alps; Northern Italy.

Abstract – Eight geological and structural zones are recognised in the Brescia Province on the basis of their prevalent lithologic pattern, different age and structural style, as due to the Alpine orogeny.
Regional tectonic lineaments or stratigraphical boundaries divide these zones, shown both in a simplified geological map of the region and in a complete N-S trending geological profile, 80 kilometres long, through the entire Brescia Province.
The profile crosses the igneous and sedimentary continental deposits of two of the major Permian basins in the region (Orobic and Collio Basins).
They corresponded to continental intramontane basins with a WSW-ENE orientation, fault-bounded by basement structural highs.

Parole chiave – provincia di Brescia; profilo geologico N-S; stratigrafia; tettonica; Sudalpino; Italia settentrionale.

Riassunto – La provincia di Brescia è stata suddivisa in otto zone geologico-strutturali, distinguibili sulla base dei litotipi prevalenti, dell'età delle formazioni e del differente stile strutturale assunto dalla successione in conseguenza degli eventi compressivi alpini. Lineamenti tettonici di importanza regionale o superfici stratigrafiche delimitano le zone geologico-strutturali, la cui estensione areale è riportata nella carta geologica semplificata della provincia, che accompagna un profilo geologico della lunghezza di circa 80 chilometri che attraversa da nord a sud l'intera provincia di Brescia. Il profilo attraversa i depositi continentali ignei e sedimentari dei due maggiori bacini deposizionali permiani della regione (Bacino orobico e Bacino di Collio). Essi rappresentano dei bacini continentali intramontani orientati in senso WSW-ENE, separati dagli alti strutturali del basamento ad opera di attivi sistemi di faglie sinsedimentarie.

INTRODUCTION

A geological profile across the entire Brescia Province, along a N-S trend, is here proposed in order to highlight the main geological features of this region. It is also supported by a concise explanation aimed at clarifying the local geological framework.

The profile (Plate 1), derived from published (Boni & Cassinis, 1973; Boni *et al.*, 1968, 1970, 1972; Bianchi *et al.*, 1971; Brack, 1984; Cassinis, 1980; Desio *et al.*, 1970; Schiavinato *et al.*, 1969; Cassinis & Castellarin, 1981; Cassinis & Forcella, 1981; Forcella, 1981; Cassinis & Castellarin, 1988) and unpublished (Brack; Schirolli) cross-sections and data on the Brescian Alps, links up a point to the north of the Tonale Line (part of the Insubric Line), near the Aprica Pass, to the last hills in front of the Po Plain, in Rezzato, a few kilometres to the east of Brescia. It allows better understanding of the geological setting of this part of Southern Alps, and the visualisation of the pellicular tectonic and stratigraphic pattern in which the Permian rocks crop out, immediately to the west of the Adamello batholith.

The most important regional faults and folds are represented along this N-S profile, which is 80 kilometres long; two other shorter cross-sections (from Brack) cut the western part of the Adamello intrusive massif, showing the structures superimposed by the magmatic Tertiary intrusion on the local stratigraphic succession of the Permian and Triassic cover.

GEOLOGICAL SETTING

A glance at the simplified geological map of the Brescia Province (Plate 1) shows the main geological features of this area. As in the whole Southern Alps, the Brescian Alps also reacted to the Tertiary Alpine compression through the imbrication of south to southeastward-verging tectonic blocks. This geological feature created the present pattern that shows older formations proceeding by degrees

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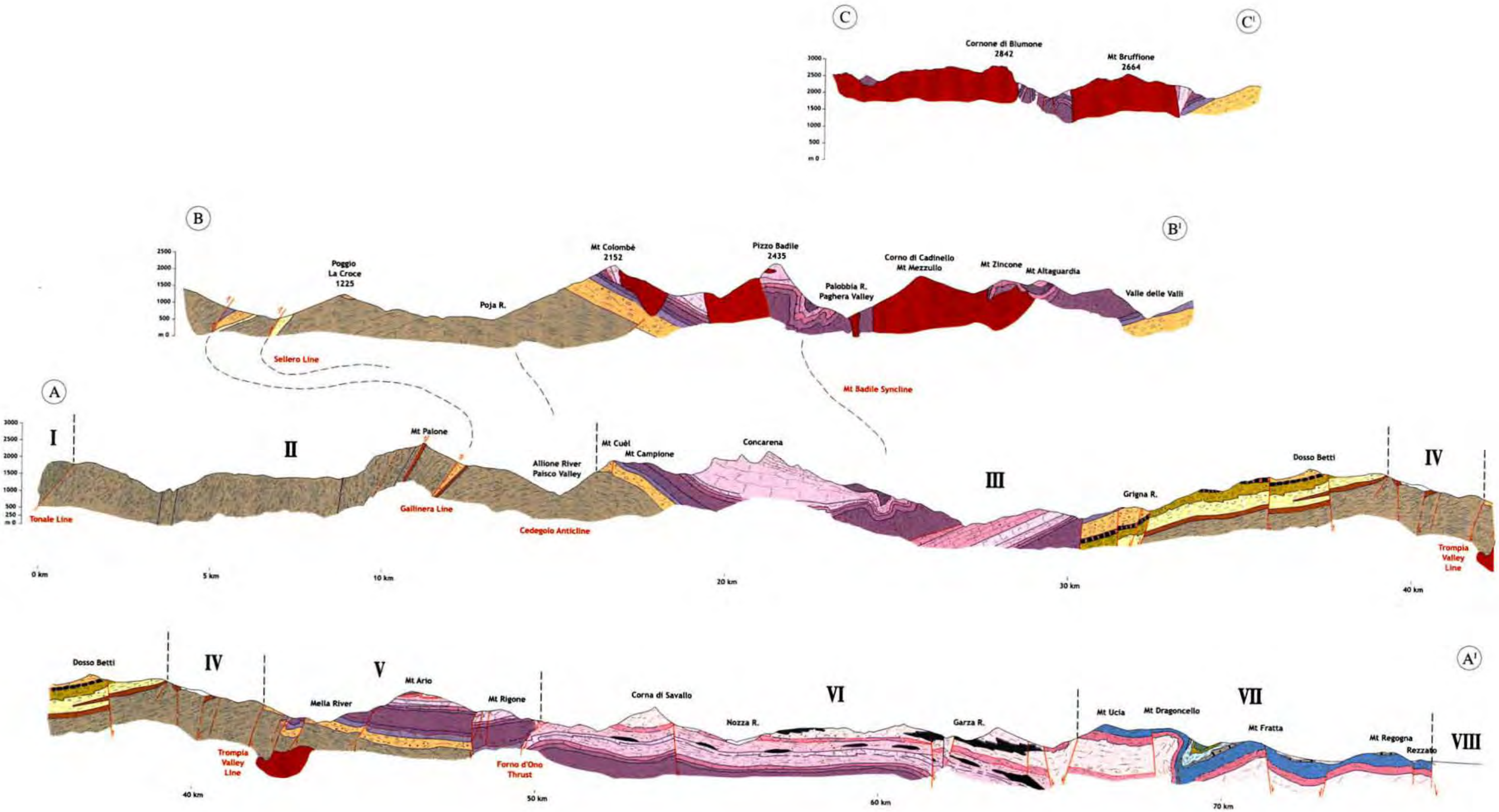
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Cornone di Blumone
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Poja R.
Mt Colombè
2152
Pizzo Badile
2435
Palobbia R.
Paghera Valley
Corno di Cadinello
Mt Mezzullo
Mt Zincone
Mt Altaguardia
Valle delle Valli
Sellero Line
Mt Badile Syncline
A
I
II
Mt Palone
Allione River
Paisco Valley
Mt Cuèl
Mt Campione
Concarena
III
Grigna R.
Dosso Betti
IV
Tonale Line
Gallinera Line
Cedegolo Anticline
Trompia
Valley
Line
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Dosso Betti
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Mt Ario
Mella River
Mt Rigone
Corna di Savallo
VI
Nozza R.
Garza R.
Mt Ucia
Mt Dragoncello
VII
Mt Fratta
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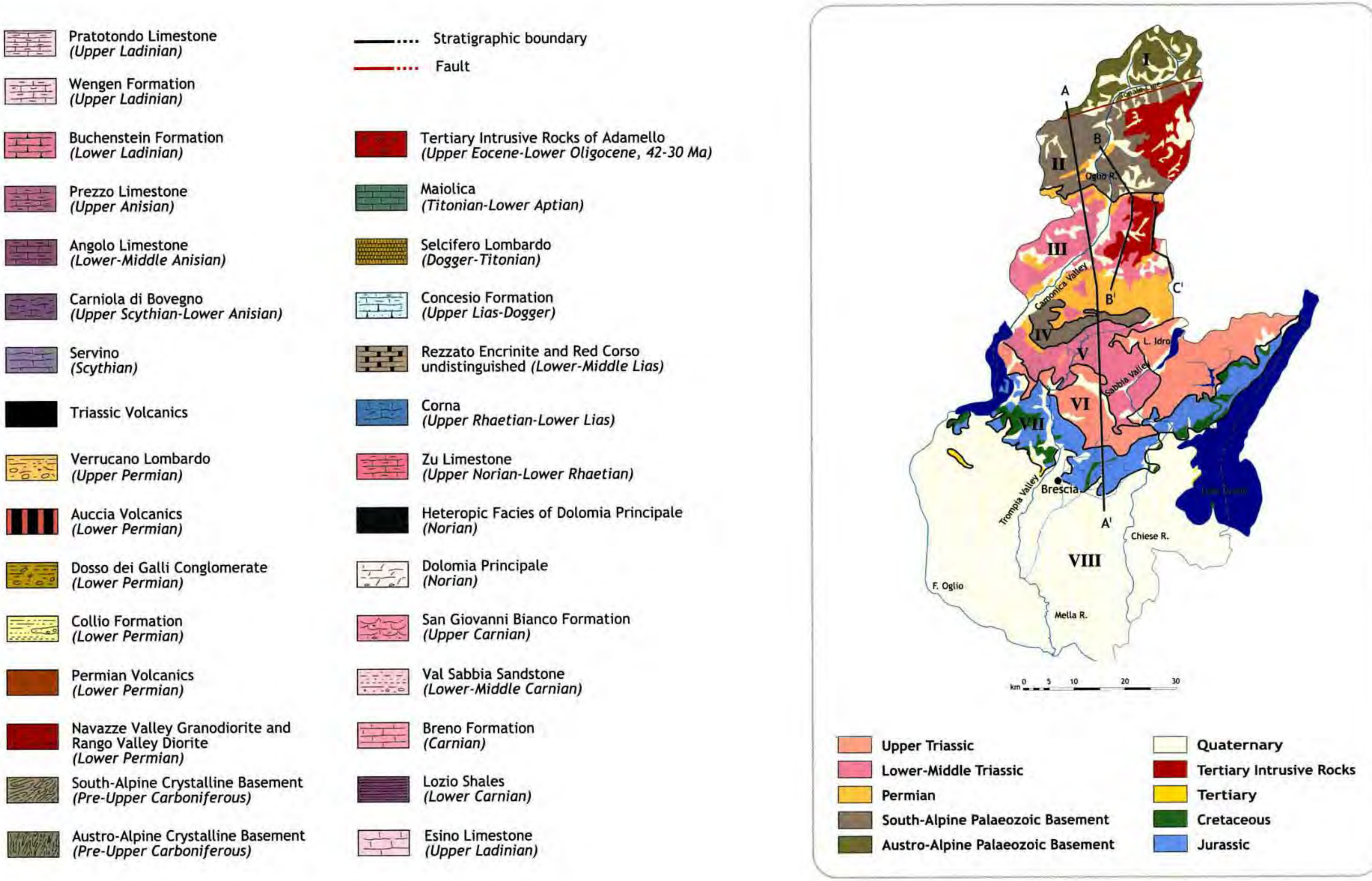

Plate I – N-S trending geological profiles across the Brescia Province. The simplified geological map of the Brescia Province shows the traces of the three profiles A-A', B-B', C-C'. Roman numbers, both in the small geological map and in the cross-section A-A', are referred to the geological and structural zones described in the text. Profiles B-B' and C-C' through the western Adamello intrusive massif derive from P. Brack unpublished data.

from the south (northern border of the Po Plain) to the north. In fact, from the south, we pass from the Quaternary alluvial deposits to the Tertiary and Mesozoic sedimentary cover of the piedmont, and from here to the first South-Alpine and then Austro-Alpine Palaeozoic metamorphic basement (Pre-Upper Carboniferous), widely cropping out to the north.

Part of South-Alpine basement, known as the "Brescia Three Valleys Massif", appears isolated between the Permian and Lower Triassic cover in the so-called "Trompia Valley Culmination".

The appearance of the Adamello batholith cuts off eastward both the South-Alpine basement and the Permian and Triassic sedimentary rocks, which crop out in the northern part of the Province. This magmatic body is represented by intrusive rocks emplaced during the Tertiary (from 42 to 30 Ma according to Del Moro *et al.*, 1986).

THE MAJOR TECTONIC DIRECTIONS

The Brescian Pre-Alps and Alps are characterised by the presence of two most important groups of tectonic lineaments. The first, with an E-W to ENE-WSW trend ("Orobic" or "Trompia" direction), is generally linked to south-verging thrust and fold Alpine structures; the second, NNE-SSW trend ("Giudicarie" direction), is connected to southeast-verging structures. These two groups of lineaments gave rise to a wide structural arc that involves both the outcropping formations of the region and the substrata below the alluvial deposits of the Po Plain (Castellarin & Sartori, 1980, 1983; Castellarin & Vai, 1986; Castellarin *et al.*, 1987; Picotti *et al.*, 1995).

The compressive regime that induced south-verging thrusts is ascribed unanimously to the structures with an Orobic trend, while the amount of transcurrent movement linked to the compression is still being defined for the Giudicarie-oriented structures (Trevisan, 1938; Laubscher, 1973, 1990; Castellarin & Sartori, 1983; Castellarin *et al.*, 1987; Doglioni & Bosellini, 1987).

The deformation of some folds with E-W trending axes, induced by the emplacement of the Adamello intrusion in the upper Camonica Valley (Cozzaglio, 1894; Brack, 1981), ascribes the Orobic direction to the first deformational phase at the origin of Brescian Pre-Alps such as the entire Southern Alps (Late Cretaceous Eo-Alpine Phase and/or Eocene Meso-Alpine Phase in age).

The Giudicarie structures were generated in a different (shortening axis rotated by about 45°) and subsequent deformational phase if compared with the above-mentioned Orobic-trending structures, *i.e.* during the Miocene Neo-Alpine Phase in the opinion of Castellarin *et al.* (1987); according to Doglioni & Bosellini (1987), they were produced by the Eo-Alpine sinistral transpressional movements along the Giudicarie Line. The southwards movement of both the Adamello crustal block and the Trompia metamorphic basement could explaine both the contemporary compression towards the south along the Trompia Valley Line and the sinistral transcurrent movement along the Giudicarie Line in the same deformation phase, following the Adamello intrusive emplacement (Middle Miocene to Messinian Neo-Alpine Phase, from Laubscher, 1990).

THE MOST IMPORTANT TECTONIC STRUCTURES

The main geological profile, N-S oriented, cuts all the major Orobic structures, firstly the Insubric Line. Under the name of the *Tonale Line*, it crosses the northern part of Brescia Province from Tonale (east) to the Aprica Pass (west), separating the northern metamorphic basements of the Austro-Alpine domain from the basement of the South-Alpine palaeogeographical-structural domain to the south.

The *Gallinera Line* (Cassinis & Castellarin, 1988) represents the eastwards termination of that important fault system (*Sellero Line* included) named the *Orobic Line*. South-verging high-angle reverse faults with north-dipping planes comprise this system. It permits the overthrust of the northern metamorphic rocks of the South-Alpine basement on to the southern Permian and Triassic cover of the north side of the Cedegolo Anticline. This structure, Orobic-oriented, formed because of the low competence of the "Carniola di Bovegno" formation.

The *Trompia Valley Line* is a complex E-W trending fault system that rose and overthrust southwards the "Brescia Three Valleys Massif" on to the Triassic cover rocks. The Trompia Line, together with the *South Giudicarie Line* (that part of the Line located to the south of its crossing with the Tonale Line), NNE-SSW oriented, give rise to a structural arc (Castellarin & Sartori, 1983; Castellarin *et al.*, 1987; Picotti *et al.*, 1995). Towards the eastern border of Adamello massif, the subvertical faults of the Giudicarie system allowed the significant uplift of the Brescia area in comparison with the eastern neighbouring sector of the Southern Alps (Castellarin & Sartori, 1980; Cassinis *et al.*, 1982).

To the south of the Trompia Line a large number of fault planes guided the thrust of the Lower-Middle Triassic rocks on to the southern Upper Triassic formations ("Dolomia Principale"). For example, the main section shows the important *Forno d'Ono Thrust*, by which the Anisian "Angolo Formation" was carried on to the younger "Sabbia Valley Sandstone", thanks to the deforming bed of the Scythian-Anisian "Carniola di Bovegno" formation.

In its turn, the wide "Dolomia Principale" plate overthrust the younger Mesozoic formations located to the south and east. Especially in evidence is the *Tremosine-*

Tignale Thrust, with a NNE-SSW orientation, running close to the western coast of Garda Lake, out of the range of the geological cross-sections presented here. The movement of this very thick dolomitic formation, favoured by the underlying Carnian evaporitic deposits, shows a south-vergence (Orobic trend) in the central-western Brescia Province, and a SSE-vergence (Giudicarie direction) in the eastern part of the Province. It forced the Jurassic and Cretaceous formations into folds of varying width, characterised by axes parallel to the regionally predominant structural directions.

GEOLOGICAL AND STRUCTURAL ZONES

With regard to its geological framework, the Brescia Province can be divided into eight geological and structural zones, bounded by regional tectonic lineaments and/or stratigraphical boundaries. The lithological pattern of the formations, their age and structural style caused by the compressional Alpine tectonic events are the major features distinguishing the eight zones below (Cassinis *et al.*, 1991).

Zone I is located to the north of the Tonale Line. It includes the metamorphic basement (generally pre-Late Carboniferous in age) of the Austro-Alpine palaeogeographical– structural domain.

Between the Tonale Line and the Cedegolo Anticline, the crystalline basement of the South-Alpine domain (locally known as the "Edolo Schists") occurs in *Zone II*. These pre-Upper Carboniferous metamorphic rocks appear as a strip immediately to the south of the Insubric Line in Lombardy. The southern boundary of Zone II is represented by the Gallinera Line. Along it, the basement thrusts southwards to cover stratigraphic units of the Cedegolo Anticline northern limb, belonging to *Zone III*. This Upper Permian to Lower-Middle Triassic sedimentary cover unconformably overlies the "Brescia Three Valleys Massif" crystalline basement of *Zone IV*. The metamorphic basement of Zone II and the Permian and Triassic sedimentary rocks of Zone III are cut by the eastern Adamello intrusion and the South Giudicarie Line (Cassinis, 1985; Brack *et al.*, 1985).

The Trompia Valley Line induced the rise and the south-verging thrust of the Zone IV crystalline basement on to the strongly deformed rocks of *Zone V*. In fact these pre-Norian Triassic formations are divided into numerous south-verging blocks by a local reverse fault system with northward-dipping planes. Both the Scythian-Anisian "Carniola di Bovegno" and the Carnian deposits aided the tectonic movements within this zone.

The Forno d'Ono overthrust is one of the most important tectonic planes that allowed the thrust of the Lower-Middle Triassic rocks on to the southern Norian "Dolomia Principale" formation of *Zone VI*. It consists of a thick and strongly rigid dolomitic plate, cut by reverse and wrench faults.

Also the Dolomia Principale generally thrust the younger Mesozoic formations of the neighbouring zone. Jurassic and Cretaceous rocks, locally unconformably overlain by Tertiary deposits, represent *Zone VII*. They reacted to the Alpine compressional movements by folding, which generated broad anticlines and synclines in the well-bedded Mesozoic limestones.

The final *Zone VIII* is characterised by the plain to the south of Brescia, where the whole stratigraphic succession described in the above-mentioned zones is buried under southward-thickening Neogene-Quaternary alluvial deposits.

PERMIAN TECTONO-SEDIMENTARY CYCLES AND PHYSIOGRAPHICAL SETTING OF THE REGION

The regional tectonic lines involved in the Alpine orogeny frequently show evidence of a long previous structural history, usually named "ancestral" character (Cassinis *et al.*, 1982; Castellarin, 1982; Castellarin & Vai, 1982; Doglioni & Bosellini, 1987; Cassinis & Castellarin, 1988; and so on). In many cases they had an active role of inversion during the Tertiary compressional regime. In fact, in former times many of these lines controlled the most intense phases of continental rifting: firstly during the Permian (Late-Hercynian event), then during the Norian and still later the Jurassic (Neo-Tethys rifting). During all these periods, the Brescia Province underwent extension.

The geological profile crosses two of the major Permian depositional basins in the region. These corresponded to continental intramontane basins with a WSW-ENE orientation, fault-bounded by basement structural highs. Only the activation of the aforementioned tectonic lines allowed the creation of these basins (Cassinis, 1982, 1985, 1988; Cassinis & Neri, 1990, 1992; Cassinis *et al.*, 1990, 1995, 1997; Perotti & Siletto, 1996).

South of the Tonale Line, the Orobic Basin is located between the Orobic Line-Gallinera Line, E-W trending system, and the southern Cedegolo Anticline. Further to the south, the sigmoidal shaped Collio Basin is bounded by the structural arc defined by the Trompia Valley Line to the south and the South Giudicarie Line to its eastern border.

Typology and areal variability of the Permian stratigraphic formations allow us to highlight the synsedimentary activity of these lines and their consequent cause and effect role in the birth and evolution of the Permian basins.

Permian formations (Cassinis, 1988) lie unconformably on the Hercynian (or Variscan) crystalline basement composed of phyllites, mica-schists and gneisses. Igneous and sedimentary continental deposits characterise the Permian succession in the Brescia Province (Cassinis, 1985; Ori *et al.*, 1988; Cassinis & Perotti, 1994). It can be divided into two major stratigraphical units, directly linked to two different tectono-sedimentary cycles (following the Hercynian orogeny) separated by a marked regional unconformity (Italian IGCP 203 Group, 1986; Cassinis *et al.*, 1988).

In the upper Trompia Valley, red-violet rhyolitic ignimbrites, lavas and tuffs of calc-alkaline geochemistry usually overlie the metamorphic basement. Only locally discontinuous outcrops of "Basal Conglomerate" (Upper Carboniferous?-Lower Permian) occur between these two units in the Brescia area. The basal volcanics appear as concordant with the overlying "Collio Formation" bodies (Peyronel-Pagliani, 1965; Origoni Giobbi *et al.*, 1979; Breitkreuz *et al.*, 1999).

Furthermore, the late Hercynian period is characterised by granodioritic and dioritic intrusives, such as the Navazze and Rango small bodies cropping out close to Collio (Upper Trompia Valley), which could be directly related to the Permian volcanic eruptions.

Continental alluvial-lacustrine well-bedded sandstones, siltstones and shales associated with interbedded volcanic rocks (Peyronel-Pagliani, 1965; Origoni Giobbi *et al.*, 1979; Breitkreuz *et al.*, 1999), characterise the "Collio Formation" (Cassinis, 1966a, 1966b; Cassinis *et al.*, 1975, 1978). This is a very thick sedimentary succession (up to 1000 m), infilling the subsiding Lower Permian basinal areas.

The reddish "Dosso dei Galli Conglomerate" (Cassinis, 1969a), rich in basement and Lower Permian volcanic fragments, lies on and laterally passes into the Collio Formation. It shows the alluvial conoid progradation from the borders to the depocentre of the basin. But the "Dosso dei Galli Conglomerate" laterally can also pass to the red-brown bioturbated fine sandstones and siltstones of the "Pietra Simona Member".

The rhyolitic-rhyodacitic ignimbrite unit of the red-violet "Auccia Volcanics" (Cassinis, 1969b; Peyronel-Pagliani & Clerici Risari, 1973; Origoni Giobbi *et al.*, 1979; Breitkreuz *et al.*, 1999) ends the Lower Permian tectono-sedimentary cycle, as well as the basin existence.

A time-gap of uncertain duration, most likely limited to the middle part of the Permian period, is associated with the regional unconformity that marks the contact between the two cycles (Cassinis & Doubinger, 1991, 1992; Cassinis & Perotti, 1997; Cassinis *et al.*, 2000 and in press).

The Upper Permian Cycle is defined by the red fluvial siliciclastic deposits ("red beds") of the "Verrucano Lombardo" and the disappearance of volcanic products. Between Camonica Valley and Giudicarie Valley, the "Verrucano Lombardo" is generally represented by fine-grained sandstones and siltstones.

The geological profile clearly shows rapid lateral facies and thickness changes in the Lower Cycle Units. In fact, the "Collio Formation" and associated volcanics disappear in the areas that were structural highs during the Early Permian. But, in the whole province, the "Verrucano Lombardo" is always present. It covers the Lower Group Units in the Permian basinal areas but directly overlies the crystalline basement on the Permian highs.

Consequently, both the cycles of the Permian succession occur where the profile crosses the Orobic and the Collio basins. In contrast, on the Camonica Valley High, which is located between the aforementioned basins, only the Verrucano lies on the metamorphic basement.

In a few kilometres, a step-by-step E-W trending fault system represents the tectonic boundary between the Collio Basin and the southern Trompia Valley High. In fact, the profile shows the Verrucano gradually overlying the basement southwards. So a thinner Permian succession crops out along an E-W oriented strip to the south of Trompia Valley Line. Moreover, the Permian granitoid intrusives within the upper Val Trompia basement are in accord with this structural pattern.

Another ridge, bounded by synsedimentary faults with a NNE-SSW orientation, earlier separated the Collio Basin from the western, smaller Boario Basin.

A large part of the above-mentioned volcanic products clearly came from the synsedimentary fault systems bounding the Permian basins. In fact, these tectonic scarp slopes were the source of strong magmatic activity following the Hercynian orogenic event.

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VOLCANISM AND ASSOCIATED SUB-LACUSTRINE CRYSTAL-RICH MASS-FLOW DEPOSITS IN THE EARLY PERMIAN COLLIO BASIN (ITALIAN ALPS)

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Key words – rhyodacite dome; phreatomagmatic explosion; turbidite; Permo-Carboniferous magmatism; post-Variscan.

Abstract – The Collio Basin is an intramontane continental basin developed as a consequence of post-Variscan orogenic collapse in southern Europe. Its evolution was controlled by extensional tectonics associated with calc-alkaline, intermediate and acidic magmatism.
Two episodes of ignimbrite-forming eruptions bracketed the development of the Collio Basin. During the intervening period, alluvial to lacustrine sedimentation was accompanied by episodic volcanism, recorded as fragments within sedimentary deposits, by syndepositional sills and dykes, massive acid laccoliths and domes with brecciated bases, and lastly, as volcaniclastic mass-flow deposits.
Immediately above the lower Collio Fm., several of these volcaniclastic mass-flows crop out in the central and eastern part of the basin; the most prominent are the Dasdana I Beds between the Trompia and the Caffaro valleys. These 10-20 m thick beds consist of: (1) amalgamated coarse-sandy to gravelly crystal-rich turbidites; and (2) a well-bedded sandy-pelitic unit rich in whitish lava fragments.
The volcaniclastic mass flows originated at the eastern margin of the basin from a porphyritic acid dome, presumably fragmented by phreatomagmatic explosions and/or seismic liquefaction. The metamorphic basement clastics and the high proportion of lacustrine pelitic fragments in the lower subunit indicate a (partially) intrusive position for the dome.
The sub-lacustrine/subaerial eruption column generated dense, crystal-rich turbidity currents.
Later, in a second phase, the foamy lava fragments sedimented from dilute turbidity currents, together with sandy-pelitic detritus, and from fall-out.

Parole chiave – duomo riodacitico; esplosione freatomagmatica; torbiditi; magmatismo permo-carbonifero; post-varisico.

Riassunto – Il Bacino di Collio in Val Trompia è un bacino continentale intramontano creatosi a seguito del collasso post-orogenico varisico nell'Europa meridionale. La sua evoluzione è controllata da una tettonica estensionale associata ad un magmatismo calc-alcalino intermedio e acido. Lo sviluppo del Bacino di Collio è scandito da due eventi ignimbritici basale e sommitale, ed è caratterizzato da una sedimentazione alluvio-lacustre alternata ai prodotti di un' attività vulcanica, che include clasti di questa natura all'interno di depositi sedimentari, *sill* e filoni sin-deposizionali, laccoliti e duomi acidi con basi brecciate, e torbiditi vulcanoclastiche. Quest'ultimi depositi si stagliano in un certo numero al di sopra della porzione inferiore della Formazione di Collio, che è di origine alluvio-lacustre, nella zona centro-orientale del bacino; il più significativo tra essi è quello affiorante sul M.te Dasdana ("Dasdana I Beds"), che dall'alta Val Trompia giunge fino in Val Caffaro. I "Dasdana I Beds" presentano in genere spessore tra 10 e 20 metri e sono formati da (1) una sotto-unità inferiore di torbiditi ricche in cristalli di taglia da grossolana a media, e (2) da una sotto-unità superiore arenaceo-pelitica ben stratificata, ricca in frammenti lavici di colore biancastro. Le torbiditi vulcanoclastiche si sono originate al margine est del bacino a seguito dell'esplosione freato-magmatica di un duomo acido di lava porfirica, unita possibilmente ad un fenomeno di liquefazione di origine sismica. I clasti di basamento metamorfico e l'alta percentuale di frammenti neri pelitici nella prima sotto-unità suggeriscono una posizione parzialmente intrusiva del duomo. La colonna di eruzione sub-lacustre/sub-aerea diede origine a correnti di torbidità dense (sotto-unità inferiore). Successivamente, i frammenti di pomici e di lave, anche a tessitura bollosa, furono sedimentati all'interno di correnti di torbidità più diluite, o si depositarono a seguito della precipitazione del "*fall-out*" (sotto-unità superiore).

INTRODUCTION

The Collio Basin in the Brescian Alps is one of the best-preserved intra-Variscan post-orogenic basins of southern Europe. The evolution of the Collio Basin occurred between 283±1 and 281±2 Ma (Schaltegger & Brack, 1999), *i.e.* during the Early Permian, and its dynamics included tectonism, sedimentation and volcanism (Cassinis, 1966;

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Ori *et al.*, 1988; Cassinis & Perotti, 1997). The Collio Basin presumably formed as a half-graben pull-apart structure within the framework of Late Paleozoic dextral transcurrent tectonics (Fig. 1).

The succession of volcanic events within the Collio Basin can be correlated with many intramontane troughs in southern Europe, such as those in Sardinia (Perdasdefogu, Escalaplano and Seui) and in the Ligurian Briançonnais. The timing of volcanism in relation to regional transtensional tectonics points to a Variscan post-collisional setting (Cortesogno *et al.*, 1998). The volcano-sedimentary deposition in the Collio Basin was bracketed in time by two episodes of ignimbrite-forming eruptions. The present contribution summarises recent research carried out in the (sub)volcanic units and major volcano-sedimentary mass-flow deposits intercalated within the alluvial-to-lacustrine sediments of the Collio Formation.

VOLCANIC ACTIVITY AND DEPOSITIONAL DEVELOPMENT IN THE COLLIO BASIN

The earliest volcanic activity is represented by subaerial emplacement of rhyolite ignimbrites directly on the Variscan basement, which in places was affected by pedogenic alteration. Volcaniclastic deposits are locally separated by decimetric tuffs with accretionary lapilli, indicating intermittent phreatomagmatic activity, and sandy to gravelly alluvial sediments. Grey to reddish-grey ignimbrite layers are homogeneous in composition, rich in pumice clasts and poorly welded; they extend far to the east, with a progressive increase in thickness (up to100 and more metres) westwards of the basin. This suggests a relatively flat paleotopography and probably a distal deposition from an extrabasinal source zone. The occurrence of andesite lava clasts within the ignimbrite indicates an earlier, intermediate volcanic activity.

Deposition of conglomeratic and arenitic alluvial fans intermittently followed, indicating progressive subsidence of the basin, triggering erosion along the margins. The common occurrence of rounded andesite clasts in the conglomerates suggests that the volcanic centres were localised at and near the eroded border of the Collio Basin. The subsequent depositional phase is represented by alluvial plain to lacustrine finer-grained clastics, and probably indicates a period of relative tectonic stability.

Rhyolite and rhyodacite laccoliths and domes were essentially emplaced along the southeastern margin of the basin (Malga Fontana, Dosso del Bue), probably related to basin-controlling fault activity.

Repeated phreatomagmatic explosions are indicated by: (i) breccias with rhyodacite clasts up to one metre in size underlying the dome structure; (ii) dome portions with subintrusive microtextures directly covered by tuffs, vesicular hyaloclastites and sediments; (iii) occurrence of thick tuffs with accretionary lapilli that probably represent surge deposits. Explosion of some porphyritic lava domes led to the formation of widespread sub-lacustrine mass-flow deposits (see below).

Towards the western development of the basin, distal deposits of the volcanic activity comprise green-blackish, highly fine silicified deposits (cinerites ?), characterised by laminar and convolute structures, locally interbedded with pelites rich in plants. Also, in places, late intermediate magmatic activity is recorded in the form of rare dykes.

The peak of the volcanic activity was followed by the second sedimentary cycle (upper Collio Fm.) characterised by prevailing arenitic layers, with frequent reworked, intermediate-to-acid volcanic clasts. Sedimentation within the basin ended with the deposition of con-

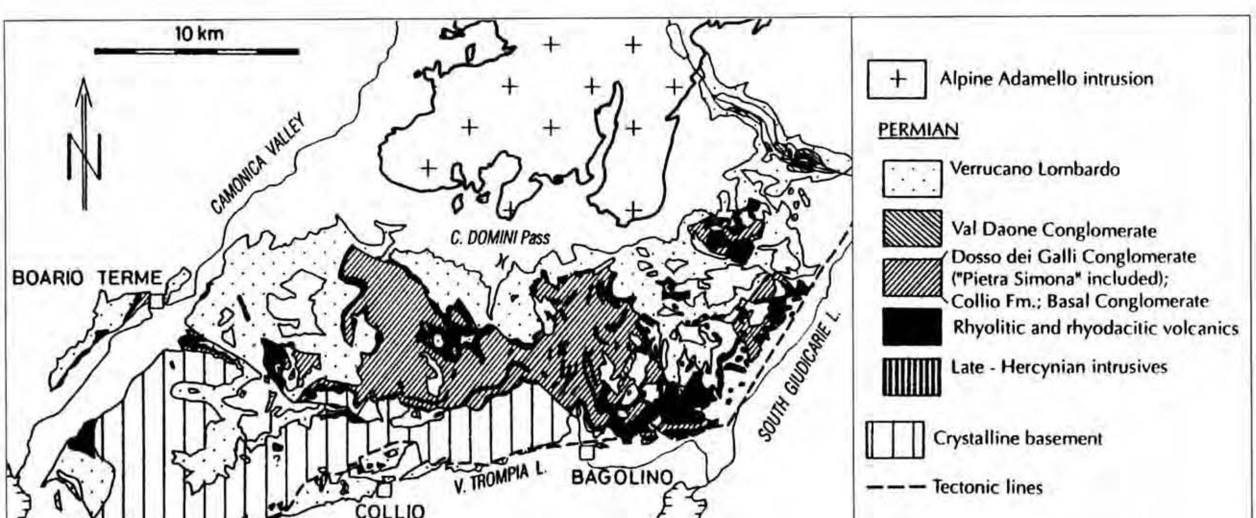

Fig. 1 – Location of the Early Permian continental Collio Basin between the Camonica and South Giudicarie Valleys, central Southern Alps.

glomerates rich in a hematite matrix (Dosso dei Galli Fm.).

Extended and thick (>100 m) subaerial ignimbrites (Auccia ignimbrite), characterised by a prevailing violet colour, abundant phenoclasts and eutaxitic *fiammae* and a strongly welded glassy matrix, covered the whole basin. Locally, a thin hydroxide-rich paleosol separates the top of the ignimbrite from the overlying Verrucano Lombardo fluvial clastics.

CHEMICAL FEATURES OF (SUB)VOLCANIC ROCKS AND OF THE DASDANA VOLCANICLASTIC MASS-FLOW

Chemical data on the magmatic rocks of the Collio Basin can be found in Peyronel-Pagliani (1965), Peyronel-Pagliani & Fagnani (1965), Peyronel-Pagliani & Clerici Risari (1973), Cassinis *et al.* (1975), Origoni *et al.* (1979), and Cortesogno *et al.* (1998).

The present data are addressed in order to compare the volcaniclastic mass-flows with the massive acid volcanic rocks in the basin. Samples of the volcaniclastic mass-flows were selected in order to avoid any sedimentary or basement components. The subvolcanic bodies occurring eastwards of the basin are dacitic to rhyolitic in composition. The analysed lithologies (Fig. 2A) included: (1) high-K dacite (possibly a complex laccolith) intruding pyroclastic deposits (Malga Fontana); (2) high-K dacite clasts in an explosive magmatic breccia covered by a rhyolite flow (Malga Scaie, east of the Dorizzo valley); and (3) five samples from rhyolite domes (Dosso dei Lupi, Rio Secco). The Malga Fontana intrusion and the Malga Scaie breccia are interpreted as dacite on the basis of mineral modes in the phenocryst assemblage, although they fall in the tra-

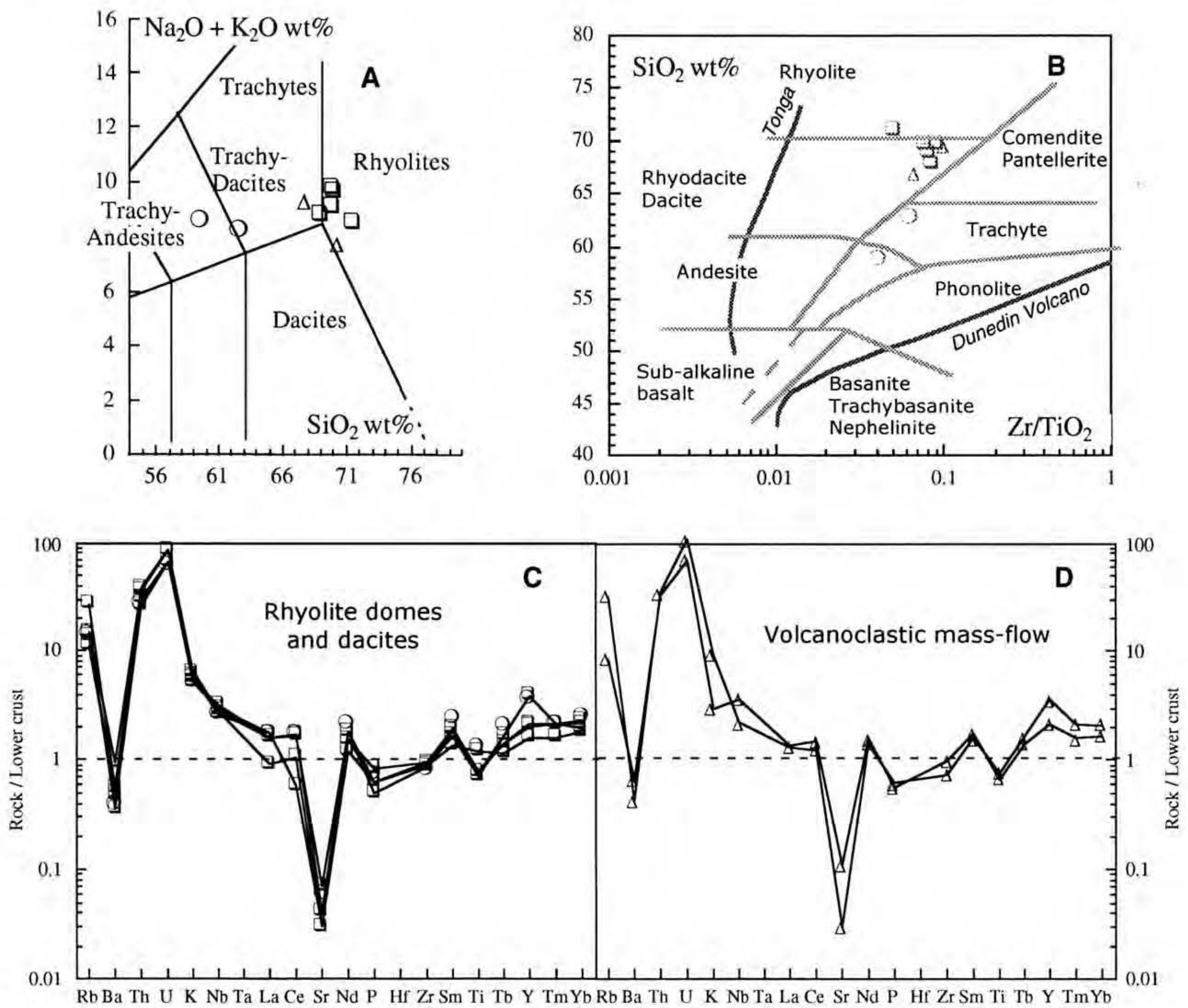

Fig. 2 – Chemical features of the volcaniclastic mass-flows in comparison with the dacites and rhyolites. A) Total alkali–silica classification (Le Maitre *et al.*, 1989) for dacites (○) rhyolites (□), and the Dasdana I Beds (△). B) $Zr/TiO_2 - SiO_2$ (Winchester & Floyd, 1977). Symbols as in A). C) Rock/lower crust normalised spidergrams (Weaver & Tarney, 1984) for dacites (○) rhyolites (□) and the Dasdana I Beds (△).

chyandesite-trachydacite field in Fig. 2A. They also differ from the probably younger dacite domes in having higher modal biotite. The marked differences between dacite samples, mostly in SiO_2 and Na_2O/K_2O ratio, are controlled by high porphyricity (20 < P.I. < 35) and by the irregular distribution of phenocrysts.

All lithologies show relatively high alkali contents ($Na_2O + K_2O$ in the range of 8-10 wt%). Affinity with the high-K transitional series is also consistent with the Zr/TiO_2 ratios (Fig. 2B). The relatively high CaO content in one of the mass flows (1.56 wt%) is related to the presence of carbonates in the matrix. Also, REE patterns (not reported) are homogeneous, with significant fractionation of LREEs, almost flat HREEs, and a weak negative Eu anomaly.

Lower-crust normalised spidergrams (Weaver & Tarney, 1984) for the dacites, rhyolites and volcaniclastic mass-flows (Fig. 2 C, D) are highly homogeneous, showing evident positive anomalies for LILEs except for Ba_N, being close to unity or slightly lower, and a marked negative anomaly for Sr_N. HFSEs show normalised values near to unity, with weak negative anomalies for Sm, Tm, Y and Yb. The volcaniclastic mass-flows strongly resemble the rhyolitic domes for major, trace and rare earth elements, pointing to a common origin. The slightly lower average silica and alkali contents could correspond to secondary enrichment in early crystallised phenocrysts, as a consequence of loss of floating glassy material.

THE SUB-LACUSTRINE VOLCANICLASTIC MASS-FLOW DEPOSITS: OCCURRENCE AND SEDIMENTOLOGY

The alluvial to lacustrine sequences of the Collio Fm. contain repeated intercalations of volcaniclastic mass-flow deposits, some of which cover a large part of the homonymous basin (Cassinis, 1988; Cassinis & Perotti, 1997). The slopes of the Val Caffaro, in the east of the Collio Basin, expose at least three of these crystal-rich competent units.

For its massive look and thickness, one of the most prominent is represented by the Dasdana I Beds, cropping out in the upper part of the lower Collio Fm. from Val Caffaro to M. Colombine (Fig. 3). The 10-20 m thick Dasdana I Beds consist of two subunits: (1) amalgamated coarse-sandy to gravelly crystal-rich turbidites; and (2) a well-bedded sandy-pelitic unit rich in whitish, frequently large, lava fragments (Plate 1, A-C). Both these units correspond to the old informal members "D" and "E" , respectively, which were introduced by Cassinis in 1966.

Detailed sedimentological sections, a representative selection of which is shown in Fig. 4, document the geometry and granulometric features of the Dasdana I Beds: the thickness of the lower subunit (1) increases from Mt. Colombine in the west to Val Caffaro in the east. The thickness of subunit (2) is almost constant; conversely, apparent changes occur in clast nature and in the clast/matrix ratio (see below). At the base of the Dasdana I Beds, soft sediment deformation results in decimetre to metre-scale load casts and balls of volcaniclastic material sunk into the substrate. In places, roughly E-W oriented channels occur at the base.

At Mt. Dasdana, mineralised danburite, tourmaline, ankerite, and trace gold-bearing layers in alternating millimetre to centimetre-scale laminae have been discovered in pelites a few metres just below the Dasdana I Beds. Generally, borate and boron silicates can form in evaporitic environments, and precipitation of less soluble boron

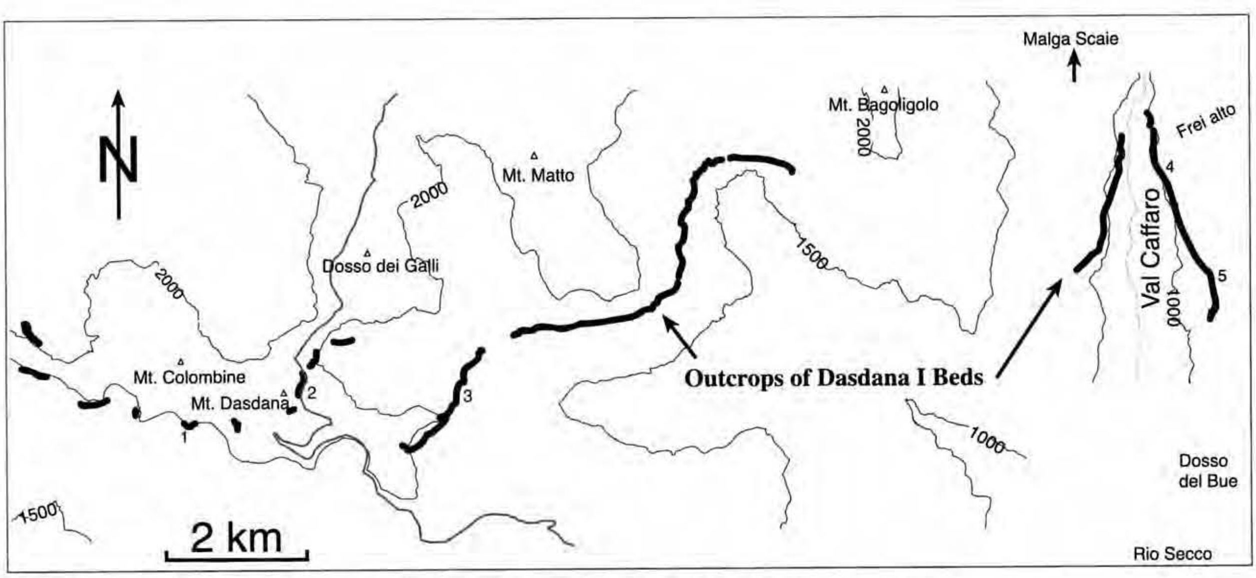

Fig. 3 – Outcrops of the Dasdana I Beds in the Collio Basin. Numbers indicate the sites of measured sections.

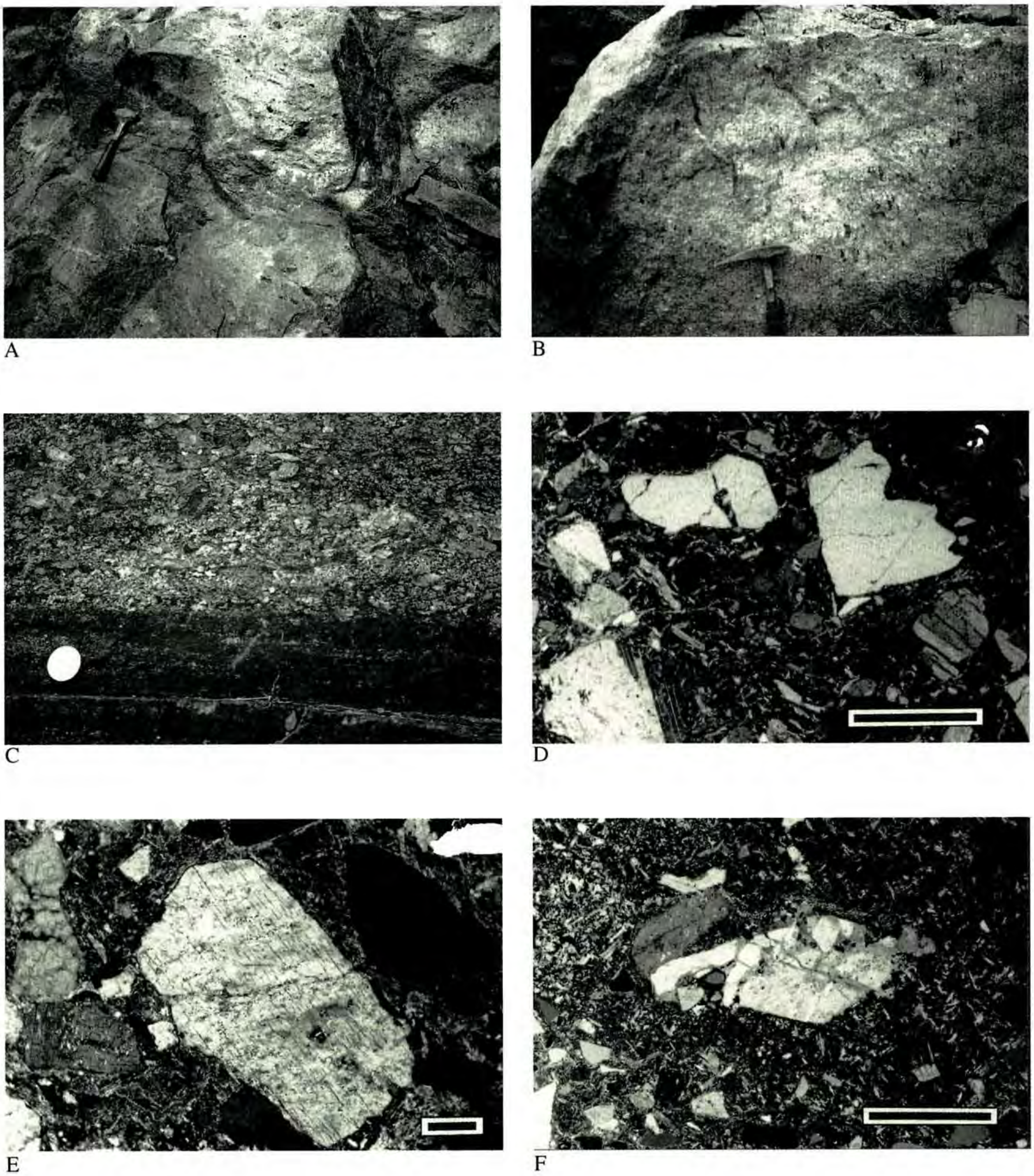

Plate 1

A. Field features of the Dasdana I Beds: erosional contact between the lower and upper subunits.

B. Lower subunit of the Dasdana I Beds including centimetre-scale clasts of prevailing andesitic and subordinate acidic volcanic rocks and metamorphic basement.

C. Upper subunit of the Dasdana I Beds: layer rich in porphyritic lava clasts.

D. Dasdana I Beds: microphotograph from the upper subunit, showing phenoclasts of volcanic quartz and plagioclase. Crossed polars, scale bar = 1 mm.

E. Dasdana I Beds: microphotograph from the upper subunit, showing a volcanic K-feldspar. The inner deformation of the phenoclast precedes its inclusion in the fine-grained matrix. Crossed polars, scale bar = 1 mm.

F. Dasdana I Beds: microphotograph from the upper subunit, showing a lava clast of felsitic dacite and a K-feldspar phenoclast in a matrix of neoblastic quartz. Crossed polars, scale bar = 1 mm.

silicates (danburite and tourmaline) instead of borate is favoured by high alkalinity. In most cases, occurrences of danburite are associated with volcanism, and the related hydrothermal activity provides the source for boron (Harder, 1959). Traces of danburite and tourmaline in the sediments exclude any post-diagenetic origin.

Subunit (1) of the Dasdana I Beds comprises light grey, amalgamated gravelly Bouma A (B) divisions. Faintly stratified Bouma B sequences occur in the form of discontinuous erosional remnants. Some of the 1-2 m thick Bouma A divisions start with darker grey matrix-rich deposits and have distinct bases showing erosional W-E oriented channels (Plate 1A). Black pelite (lacustrine Collio rip-ups) and various volcanic and metamorphic rocks occur as outsized clasts (up to metre-sized), isolated or concentrated in lenses. The modal composition of the lower subunit comprises fragments of plagioclase (20-30%), K-feldspar (7%), quartz (20%), biotite (3-10%), and por-

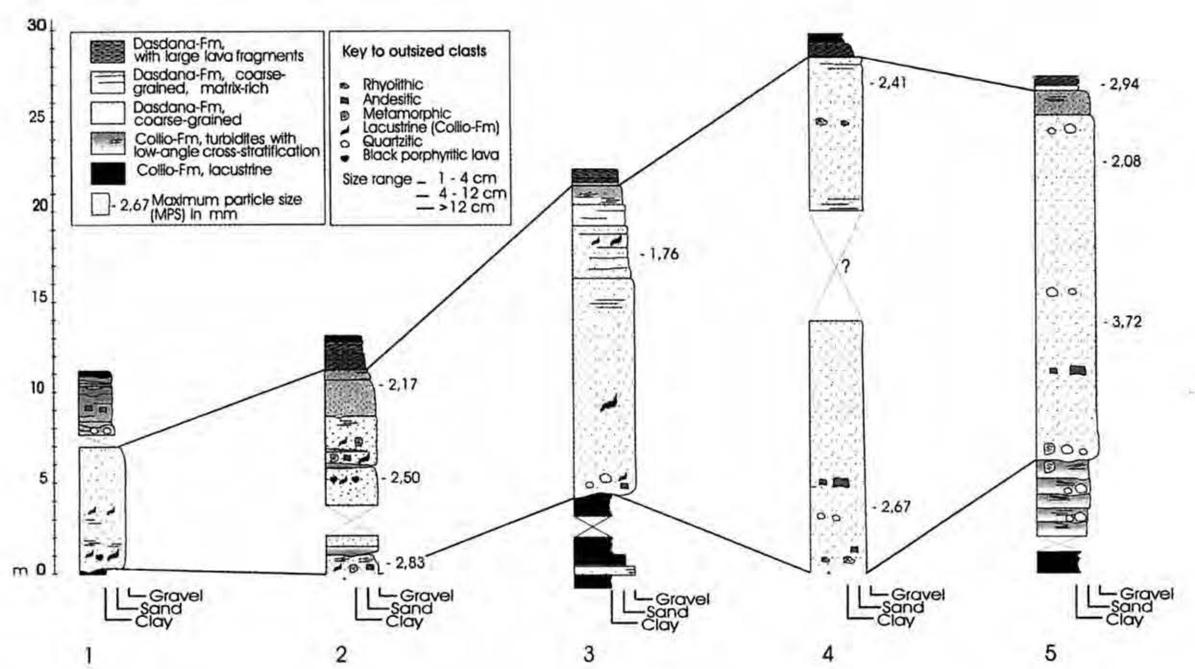

Fig. 4 – Representative sedimentological sections for the Dasdana I Beds.

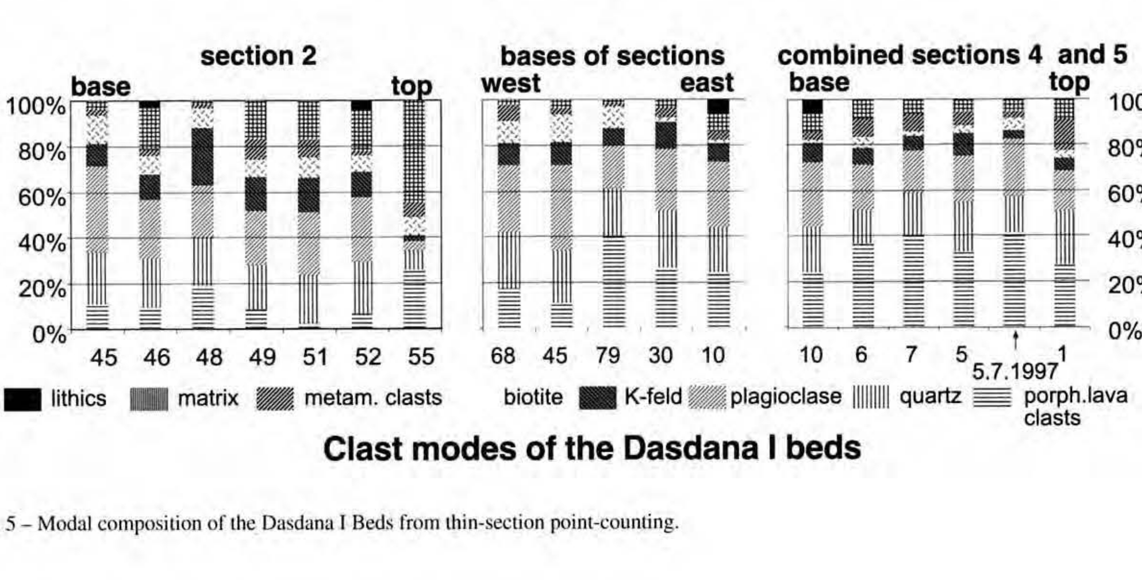

Fig. 5 – Modal composition of the Dasdana I Beds from thin-section point-counting.

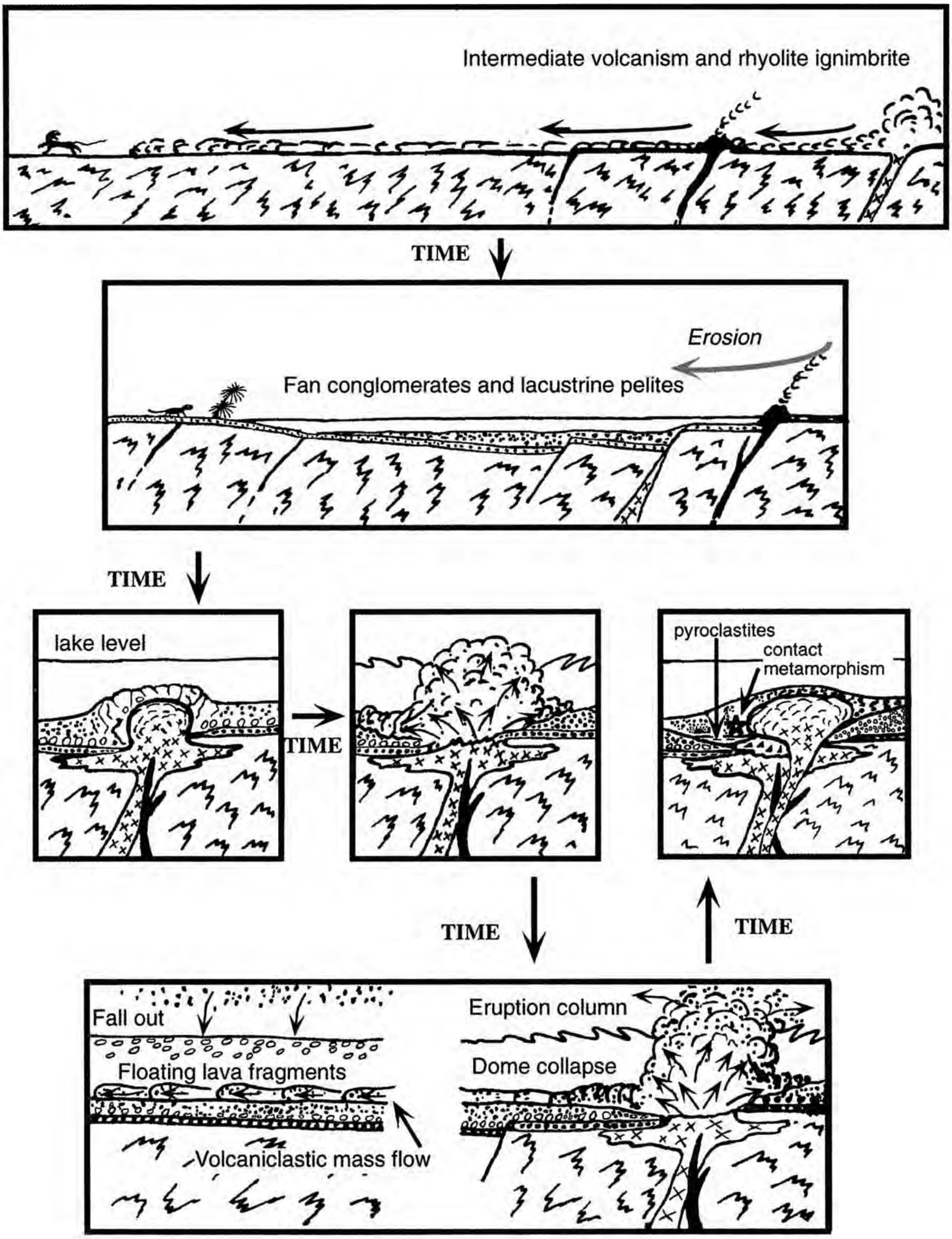

Fig. 6 – Highly idealised tectonic, sedimentary and magmatic evolution for the Collio Basin, inferred from local relationships.

phyritic rhyodacitic/rhyolitic lava (20-30%) as well as metamorphic basement clasts (5%, biotite, muscovite, garnet-bearing quartz mica-schist) and pelite clasts (Fig. 5). The lower Dasdana I Beds revealed some systematic proximal-distal trends such as a higher plagioclase and biotite content together with a diminishing unit thickness, maximum particle size and proportion of porhyritic lava fragments towards the W (Fig. 4).

The rhyodacitic-rhyolitic porphyritic lava clasts (Fig. 5) of the Dasdana I Beds resemble a single population displaying the same phenocryst assemblage (quartz, plagioclase, K-feldspar, biotite; Plate 1D, F) that is present as crystal fragments (in the mass-flow deposits). Predominantly the groundmass of the lava clasts consists of a micrographic mosaic of quartz and feldspar; subordinately, illite-chlorite groundmass occurs, produced by *in situ* alteration of glass. Uncompacted lava clasts with micrographic groundmass show irregular ragged to cauliflower shapes typical of the phreatomagmatic fragmentation of viscous magma. From these textures we infer that the lava fragments originated from a lava dome that was already cooling down, forming an outer glassy carapace and an inner crystallised core. Computer-aided image analyses indicate that the phenocryst content of the fragmented dome was of the order of 20%.

The upper subunit (2) of the Dasdana I Beds consists of well-bedded, partly turbiditic, sandy to pelitic deposits (Figs 4, 5), which contain varying amounts of gravel-sized, porphyritic, acidic lava fragments. The phenocryst assemblage within the lava fragments is similar to that present in the lower subunit; however, spindle to cauliflower shapes are dominant (Plate 1C). Where illite and/or chlorite replaced the groundmass of the fragments, strong compaction took place. In addition to clay minerals, quartz, albite and carbonate formed. These strongly compacted lava fragments presumably had a glassy groundmass during transport and deposition.

POSSIBLE SCENARIO FOR THE FORMATION OF THE LACUSTRINE VOLCANICLASTIC MASS-FLOW DEPOSITS

Striking features of the Dasdana I Beds and similar units of the Collio Fm. include the compositional homogeneity within a given unit, and the strong concentration of crystals in the thick, gravelly lower subunit. Comparing the content of crystals originating from the lava dome (in places more than 60%, Fig. 5) with the estimated crystallinity of the lava fragments (about 20%), strong fractionation between lava and crystal fragments during eruption and transport has to be considered. The groundmass textures found in the porphyritic lava fragments (from granophyric to (glomero)porphyritic, with felsitic poikilomosaic and spherulitic up to glassy mesostasis) indicate that the source for the mass flows was a texturally zoned lava dome, and not an erupting magma chamber. Thus we propose that the Dasdana I Beds and similar sub-lacustrine deposits of the Collio Basin formed as a consequence of phreatomagmatic explosions and/or seismic liquefaction of rhyodacite domes, which had already developed a textural zonation upon cooling. The common occurrence of fine to large fragments of metamorphic basement and of lacustrine pelite clasts suggests, at least for the Dasdana I Beds, an intrusive position for the cryptodome, presumably at the transition between the Variscan basement and the overlying Collio Fm. As a consequence of the liquefaction of the cryptodome, a sub-lacustrine/subaerial eruption column formed, from which pulses of dense, crystal-rich mass-flows originated, leading to the deposition of the Bouma A (B) units. In the first instance, much of the foamy lava remained in the column and sedimented later in a second phase, from dilute turbidity currents together with sandy-pelitic detritus, and from fall-out, forming the upper sandy-pelitic subunit rich in large lava fragments.

CONCLUSIONS

The subsiding half-graben of the Collio Basin was controlled by approximately N-S and E-W oriented faults that represented the main conduits for the ascent of magmas at its eastern marginal areas (Fig. 6). The magmas produced minor andesite effusions and rhyodacite laccoliths and domes, mostly emplaced at the basement-cover contact. Upon emplacement some of these lava domes and laccoliths were affected by fluidisation, perhaps triggered by earthquakes, which led to strong phreatomagmatic eruptions and to the formation of crystal-rich volcaniclastic mass-flow deposits. Similar crystal-rich volcaniclastic deposits have been reported from other Permo-Carboniferous basins in Europe, such as in the Pyrenees (Marti, 1996).

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NEW PALAEONTOLOGICAL DATA FOR THE VAL GARDENA SANDSTONE

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Key words – ichnology; northern Italy; Permian; tetrapod footprints; biochronology.

Abstract – A new find of tetrapod footprints – *Rhynchosauroides* sp. cfr. *R. palmatus* (Lull, 1942) – from the lowermost portion of the Val Gardena Sandstone allowed us to improve our knowledge of the stratigraphy of this formation.
The section, from which the new material comes, was studied at various times from both the sedimentological and the paleontological point of view. It was subdivided into five (and a lower part of a sixth) third-order depositional cycles. The very rich paleontological data (ichnofossils and sporomorphs) and some peculiar depositional characteristics (interfingering and overlying marine layers, a clearly exposed P/T boundary) made the section optimal for stratigraphical purposes. Unfortunately botanical and ichnological data allowed good results only for the upper cycles, while the data from the first cycle were not completely satisfactory.
The new find, from within 20 m of the underlying volcanics, ultimately solved that problem, allowing us to ascribe the whole sequence to a very short depositional time interval.

Parole chiave – icnologia; Italia settentrionale; Permiano; impronte di tetrapodi; biocromologia.

Riassunto – Viene segnalato il ritrovamento di una icnofauna con *Rhynchosauroides* sp. cfr. *R. palmatus* (Lull, 1942) nella parte inferiore delle Arenarie di Val Gardena, nella sezione del Bletterbach (Bolzano). Il nuovo ritrovamento ha permesso di chiarire la stratigrafia di questa formazione.
La sezione da cui proviene il nuovo materiale è stata studiata a più riprese sia dal punto di vista sedimentologico che paleontologico. I dati paleontologici (icnofossili e sporomorfi) e le caratteristiche deposizionali (livelli marini intercalati e a copertura, ottimo affioramento del limite P/Tr) ne fanno una splendida sezione per la stratigrafia del Permiano superiore. La sezione comprende cinque cicli deposizionali di terzo ordine e la parte inferiore di un sesto ciclo; sfortunatamente i dati icnologici e paleobotanici davano buoni risultati soltanto per i cicli superiori mentre non erano soddisfacenti per il I ciclo. La nuova faunula, proveniente da un livello posto a meno di venti metri dalle vulcaniti del substrato e proprio da sedimenti del I ciclo, permette di risolvere il problema e di attribuire l'intera sequenza ad un breve intervallo di tempo.

GEOLOGICAL SETTING

A new find of tetrapod footprints from the lowermost portion of the Val Gardena Sandstone has allowed us to improve the biochronological calibration of that formation and to eliminate a large degree of uncertainty from previously known data on stratigraphy of Permian deposits in the alpine region. The Permian rocks cropping out in the central and eastern Alps are traditionally subdivided into two major tectonostratigraphic complexes or cycles (Italian IGCP 206 Group, 1986), the first (pre-Permian-Lower Permian) composed of a complex of volcanic, sedimentary and volcano-sedimentary rocks, the second (Upper Permian) of clastics and calcareous rocks. The two complexes are separated by a regional unconformity.

The Upper Permian cycle is in its turn mainly composed of siliciclastic sediments deposited in environments related to braided river plains and deltas, and by carbonate limestone laid down in a clearly marine environment. The former are ascribed to the Val Gardena Sandstone (VGS), and the latter the Bellerophon Fm. The upper boundary of the Bellerophon Fm. Finds is the sharp contact with the Tesero Mb., the first member of the overlying Lower Triassic Werfen Fm.; at that contact is traditionally located the P/T boundary (Assereto *et al.*, 1973).

As a whole the Permian deposits form a transgressive sequence. They were subdivided into minor cycles: originally into three third-order sedimentary cycles (Massari *et al.*, 1988), and then subsequently into five and a lower part of a sixth (Massari *et al.*, 1994). Moreover, the VGS was dated using scattered marine elements in its upper part (Neri *et al.*, 1994), and by means of sporomorph assem-

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blages and tetrapod footprints (Conti *et al.*, 1977; Ceoloni *et al.*, 1988; Massari *et al.*, 1988, 1994). The presence of a magnetic reversal recognised as the Illawarra Reversal Event (Mauritsch & Becke, 1983) constrained the lower boundary.

The outcrop from which the new find comes is the most famous Permian track-site of northern Italy; the section is located along the Bletterbach near Redagno (Aldein, Bozen, Italy) (Fig. 1). It was studied at various times both from the sedimentological (Massari *et al.*, 1988, 1994) and the palaeontological point of view (for a historical review, see Blieck *et al.*, 1995, 1997). Subsequently, footprints from the same outcrop have been used as a database to establish a faunal unit and the corresponding faunal age (Conti *et al.*, 1997), obviously named after the section as the Bletterbach Faunal Unit and the Bletterbach Faunal Age. The chronostratigraphic boundaries of the faunal unit were constrained, by the assumed Illawarra Reversal Event at the base, and by the Fungi Blooming Event at the top, to an interval ranging from the Midian (*p.p.*) to Late Djulfian (Conti *et al.*, 1997).

The very rich paleontological data (ichnofossils and sporomorphs) and some peculiar depositional characteristics (interfingering and overlying marine layers, and the clearly exposed P/T boundary) make the section optimal for the stratigraphical purposes. Only the paleontological data from the sediments pertaining to the first third-order cycle were not completely satisfactory. Footprints were lacking from the first 60 m of the section, and sporomorphs were too poorly preserved within the same interval. The new find, from within 20 m of meters the underlying volcanics, has solved that problem, allowing us to progress with the stratigraphy.

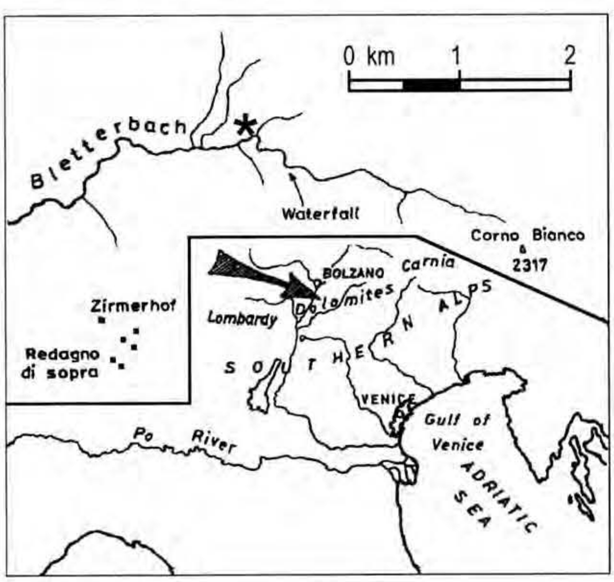

Fig. 1 – Location of the Bletterbach section; the asterisk shows the exact location of the new find.

TETRAPOD FOOTPRINTS

The VGS and the Bellerophon Fm. were subdivided into cycles, and the same cycles were recognised along other sections over the whole southern Alpine region. All the examined sections show the same vertical organisation (Massari *et al.*, 1988, 1994). In the Bletterbach Gorge the upper boundary of the first cycle was set at around 74 m from the base in Massari *et al.* (1988), and at 29 m in Massari *et al.* (1994). Up to that time, no footprints had been found within the first 60 m of the succession. The new footprints were found along a lateral channel of the Bletterbach, imprinted on a small surface 13 m above the lower boundary between sandstone and the underlying volcanics (Fig. 2).

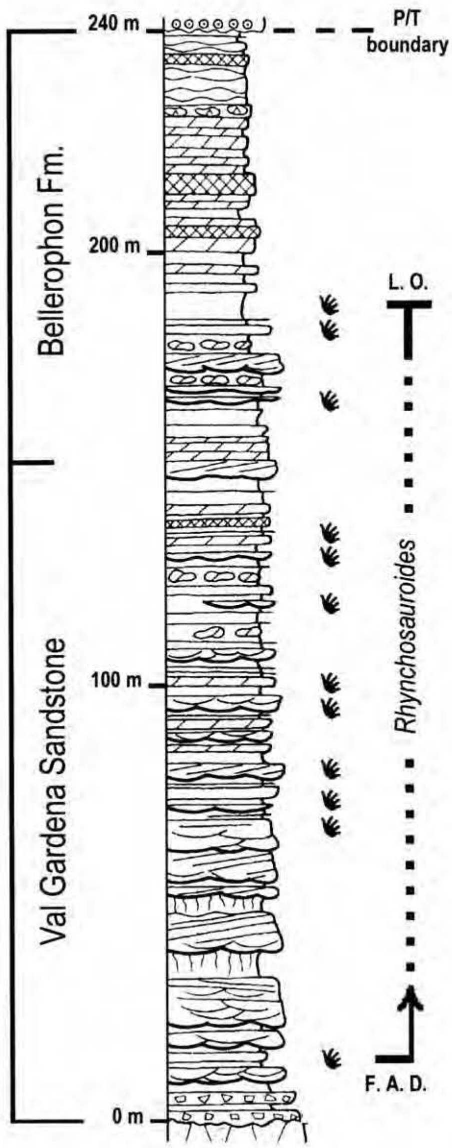

Fig. 2 – Simplified stratigraphic column of the section; the *Rhynchosauroides* F. A. D. arrow points to the new ichnological level.

The first part of the formation is composed of coarse to medium sandstone, reddish-purple in colour, poorly cemented, poorly sorted and rich in porphyric pebble. Deposits are upward-fining, passing to medium-grained arenites, structureless and poorly cemented. The imprinted surface, about 160 by 80 cm in size, was found within a better cemented medium-grained arenitic body, with some small subspherical cavitie, probably due to the dissolution of gypsum nodules. The rock unit containing the new find is less than one metre thick and extends laterally only for some tens of metres. Below and above there are numerous levels with gypsum nodules, and others with pedogenetic structures like calcrete, colour mottling and deep dessication cracks.

At 15 m from the base the situation changes abruptly, thanks to the presence of a cross-laminated, coarse-grained arenitic body with an erosional base. This is the first of the frequent channelised bodies that one can find going up-section.

According to Massari *et al.* (1988), such parts of the section "may represent the distal part of a semiarid alluvial fan, locally merging into an inland sabkha where high evaporation rates may have caused precipitation of sulphates in the capillary fringe above the water table". The same depositional environment is confirmed by Massari *et al.* (1994): "the deposits may record sedimentation on "flashy alluvial" fans typical of semi-arid areas".

Surface analysis

The poor condition of the exposure (the surface is periodically flooded and covered by large amounts of debris) and the risk of destruction of the specimen by weathering, compelled us to make a plastercast. The first time we moulded the natural casts, they were subsequently lost be-

Fig. 3 – Slab with some natural casts of footprints and ripple marks.

cause they were imprinted on a thin, uncemented clay cover. A few months later, with a more specific technique, it was possible to mould the corresponding underprints, still in place. In both cases we moulded the most footprint-rich portion of the surface, about 80 by 60 cm in size. Thus we have studied the original surface and some isolated natural casts (Fig. 3), a plastercast of the prints and another of the underprints.

The track-bearing surface is subdivided into two parts, the first flat and the second covered by ripple marks. The ripples are asymmetric and irregular, showing discontinuous and sinuous crests. The distance between crests ranges between 5 and 8 cm and their maximum height varies between 1,5 and 3 cm. Such bottom structures were probably made by unidirectional, low-intensity currents in very shallow waters.

The ripple crests are marked by some slight impressed and discontinuous groove marks, probably made by drifting plant fragments. Neither the footprint bearing bed, nor the few overlying layers show any internal structures.

Systematics

The material consists of nearly 30 footprints, ascribed to the ichnogenus *Rhynchosauroides* Maidwell, 1911. Some of them are preserved as natural casts, others as prints and underprints. Sometimes we have prints and reverse moulds of the same footprints. The material includes impressions of somewhat unclear trackways, as well as well-preserved isolated footprints and sets (Fig. 4). Because of the rich sample the analysis of extramorphologies has yielded clear results. As already mentioned, the footprints are quite irregular but, in some cases, extremely well-preserved. The greater irregularities seem to involve digit divergences. This extramorphology is more characteristic of lacertoid reptile footprints, and is frequently recognised when footprints were impressed on very plastic ground. It seems connected to be related to the greater equilibrium and stability of the track-maker.

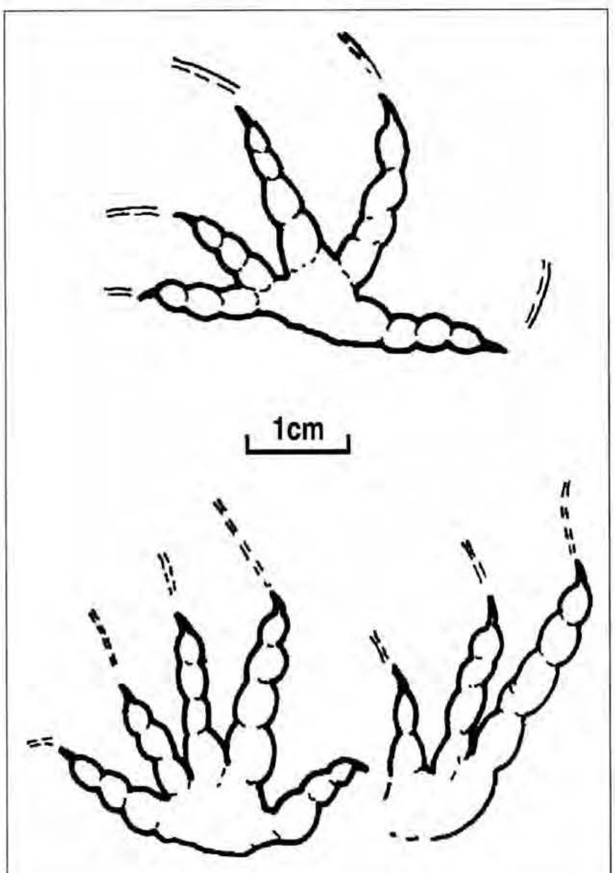

Fig. 4 – Drawing of a single manus and of a set of footprints from the new surface.

Ichnogenus *Rhynchosauroides* Maidwell, 1911
(type species: *R. palmatus* (Lull, 1942))

DESCRIPTION: quadruped lacertoid footprints, ectaxonic, asymmetric. Digit length increases from I to IV, V is strongly abducted and is as long as the I digit. Digits of *manus* are more or less curved inward, with the exception of digit V which forms an angle of about 60° with digit IV. The *pes* is slightly larger than the *manus* and shows a less rounded outline; *pes* digits are less curved. *Manus* often superimposing *pes* and sometimes surpassing it.

Even if showing functional prevalence especially on digits III and IV, it can be defined as semi-plantigrad showing the consistently clear proximal attachment of the metapodial-phalangeal pads and often the posterior edge of the sole.

OCCURRENCE: the genus is characteristic of the European Lower Triassic but was already used for some forms present in Permian VGS, from levels well above the one presently described (Conti *et al.*, 1977). The new finding can thus be considered as the first appearance datum level of the ichnogenus.

Rhynchosauroides cfr. *R. palmatus* (Lull, 1942)

MATERIAL: Seventeen isolated natural casts and two surface plaster casts, one with prints and another with underprints. On the plasters are preserved two trackways and some isolated footprints.

DEPOSITORY: footprints will be preserved in the Geological Section of the Museumsverein Aldein, at the school of Radein/Redagno (Bozen/Bolzano, South Tirol, Italy). Plaster casts of the surface are preserved at the same depository and at the Museum of Palaeontology of the Dipartimento di Scienze della Terra of the University "La Sapienza" of Rome.

DESCRIPTION: Trackway: Trackways are irregular due to environmental conditions and to the footprint-bearing bed geometry. In fact the surface shows a series of irregular ripples, and the footprints are imprinted both on the crests and within the troughs. The track-maker digits, quite vertically mobile, left footprints of similar shape while the dimensions, the relative distances and the digit divergences are very variable, depending on the ground characteristics. In the troughs, footprints are more deeply impressed, even if expulsion borders and displaced mud render the impression less clear; moreover *manus* and *pes* are often overprinted.

REMARKS: footprints closely similar to the new find were previously collected from some layers up-section. Apart from small differences in preservation, the new find shows the same size interval and the same evolutionary level as the previously discovered material and can be ascribed to the same ichnotaxon (compare specimen with collection number 75/10 in Conti *et al.* (1977, p.31).

Biochronological meaning

The new find, showing a very characteristic ichnospecies, also present in the upper portion of the sequence, enables us to ascribe the new footprints to the same faunal unit established for the higher layers (Bletterbach Faunal Unit in Conti *et al.*, 1997). It allows inclusion of the whole sediment thickness to a consistent biochronological unit (Bletterbach Faunal Age in Conti *et al.*, 1997).

CONCLUSIONS

The new finds can be ascribed to the very particular faunal association present in the upper levels of the section, on this basis we can ascribe the VGS cropping out in the Bletterbach section to a very short time interval. Such an interval can be coarsely calibrated on the basis of the rate of evolution of reptiles and the age of the last volcanics, at the base, and Late Permian marine fossils and sporomorphs at the top; it ranges between 260 and 251 Ma BP (Cassinis *et al.*, in press). This datum also brings forward to the age of the base of the VGS in the section, thus enlarging the time gap corresponding to the basal unconformity (Cassinis *et al.*, in press).

Previous studies on the same outcrop showed the presence of third-order sedimentary cycles. All but the first of them were ascribed to the same faunal unit, characterised by Late Permian ichnotaxa (Bletterbach Faunal Unit in Conti *et al.*, 1997). The lowermost cycle age remained a mystery, due to the lack of tetrapod footprints and to the presence of a sporomorph association with low stratigraphical meaning. Moreover, correlation among the third-order cycles shows the presence of the same first cycle in all the sections examined by Massari *et al.* (1994), and allows us to extend the conclusions valid for the Bletterbach section to the whole southern Alpine region. Consequently the age of the VGS can be limited to the same short time interval over the whole Alpine region. This raises strong doubts over the coincidence between the Illawarra Reversal Event and the magnetic reversal recognised by Mauritsch & Becke (1983) at the Paularo section in the Carnic Alps (Venturini, 1986).

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PERMIAN AND TRIASSIC TETRAPOD ICHNOFAUNAL UNITS OF NORTHERN ITALY: THEIR POTENTIAL CONTRIBUTION TO CONTINENTAL BIOCHRONOLOGY

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Key words – stratigraphy; biochronology; ichnology; reptiles; Permian; Triassic; Northern Italy.

Abstract – After theoretical analyses on the application of the main stratigraphic methods to continental deposits, we carried out a feasibility analysis to assess the ages of Permian and Triassic continental sediments by means of tetrapod footprints.
The data mainly originate from the Central and Southern Alps. The paleogeography of the Alpine region during Permian and Triassic times gave rise to a unique geological situation and well-exposed sections in which marine sediments, continental deposits rich in footprints and volcanic rocks are interfingered. The resulting mixed sections enable us to build a framework of biostratigraphical and chronological data in which tetrapod footprint-based evolutionary groups can be considered as biochronological units.
Such units (*Land Ichnofaunal Units*) and the corresponding biochronological units (*Land Ichnofaunal Ages*), still to be formalised, seem to reveal some advantages with respect to other dating systems.

Parole chiave – stratigrafia; biocronologia; icnologia; rettili; Permiano; Triassico, Italia Settentrionale.

Riassunto – L'applicabilità dei più diffusi e tradizionali metodi stratigrafici ai depositi continentali lascia molto dubbiosi. Di conseguenza è stata portata a termine l'analisi di fattibilità sulla possibilità di utilizzare le orme di tetrapodi per definire le età dei sedimenti continentali del Permiano e del Triassico. La base-dati utilizzata è stata raccolta dalle Alpi Centrali e Meridionali.
La paleogeografia del Permiano e del Triassico in tali regioni determinò una situazione geologica particolare e sezioni, ben esposte, nelle quali si intercalano sedimenti continentali a impronte di tetrapodi, vulcaniti e livelli marini. Le sezioni miste che risultano da questa situazione ci hanno permesso di costruire una maglia di dati cronologici e biostratigrafici nella quale insiemi evolutivi basati sulle orme di tetrapodi (*Land Ichnofaunal Units*) possono essere considerati come unità biocronologiche (*Land Ichnofaunal Ages*). Tali unità, ancora da formalizzare, sembrano rivelare parecchi vantaggi rispetto agli altri sistemi di scansione dei fenomeni geologici nei sedimenti continentali.

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INTRODUCTION

Over the last 30 years the authors have studied Permian and Triassic tetrapod footprints in the Southern and Central Alps (see bibliography). After the first, mainly systematic studies (Leonardi & Nicosia, 1973; Conti *et al.*, 1977, 1991, 2000; Ceoloni *et al.*, 1988b; Mietto, 1981, 1987; Santi, 1992 and in press; Avanzini & Neri, 1998; Leonardi, 2000; Avanzini, in press), some attempts were made to assess the age of track-bearing continental dep osits using different methods. Some of us (Conti *et al.*, 1979, 1989; Neri *et al.*, 1994) tried to assess the ages of continental deposits by studying them using classical biostratigraphical methods. Another attempt (Ceoloni *et al.*, 1988b) examined the possibility of correlation of our deposits with bone-based Bakker's "Dynasties" (Bakker, 1977) or with the "Empires" of Anderson & Cruikshank (1978). An attempt was also made, in the same paper, to use variation of the degree of biodiversity for the Permian interval (Ceoloni *et al.*, 1988b).

In a different approach tetrapod footprint-bearing levels were used as markers of transgressive or high-stand events (De Zanche *et al.*, 1993; Gianolla *et al.*, 1998) for Triassic sequence-stratigraphy in the Dolomites.

Each attempt, even if improving the subdivision and correlation of Permian and Triassic sediments, was quite unsatisfactory with respect to our efforts, and showed a variety of problems and uncertainties. At first we hypothesised that such negative aspects were linked either to the particular paleogeography/geology of the Alpine region, or to an incomplete ichnological literature. Subsequently we recognised the same problem in widespread geological domains (Lucas, 1996, 1998a, 1998b, 1999).

Indeed, the problems, large enough in marine stratigraphy and complicated by different codes and philosophies (Harland, 1992) seem to be enhanced during subdivision and correlation of continental deposits. Thus we began a feasibility study of a different stratigraphical tool, which is the use of tetrapod footprints for stratigraphical purposes using a biochronological approach. In this paper, after having analysed theoretical and practical problems, we will show the results of this analysis.

MAIN THEORETICAL PROBLEMS

The stratigraphical subdivision of continental deposits is characterised by enormous problems which depend on their intrinsic characteristics such as:

- the sharply changing depositional environments;
- the shape and dimensions of basins;
- the laterally and vertically discontinuous geometry of sedimentary bodies;
- the discontinuous fossil record: the frequent occurrence of non-fossiliferous units as well as units in which one does not find body fossils but just the less diagnostic footprints.

Problems also arise because of the kind of study, and because of real theoretical difficulties in applying classical stratigraphy. These last can depend on the systematics of land animals or, last but not least, can be due to the traditions of the discipline and to the conservative approach of many researchers to the argument. Moreover, most researchers work in this field after having been trained in marine sediment stratigraphy.

Problems in chronostratigraphy and geochronology

The lack of a specific standard reference scale for geological time is the first peculiarity that a "marine-trained" stratigrapher notices when attempting time subdivision and correlation in the continental deposits. Such a lack of standard reference units is not devoid of meaning, but corresponds to the actual situation with continental deposits.

Indeed, a conflict exists between the situations as recognised in continental deposits and the basic stratigraphic rules (mainly derived from the study of marine sediments). This conflict is well illustrated by sentences of this kind: - "*... Boundary-stratotypes of a stage should be within sequence of* ***continuous*** *deposition preferably marine ...*" (Hedberg, 1976, p. 71). - "*... Boundary-stratotypes should be chosen in sequences of essentially* ***continuous*** *deposition ...*" (Hedberg, 1976, p. 84). - "*... the correlation potential of any boundary level should be tested through a detailed study of several* ***continuous*** *successions covering the critical interval ... The most suitable of these sections can then be selected for definition of the GSSP.*" (Remane *et al.*, 1996). – "*... The boundary-stratotypes of a stage should be within sequences of essentially continuous deposition, preferably marine (except in cases such as the stages based on mammalian faunas in regions of nonmarine Tertiary sequences or the Quaternary glacial stages.*" and next "*... If major events in the geological development of the Earth can be identified at specific points in sequences of continuous deposition, these may constitute desirable points for the boundary-stratotypes of stages.*" (Salvador, 1994).

It is not by chance that the recommendation to use continuous sequences (*e.g.* marine pelagic) to establish unit-stratotypes or boundary-stratotypes (GSSPs) is present within the stratigraphic codes (Hedberg, 1976; Salvador, 1994) and within IUGS Guidelines (Remane *et al.*, 1996). We believe that "continuous" in that sense means not only and simply "without gaps" but mostly "constant" in every sense. This need is based on the theoretical constraint of the maximum continuity in type and rate of sedimentation and in the constancy of the deposition environment, "*... Absence of vertical facies changes at or near the boundary. A change of* *litho- or biofacies reflects a change of ecologic conditions which may have controlled the appearance of a given species at the boundary level. ...*" (Remane *et al.*, 1996). Continuity and constancy are needed for full confidence in the recognition of both faunal changes and all other events.

These characteristics contrast sharply with the evident characteristics of continental deposits. Indeed, the continental environment is known for discontinuities and sharp variations in the rate and type of sedimentation. Moreover basins are laterally discontinuous and the geometry of rock bodies is irregular; thus direct correlation, superposition and continuity are almost always impossible in practice and incorrect from a conceptual point of view.

From the above we recognised the theoretical impossibility of establishing "continental stages" and the use of special chronostratigraphical units. Also the selection of continental sections as assumed auxiliary type-sections seems unworkable. It is well known that the only geochronological units that we can use are based on standard marine stages, but also that correlational elements between continental and marine sequences are usually few or completely absent. Thus chronostratigraphy and geochronology are both extremely difficult to use for assessing the age of continental sediments.

Problems in lithostratigraphy

Problems also arise if one limits oneself to the use of lithostratigraphy; strong similarities in composition of sediments and the repetition of their depositional environments show the substantial inadequacy of the lithostratigraphic method, as demonstrated by the frequent diachronic correlation present in the old literature. One exemplar in this field is the widespread use of the names Rotliegendes or Buntsandstein (both actually lithostratigraphic units at a group level; Menning, 2000a) with chronostratigraphical meaning. The same is true for the term "Weald", originating in England but used the world over. A recent example of the low degree of confidence of the lithostratigraphic criteria is the systematic attribution of Cretaceous rocks to Palaeozoic units and ages in the several intracratonic basins of NE Brasil, because of misleading lithostratigraphic similarity, until the recent systematic discovery of dinosaur footprints within the units (Leonardi & Avanzini, 1994, p. 54; Carvalho *et al.*, 1994).

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The institution of the unconformity-bounded stratigraphic units (*e.g.* allostratigraphic units, synthems – see Salvador, 1987; 1994) although partially solving the above-mentioned problem, within an intrabasinal lithostratigraphical framework and for sequence-stratigraphy, seems useless for continental chronostratigraphy and geochronology.

"*...The worst possible boundary (for chronological units) is an unconformity ...*" this sentence (Hedberg, 1976, p. 84) finds its theoretical basis in the evidently diachronus nature of the first sediments onlapping the unconformity surfaces. This agrees with the opinion of Cowie *et al.* (1986) "*An obvious boundary should be suspect*". Even if the isochrony of each cycle of higher orders (global cycles after Haq *et al.* (1987, 1988) may be stated, the exact age of the onlapping sediments seems to depend strongly on the morphology of the basin and on the physical location of each examined section.

Problems in continental biostratigraphy

The use of biostratigraphy in continental sequences presents in its turn various types of problems. Within such sediments fossils are not usually rock-forming elements, as is true for most marine sediments. Fossils are usually scattered, and mainly as for the large land vertebrates only partially preserved and displaced within sedimentary traps. Very rare but not impossible (see the case of Passo Palade section in the Southern Alps: Avanzini & Neri, 1998), are the cases in which fossils of different ages or evolutionary levels are superimposed in the same sections. This presents a major difficulty to the needs of stratigraphy; a GSSP definition starts with a marker event that may be the first appearance or last occurrence of a fossil species. "*... First appearances are generally more reliable than extinction events, especially if the gradual transition between the marker and its ancestor can be observed. ...*" (Remane *et al.*, 1996).

One of the requirements for a GSSP which is proper and pertinent to our situation, is: "*Favourable facies for long-range biostratigraphic correlations; this will normally correspond to an open marine environment where species with wide geographic range will be more common than in coastal and continental settings. The latter should therefore be avoided*". (Remane *et al.*, 1996).

Moreover many types of biozones (obviously, except the chronobiozones) need to be bounded by isochronous surfaces; thus these types of subdivision also partially run into the above-mentioned problems.

The most frequently used biozones for continental sediments are a kind of Oppel's zone; a type of biozone that last that will disappear from the international codes, because it can be assimilated with the Assemblage-zone or the Multi-taxon-concurrent-range zone (Salvador, 1994). Actually the biostratographic approach uses the Interval-zone for preference, instead of the Oppel's zone. Also discarding these theoretical problems we know that a practical, reliable solution is a long way off. Indeed, in recent decades some attempts have been made to use biostratigraphy for solving such problems; reptile-based biozones or associations were repeatedly proposed by many authors (*e.g.* Bonaparte, 1973; Bakker, 1977; Anderson & Cruikshank, 1978; Lucas, 1999) (see also Lucas, 1996 for a synthesis of Permian biostratigraphy). All these attempts must be considered difficult to apply, and only suitable for a few exceptional and isolated basins, in confined regions. This may be due to:

1. the characteristics of land vertebrates, which are more provincial with respect to marine animals;
2. the continental deposits which have little potential for fossil preservation;
3. the fact that, within such deposits, fossil finds are scattered and punctuated events, mainly restricted to certain layers (Lucas, 1998 b).

The same considerations also seem to apply for the biostratigraphy based on tetrapod footprints. Ichnostratigraphic schemes or ichnozones were proposed by Holub & Kozur (1981), Ellemberger (1983 a, b, 1984) and Boy & Fichter (1988 a, b), while Gand (1987) and Gand & Haubold (1988) suggested the use of ichnoassociations for the Permian. Demathieu & Haubold (1974) and Haubold (1986) suggested the use of ichnostratigraphy for Triassic time.

However the discontinuous record, the incorrect theoretical approach and very "provincial" systematics made these attempts unsatisfactory, putting aside the different opinions on ichnotaxa and the resultant over-splitting that makes correlation difficult.

In conclusion, we must emphasise that in continental deposits the basal principles informing the stratigraphy (superimposition and continuity) are almost impossible to apply. The practical and theoretical impossibility of recognising boundary stratotypes, establishing stages and, frequently, setting biozones for fossils with discontinuous distribution are huge obstacles to the correct and useful time subdivision of continental sediments. All the above makes the use of lithostratigraphy, chronostratigraphy and geochronology unreliable and inhibits the use of biostratigraphy. To these general problems we have to add some particular difficulties related to the stratigraphic nomenclature of the Permian and Triassic continental deposits.

Problems in Permian and Triassic stratigraphic nomenclature

Inconsistencies in stratigraphy of continental deposits of the European Permo-Triassic interval have also stemmed from a number of nomenclature problems that depend with several causes. Perhaps the main one is the traditional use, with geochronological or chronostratigraphical meaning, of units that are only lithostratigraphical in nature. For instance, Rotliegendes-Zechstein and Buntsandstein-Muschelkalk-Keuper are historical names from German stratigraphy, respectively originating from the bipartite lithostratigraphic subdivision of Central European Permian sediments (Dyas) and from the traditional tripartite Triassic (Menning, 2000 b). Their use as chronostratigraphical units poisoned the literature. In this way many important data previously published are today unusable and so it is impossible to recognise the exact position of many fossil-bearing levels.

In the same way, the frequently continued use of traditional names such as Autunian, Saxonian and Thuringian, corresponding neither with true stratotypes nor with valid biostratigraphical calibrations, gives rise to further confusion. Moreover, we do not agree with the aim of establishing auxiliary stratotype points in continental deposits (Broutin *et al.*, 1999) when GSSP are obviously in marine sequences (Remane *et al.*, 1996).

In conclusion, the landscape is bleak: lithostratigraphy seems useless, biostratigraphical and chronostratigraphical units are impossible to establish and difficult to use. We must again remember that the only available correlation units are the geochronological standard units. These are not only based on marine sediments, and thus often lacking in elements of direct correlation with continental units, but in their turn are frequently poorly-defined and confused. Moreover, the long-running and unresolved debates on these topics, frequently repeated in a cyclic way, seem to indicate that one abandons the correlation of sections, levels, fossils and events because one is too busy correlating names.

With such a desolate general view, the only possibility seems to be a change of approach and trying to utilise a system less related to the standards and which does not require all the theoretical and practical formal necessities of the other branches of stratigraphy. Then instead of repeatedly trying again direct correlation among taxa or the use of biozones, we chose to test the possibility of using tetrapod footprints as faunal elements within evolutionary units (Faunal Units). These last seem easier to use; being less formally constrained than the other stratigraphic units.

Importance of vertebrate ichnology in Permian and Triassic red beds

It is soon evident that tetrapod footprints are an important component of Permian and Triassic faunal elements. They are well known in the literature and frequently are recorded when other fossils are lacking. Now we need to consider whether they have the necessary characteristics to be used for stratigraphy (rate of evolution, dispersion, confidence, etc.).

The first condition requires that different forms show the consequences of evolution, that is to say, that they changed in an irreversible way as time passed, that changes occurred with a rate convenient to the time period, and that it is possible to recognise ancestor-descendant relationships. Clearly, footprint and trackway changes are related to the evolution of the foot, the gait and other features of reptilian behaviour.

During Permian and Triassic times, reptiles underwent major diversification and gave rise to most of the groups (Benton, 1993), so that by the end of the Triassic almost all the main patterns were already distinguishable. Consequently, in the same interval reptilian feet show the greatest variability, going from the stem-reptile foot to a mammalian foot, and from basal crocodilomorph till to the true dinosaurian foot (avian pes).

During the transition "advanced ornithosuchids-primitive dinosaurs", the foot evolved rapidly towards bipedality and the functional tridactylity of the pes. It had already reached the "avian" pattern by earliest Carnian. After that, the dinosaurian pes pattern remained substantially unchanged, at least for the theropod (non-avian and avian), for the stem-ornithischians [or basal ornithischians] and for basal prosauropods. This situation changed in the Early Jurassic, with the appearance of the quadrupedal or semi quadrupedal herbivores, like large prosauropods, sauropods and relatively large ornithopods, with their quite new and different feet. As a consequence, in Upper Triassic terrains there exists the first difficulty in subdividing footprints: the absolute domain of the dinosaurian footprints makes their use in stratigraphy less attractive, and their systematics less simple.

Thus at present we can easily recognise different associations only for Permian-Triassic time, associations in which stem-reptiles, mammal-like reptiles and thecodonts absolutely prevail over basal dinosaurs.

In conclusion, within the Permian and Triassic interval, so rich in continental sediments (geocratic), the use of ichnoassociations might be very useful for stratigraphy, even if preserving some intrinsic problems.

Problems in the philosophy of tetrapod-ichnology

The first theoretical problem is related to the two-fold philosophical interpretation that forms the basis of ichnology. Many researchers prefer a simple behavioural interpretation of traces, mainly those that come from the study of invertebrate ichnology ("ichnology mainstream", Bromley, oral comm., Halle Workshop, March 1997). They consider it almost impossible that an ichnofossil could be linked to a low-level zoological category. Such generalisation also implies that the same behaviours could repeat through time, and consequently give a very low degree of confidence in the stratigraphic use of ichnofossils.

Theoretically that concept does not fit within the concept of expanded phenotype, in which the behaviour is included and has to be considered like any other characteristic, realised on the basis of the dynamic and metabolic characteristics of an organism, pertaining to the genetics of a defined population or species.

In practice, if the behavioural interpretation could be consistent for a worm trace, it seems almost less and less valid for tetrapod trackways, frequently showing the traces of variable osteological and behavioural characteristics.

A completely different interpretation admits in principle the possibility for a paleontologist to be able to recognise a direct correspondence between footprint and trackmaker taxonomic group. It is clear to ichnologists that an ichnospecies does not correspond to a species, but it is also clear that careful taxonomic analysis and correct classification would correspond to a well-defined, if higher, taxon like a genus or a family.

If the latter interpretation is accepted, one should be able to use ichnofossils in stratigraphy with the same dignity and confidence of body fossils. This possibility was recognised by Haubold (1971 a, b) and Conti *et al.* (1977) and is implicit and clearly demonstrated by the different footprint associations already listed on a time scale by various authors (*i.e.* Haubold, 1971 a, b, 1986; Gand, 1987). Consequently we believe to be possible the use of reptile footprints for time subdivision, in contrast to the opinions expressed by Lucas (1996). This author, in analysing Permian reptile distribution, suggested a three-fold subdivision for the continental Permian similar to Romer (1973), and subsequently (Lucas, 1998 b) excluded the practical possibility of using footprints for correlation. Lucas, maintaining the impossibility of recognising the correspondence between ichnogenera and taxa below the family level, stated that "*... orders, superfamilies and families of body-fossil taxa* (of Permian reptile) *have stratigraphic ranges on the magnitude of series, so ichnobiostratigraphy based on the ichnotaxa can only discriminate correlations at the series level. ...*" (Lucas, 1998 b, p. 21).

Such an approach to the problem shows an evident weakness, misinterpreting our caution in limiting zoological attribution to family level as an actual uniformity within families. If one finds a cat-trackway and a tiger-trackway and classifies both ichnotaxa just as Felidae that does not mean that a tiger and a cat are the same animal and share the same vertical and geographical distribution. Moreover, the practical impossibility of reaching a well-defined classification, as in common for paleontologists when fossils are incomplete or poorly preserved, cannot be set as a principle, since the continual and progressive refinement of our knowledge might change this situation completely.

Lucas' way of thinking also leads to a strange interpretation. According to that system we must ascribe an ichnofossil to a family of the bone-based systematics and then use the recorded temporal distribution of families for correlation. However, stratigraphy is a practical tool, always based on the recognition of a sequence of events or forms (or of unknown objects) which repeats with the same order in different sections (homotaxy). After this process and that of calibration, correlation is possible. That also excludes consideration of the zoological attribution of the utilised taxa (*e. g.* the stratigraphy based on *incertae sedis* fossils).

We will now examine the characteristics of biochronology. We believe that the biochronological value of a fossil group is determined by the succession of its different evolutionary stages. Once the succession of different forms is verified, their evolutionary status can be determined on the grounds of eco-evolutionary valuations or of ancestor-descendant relationships within lineages.

Since our aim is to establish ichnofaunal units, we must also consider the relationships of the taxa within an association. The consistency of each taxon within an association depends not only on the evolutionary stages but also on ecological and paleogeographical characteristics. Evolutionary equivalent associations can have partially or totally different compositions because of varying environments or due to the paleogeographycal positions of depositional basins in which track-bearing sediments were laid down.

Only after having established a sequence of consecutive evolutionary steps can we try to correlate to standards, although the biochronological value of the sequences is valid even if correlation is still doubtful. The chronometric value of the evolutionary stages can be assured by calibrations with external data, and only by correlation with Ages (based on marine standard Stages).

Moreover, by the study of the associations it is possible to recognise one or more index forms whose lifetime corresponds with that of the whole association. If recognised such forms share the same biochronological value as the whole association. We must keep in mind that this kind of marker can be useful but is not necessary. Also first and last appearance datum (FAD, LAD), first and last occurrence (FO, LO) and other boundary events can be useful but are not necessary.

All that considered, the remaining obstacles to the use of ichnofossils consist of the difficult taxonomy, the resulting confused systematics, and (only marginally) the uncertain zoological value of ascribing ichnogenera and ichnospecies. In fact, these are practical and temporary problems and not theoretical obstacles to the use of ichnofossils in stratigraphy.

Problems in ichnological taxonomy

Once stated that, from the theoretical point of view, tetrapod ichnofossils can be used for stratigraphy, we will briefly examine the present position of ichnotaxonomy. Tetrapod footprint taxonomy has many problems, already largely debated but also possible to overcome with the present state of the art. The Workshop on Ichnofacies and Ichnotaxonomy of the Terrestrial Permian, held at the Martin Luther University in Halle (Germany) in spring 1997, helped to change and reinforce some ideas on ichnotaxonomy and ichnosystematics. Moreover, discussions underlined the influence of extramorphologies and pushed towards an ever-increasing caution in establishing new taxa. As a result of that meeting, we can consider that taxonomy in vertebrate ichnology has reached a greater degree of maturity. The ensuing simplification of the enormous number of names present in the literature concerned with Permian vertebrate ichnology is in progress (Haubold, 1996; Conti *et al.*, 2000). The same simplification is also in progress today for inflationary nomenclature on small dinosaurian Triassic footprints (Olsen & Galton, 1984; Leonardi & Lockley, 1995; Leonardi, 2000).

It is time to discuss another point, before passing to the actual study of footprint-based chronology; this is the problem of the relationship between ichnological and zoological systematics.

Problems in ichnosystematics

Today footprints enjoy a sufficiently mature taxonomy while the ichnological systematics and nomenclature are still influenced by their well-known historical problems. Depending on the different philosophies and sometimes on the enthusiasm and inexperience of the classifiers; as well as *"systematic obstinacy"* (the classification at all hazards and at any rate; *"accanimento sistematico" sensu* Conti *et al.*, 2000), we can have:

1. taxa absolutely split, in which different names correspond simply to different attitudes (behaviour and gait) of the track-makers (see Ellemberger, 1970, 1972, 1974) or to a type of gait or to any characteristic producing repetitive extramorphologies. In fact, in simply recognising the sliding of a pes or a particular type of impression in a normal trackway, all similar footprints among those in a trackway are automatically considered normal footprints of a different animal [see Haubold (1996) for the concept of "phantom" taxa]. An absolutely demonstrative case is that of *Chelichnus* (or *Laoporus*) in which a regularly-placed sand-sliding, in a dune environment, distinguished all the footprints and was considered as a taxonomical feature (Haubold, 1996);
2. huge ichnogenera, which consist of the sum of many previously existing taxa. This case is often a kind of surrender, when one realises our inability to subdivide footprints, either because of an overlap of characteristics, or because of the coherent (even if incorrect) knowledge that some kinds of footprints are not sufficient to define a track-maker. Sometimes, after a taxon is correctly established, with its variability, further data are added to the original variability. They lead the taxon variability range to overlap to ranges of other taxa that were well separated when originally established. In this case, a too cautious re-

vision can create some problems because the different names converge to create enormous synonymic lists and equally enormous variability. Such resulting variability is able in practice able to include many different forms. A good example of this case is *Grallator* (*i.e.* Olsen & Galton, 1984) now inclusive of the three old genera *Grallator*, *Anchisauripus* and *Eubrontes*, and probably most of the species previously ascribed to the invalid name *Coelurosaurichnus* (Leonardi & Lockley, 1995);

3. some rarer cases represented by ichnogenera including more species, in this case one may perceive a correspondence between ichnospecies and fossil species (or populations characterised by particular behaviours), *e.g.* the case of ichnogenus *Rhynchosauroides*;

4. ichnogenera, usually monotypic, for which there exists wide sampling and a well-described dimensional and gait variability; in conclusion, taxa well known and rather reliable as in the case of *Ichniotherium*. These last represent solid and reliable bases for stratigraphy, and it is desirable that after in-depth and up-to-date studies, all the taxa will reach this condition.

PERMIAN AND TRIASSIC FOOTPRINT-BASED BIOCHRONOLOGY

After examination of the problems, we believe that all the above-mentioned theoretical difficulties will be overcome. Consequently, tetrapod footprints appear available for assessing the age of continental deposits, if using a biochronological approach. We concluded that to be sure of such an assumption, we only needed a complete practical test on this topic.

We tested the possibility of recognising evolutionary stages, setting them in a sequence of evolutionary units, and using these as chronological tools. In applying this method, we will follow a path closely parallel to mainstream land vertebrate stratigraphy (Walsh, 1998).

Permian and Triassic paleogeography and the geological setting of the Southern Alps

We believe that examples of all the aforementioned problems in continental stratigraphy are present in the Permian and Triassic geology of the Southern Alps.

The Upper Carboniferous-Upper Permian succession of the Southern Alps is separated into two major tectonosedimentary cycles by a regional unconformity:

- a lower cycle represented by calcalkaline acidic to intermediate volcanic and alluvial-lacustrine continental deposits (Collio Fm.; Dosso dei Galli Cgl.; Auccia Volcanites; Ponteranica Cgl.; Tregiovo Fm.; and ignimbrites and lavas of the Atesino Volcanic District);
- an upper cycle represented by fluvial red clastics of Verrucano Lombardo and Val Gardena Sandstone, laterally and vertically replaced in part by sulphate evaporites and shallow-marine carbonate sequences (Bellerophon Fm.).

The regional unconformity between the two cycles is well documented by extensive erosional surfaces and paleosoil horizons. The stratigraphic break is different in different outcrops, with a time-gap ranging 14 to 27 Ma (Italian IGCP 203 Group, 1986; Cassinis *et al.*, 1988, 1999).

The continental Permian units from which we collected the fauna were the Collio Fm., the Dosso dei Galli Cgl., and the Tregiovo Fm. from the first cycle. In second cycle, in the Dolomites area, the fossiliferous units were the Val Gardena Sandstone and the Bellerophon Fm.

The Triassic stratigraphy in the Dolomites and the interpretation in terms of sequence stratigraphy have been made possible by a highly resolved ammonite standard scale (Mietto & Manfrin, 1995 a, b). This standard scale is based on the definition of a zone succession characterised by genera; in turn each zone is subdivided into a number of subzones defined by species. Within the Middle Triassic these subzones allow a biochronostratigraphic resolution which can be extended throughout the Tethyan region.

The Triassic succession in the Dolomites and surrounding areas can be schematised as follows:

on the whole the Lower Triassic consists of terrigenous and terrigenous-carbonate units essentially deposited in shallow marine to tidal flat environments (Werfen Fm.). The Scythian–Anisian boundary seems to occur at a peritidal carbonate platform which extended throughout the Southern Alps (Lower Serla Dolomite, Lusnizza Fm.). Several terrigenous and terrigenous carbonate units follow, essentially Anisian in age (Braies Group, Upper Serla Fm. and Contrin Fm. - Pisa *et al.*, 1979; De Zanche & Farabegoli, 1982), including a number of rock units deposited in basinal, lagoonal, peritidal and continental environments. The Ladinian interval is represented in the Dolomites by basinal units (Buchenstein and Wengen Fms) and carbonate platforms (Sciliar Dolomite). In the Carnian, carbonate platforms, marls, shales and volcanic siltstones and sandstones are recognisable (San Cassiano Fm., Dürrenstein Fm.). An interval of vari-coloured terrigenous-carbonate rocks of Late Carnian age (Raibl Fm.) is covered by the carbonate tidal-flat deposits of the Dolomia Principale Fm. (uppermost Carnian to Rhaetian).

The continental units, interbedded with marine units, from which we collected the ichnofauna, were the Werfen Fm., the Braies Group, the Dürrenstein Fm., the Dolomia Principale Fm.

Tetrapod footprint database

A first attempt in using biochronology was proposed by

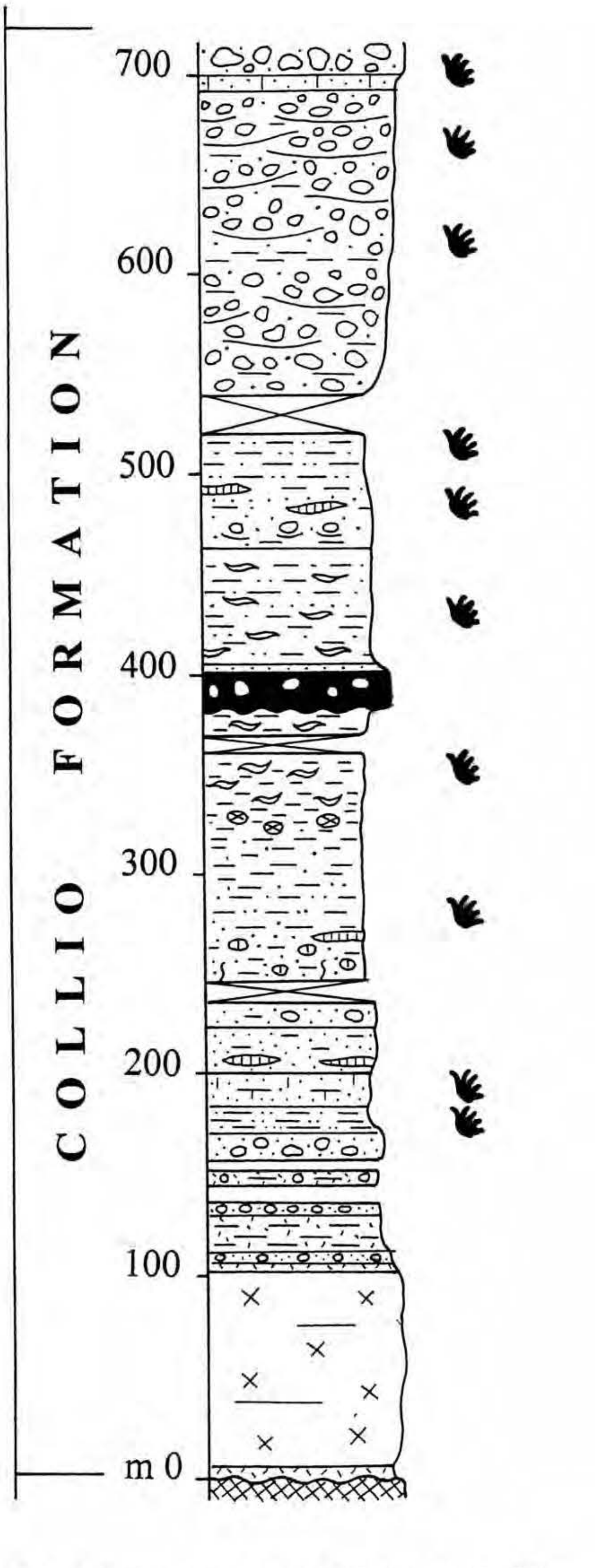

Fig. 1 – Schematic columnar section of the Collio Fm. (Early Permian) at the Dasdana Valley (Brescia, northern Italy). Footprint outlines indicate footprint levels (Modified from Italian IGCP 203 Group, 1986 and Ceoloni *et al.*, 1987).

Conti *et al.*, (1997) for the Permian; this second attempt, for the Permian and Triassic interval, is based on an updated and very much richer ichnological database. In the last few years, the ichnological literature on northern Italy has increased enormously due to the fact that many new tetrapod footprint-bearing outcrops have been found and studied. In the same way, the number of researchers partly or totally devoted to this type of fossils has increased. The recent developments in ichnology are linked both to increasing knowledge of the stratigraphical, paleoecological and paleogeographical importance of footprints and to the ever increasing number of new finds. The development is also demonstrated by the quasi-exponential growth in the number of papers on tetrapod ichnology.

All together in Italy, up to summer 1999, 35 tetrapod footprint-rich outcrops were detected, 32 of them present in the southern Alpine belt. Such outcrops obviously differ in the types of footprints (amphibians, stem-reptiles, "thecodonts", phytosaurs, dinosaurs), in the frequency of finds (single levels or multiple levels inside the same section), in dimensions of the outcrop (from a few square decimetre-sized slabs to kilometre-scale track-sites), and in degree of ichnodiversity.

After a coarse stratigraphical calibration, we selected all the data from Permian, Triassic and Lower Jurassic outcrops to check the stratigraphical value of the footprints. In many cases the systematics are still in progress, and frequently the names given to the trackway still have an uncertain systematic position. Nevertheless all the data falling in the Permian-Triassic interval were used. In order to escape temporary systematic problems, in this analysis of feasibility we used widely inclusive names, aware that many of these names represent taxa that will eventually be given a well-defined meaning. The only rule was that the names had to correspond to different objects, so that they were easy to recognise and easy to use.

Calibration data

Ichnological results were calibrated with all the other available data at our disposal from the quite rich literature (see Cassinis *et al.*, 1999, for a bibliographic review). Data under consideration were different depending on the different paleogeographical and environmental framework. In our case the tetrapod footprints are recorded from different paleoenvironments: Lower and Upper Permian outcrops are mainly within continental deposits and marine events are recorded only at the very top of the Upper Permian sections; in contrast, Triassic footprint-bearing levels crop out mainly within marine sequences.

For calibration we used data on:

a. sporomorphs, megaplants, isotopic dating, sequence stratigraphy for Lower Permian outcrops;

b. sporomorphs, megaplants, foraminifers, algae, bra-

chiopods, nautiloids, conodonts, sequence stratigraphy for Upper Permian tracksites;
c. ammonites, conodonts, sporomorphs, sequence stratigraphy for Triassic tracksites. Only in the last case was possible to use ammonite standard biozonation and then a calibration with marine standard chronostratigraphical and geochronological units.

Intrabasinal correlation and association analysis

In the most outcrops footprints were present in isolated layers, but we had the good fortune to be able to use, as main reference outcrops, three sections in which thick sequences of track-bearing sediments were superimposed.

Within these outcrops ichnofaunas are well represented and widespread, and have mostly been previously studied (see below). We compared all the recognised levels and the isolated findings with the associations recorded from the main sections. The results, all consistent, are summarised below.

Lower Permian Ichnoassociation

This association includes all the ichnotaxa recognised within the first cycle sediments of the Alpine Permian-Triassic sequence. In the past this was tentatively considered as a faunal unit under the name of Collio Faunal Unit, further subdivided into two faunal subunits named as the Pulpito Faunal Subunit and the Tregiovo Faunal Subunit (Conti *et al.*, 1997). To each of them was also ascribed a biochronological value with the names of the Collio Faunal Age (Rabejac Faunal Subage and Tregiovo Faunal Subage).

The ichnoassociation is based on data from the Dasdana Valley section (Collio Fm.; Cassinis, 1966), in which ten footprint-bearing layers were recognised through a nearly 700 m thick section (Geinitz, 1869; Curioni, 1870; Berruti, 1969; Ceoloni *et al.*, 1987; Conti *et al.*, 1991, 2000).

After a partial revision, the ichnoassociation includes reptile footprints (?*Camunipes cassinisi, Varanopus curvidactylus, Amphisauropus latus, Ichniotherium cottae, Dromopus lacertoides, D. didactylus*) and less frequent amphibian footprints (*Batrachichnus* sp.) (Fig. 1).

This set of data was successfully compared with data coming from sediments cropping out in other parts of the Orobic Basin (Gümbel, 1880; Dozy, 1935; Casati, 1969; Casati & Forcella, 1988; Nicosia *et al.*, 1999 a, 2000; Cassinis *et al.*, 2000; Santi, in press; Santi & Krieger, in press) and in the Tregiovo Basin (Conti *et al.*, 1997).

Correlational elements with marine deposits are not possible, but we can use floristic data (Cassinis & Doubinger, 1991 a, b) for correlation, and we can constrain the ages of the base and the top by using radiometric data (Cassinis *et al.*, in press). Actually, the time interval in which this fauna is widespread in northern Italy is limited to between 286/283 Ma BP, at the base, and 278/273 Ma BP at the top (Cassinis *et al.*, 1999 and in press).

Upper Permian Ichnoassociation

This ichnoassociation includes all the taxa recognised within the second cycle sediments. In Conti *et al.* (1997) this association was also considered to be a faunal unit under the name of Bletterbach FU, with a faunal age named Bletterbach FA.

The ichnoassociation is based on data from the Bletterbach section in which 12 layers were recognised in a thickness of nearly 180 m (Ceoloni *et al.*, 1988 a, b; Conti *et al.*, 1975, 1977, 1980 a, b, 1987; Leonardi & Nicosia, 1973; Leonardi *et al.*, 1975; Nicosia *et al.*, 1999 b).

The ichnoassociation is composed of footprints ascribed to reptiles (mammal-like, prolacertiforms, pareiasaurids) characterised by a very advanced evolutionary stage. Among them only some taxa were selected to have a more confident database (*e.g. Pachypes dolomiticus, Ichniotherium accordii, Rhynchosauroides pallinii, Dicynodontipus* sp.) (Fig. 2). The other outcrops bearing the same fauna were the following: SS. 48, Monte Cislon (Kittl, 1891; Abel, 1929); S. Pellegrino Pass, Seceda (Conti *et al.*, 1977), Nova Ponente (Wopfner, 1999) in the Dolomites, Ligosullo in the Carnia region (Mietto & Muscio, 1987), Recoaro near Vicenza (Mietto, 1975, 1981, 1995), the San Genesio-Meltina Plateau in the Adige Basin, and Mt. Luco in the Tregiovo Basin (Avanzini, unpublished).

The only elements that allow correlation are the Late Permian marine events at the top and the sporomorphs at the base. Nevertheless we were able to constrain the association to an interval ranging from nearly 259 to 255 Ma BP (Cassinis *et al.*, in press).

The two Permian associations still present a big problem; their evolutionary stages are very different, and between the Lower Permian and Upper Permian ichnofaunas there seems to be an evolutionary jump. From that point of view we can suppose that the lack of an ichnofaunal record for such a long interval is related to a lack of sediment. Indeed, in the Alpine region, sediments must have been lacking between the first-cycle and second-cycle sediments, because of the enormous gap, lasting 14 to 27 million years (Cassinis *et al.*, 1999, p. 10). So different chronometric ages of track-bearing sediments validate the gap between evolutionary stages. We can also observe that the same gap and the resulting change in evolutionary stage seems widespread all over the world.

Lowermost Triassic Ichnoassociation

Few footprints are recorded from the Lower Triassic sediments, Recoaro, Val Gardena (Bulla/Pufels) and Val Tra-

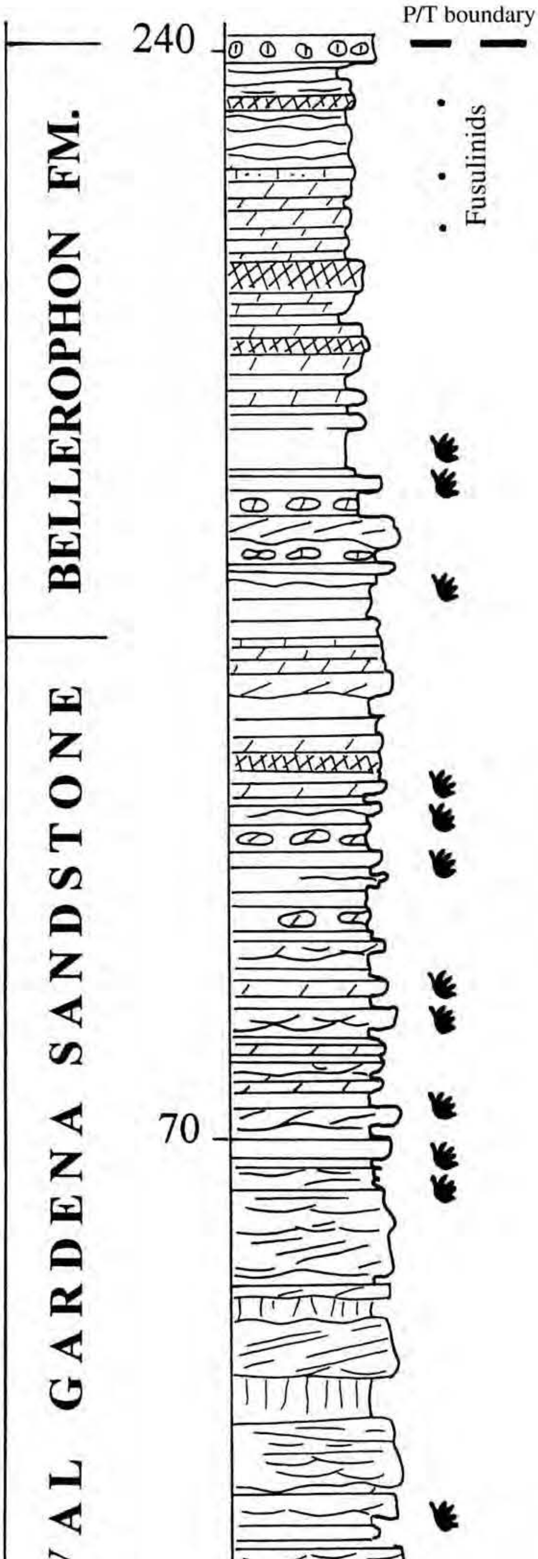

Fig. 2 – Schematic columnar section of the Val Gardena Sandstone and Bellerophon Fm. (Late Permian) at the Bletterbach Gorge (Bozen, northern Italy). Footprint outlines indicate footprint levels. (Modified from Massari *et al.*, 1988).

vignolo (Dolomites) and also in Carnia (Conti *et al.*, 2000). Some forms, ascribable to *Rhynchosauroides* sp. cfr. *R. schochardti*, were collected from the terrigenous layers of the Werfen Formation (Mietto, 1986). Other taxa, termed "Pseudosuchians" by Leonardi (1967), are to date poorly documented and represented only by incomplete specimens. The upper part of this interval is calibrated to Olenekian (Spathian) age by ammonites (*Tirolites* zone).

Mid-Triassic Ichnoassociation

Several ichnoassociations, some of which are still under study, have been discovered in the Dolomites and surrounding areas (Abel, 1926; Brandner, 1973; Mietto, 1987, 1995, 2000, and unpublished data; Sirna *et al.*, 1994).

Tracksites are located in the Braies Dolomites (northern Dolomites) (Abel, 1926; Brandner, 1973), in the eastern Dolomites (Conti *et al.*, 2000; Mietto, 2000), in the upper Val di Non and Val d'Adige, (Avanzini & Neri, 1998, Avanzini, in press) and in the Recoaro area (Mietto, 1987, 1995).

Most of the tracks pertain to Lepidosauria of the *Rhynchosauroides* ichnogenus with the typical *Rhynchosauroides tirolicus* (Abel, 1926). Subordinately Archosauria tracks have been recognised (*Synaptichnium priscum, Synaptichnium pseudosuchoides, ?Synaptichnium cameronense, Parasynaptichnium gracilis, Chirotherium rex, Chirotherium barthii, Brachychiroterium parvum, Brachychirotherium* aff. *parvum, Isochirotherium delicatum*). The sites also yielded tracks that may be ascribed to amphibians, chelonians and therapsids.

The track-bearing units of the eastern Southern Alps identify a transitional, continental-to-marine environment, characterised by terrigenous and carbonate platforms and coastal delta mouth bars deposited under relatively arid conditions. These deposits can be calibrated, using ammonites and other marine faunas, to the whole Anisian time interval.

Upper Triassic Ichnoassociations

Tracks of terrestrial reptiles were found in Upper Triassic units (Dürrenstein Fm., Dolomia Principale Fm. and equivalent units) of the Dolomites (Mietto, 1988, 1990, 1992; Leonardi & Avanzini, 1994; Avanzini *et al.*, in press) and the Carnic Pre-Alps (Dalla Vecchia & Mietto, 1998; Dalla Vecchia, 1996; Roghi & Dalla Vecchia, 1997).

The first ichnoassociation includes tracks made by "thecodonts" (chirotheroid tracks), phitosaurs, mammal-like reptiles and probable basal dinosaurs. The age of the sequence has not yet been well defined. Data from outside the Dolomites (eastern Southern Alps) seem to indicate a Carnian age (Latest Julian-Early Tuvalian) using ammonites (*Austriacum-Dilleri* zone).

A higher association is made by footprints ascribed to medium-to-large theropod dinosaurs (*Eubrontes, An-

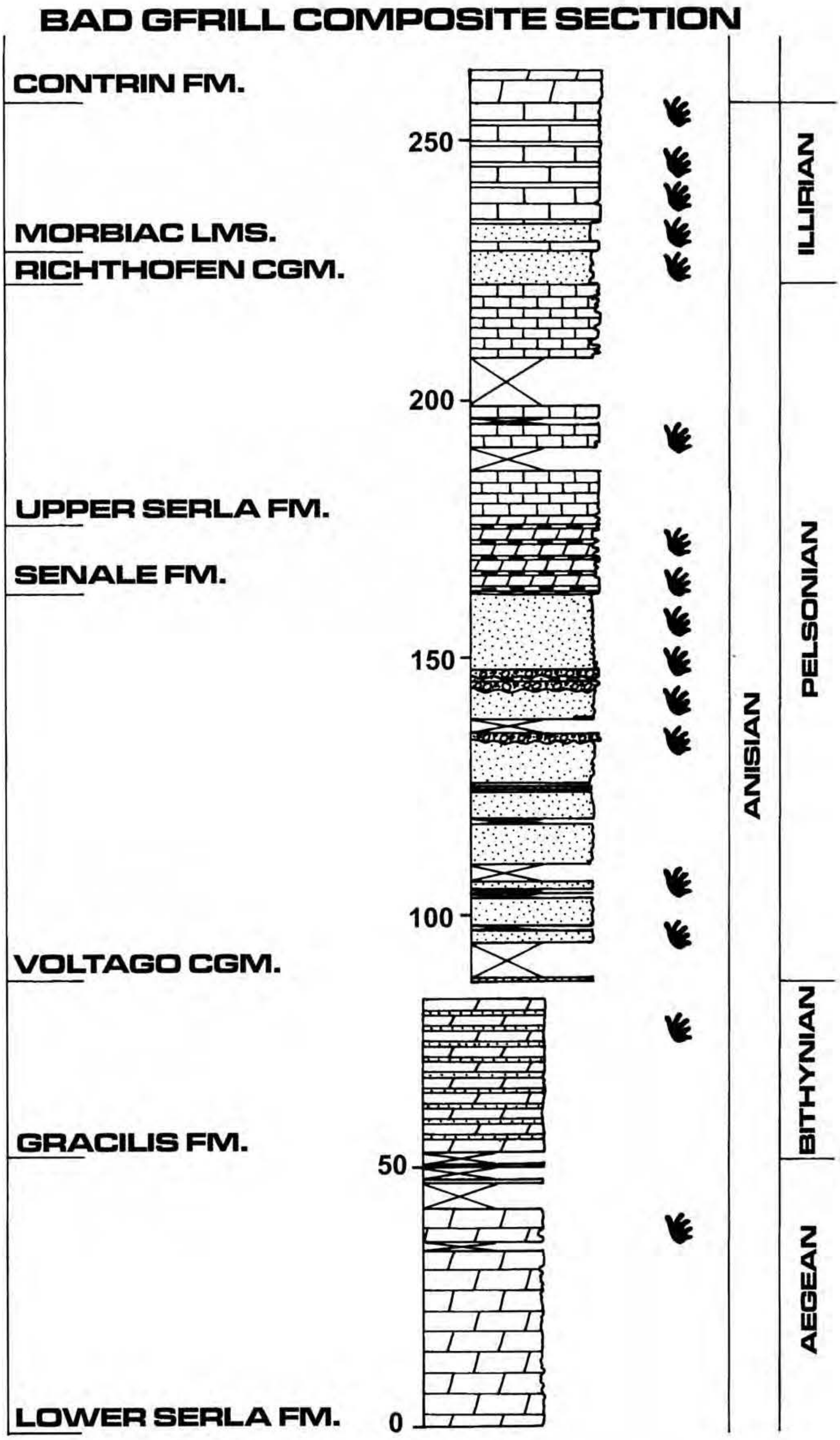

Fig. 3 - Schematic columnar section of the Mid-Triassic mainly marine sequence at the Bad Gfrill section (Bozen, northern Italy). Footprint outlines indicate footprint levels. Only for this type of sequence is it possible to use marine standard ages.

chisauripus), "thecodonts", and basal Ornithischia. The presence of prosauropods is also probable. This association is Late Carnian (Late Tuvalian) to Early Norian in age due to the presence of ammonites in the Carnic equivalent basal sequences (*Anatropites* zone).

The Triassic faunal assemblage of the Southern Alps is very interesting because track-bearing layers are interbedded with marine layers (Fig. 3). It was found possible to use ammonite standard biozonation and the resulting calibration with marine standard chronostratigraphical units.

The most recent data on the Anisian levels, the interval in which we find the most complex and rich ichnoassociations, lead us to hypothesise that it will be possible to use tetrapod footprint associations with the same resolution as ammonite biozones.

We also believe it is possible that the calibration we are performing in the Southern Alps could yield decisive results in dating sedimentary sequences still lacking in chronostratigraphical calibration.

Extrabasinal correlations

The final step, still to be accomplished, is a coarse correlation of our results with data listed from outcrops all over the world.

1. We can consider our Lower Permian Ichnoassociation as at the same evolutionary stage as the ichnoassociations from:

a. Cape John Formation, Nova Scotia, Canada (Hunt, oral com.);

b. the De Chelly Sandstone (Haubold *et al.*, 1995 b; Morales & Haubold, 1995); the Robledo Mountains Member of the Hueco Fm. (Hunt *et al.*, 1995 a; Lucas *et al.*, 1995 b); the Abo Fm. (Lucas *et al.*, 1995 a), the Earp Fm., the Sangre de Cristo Fm., New Mexico, (Lucas & Hunt, 1995); and many other formations from U.S.A. (Haubold *et al.*, 1995 a; Hunt *et al.*, 1995 b);

c. the Clear Fork Group from Castle Peak, Texas , U.S.A. (Sarjeant, 1971);

d. the Keele and Enville beds of the Birmingham region, UK (Haubold & Sarjeant, 1974);

e. the whole Nahe-Gruppe in SW Germany (Boy & Fichter, 1988 a, b);

f. all the formations of the Thuringian Forest Rotliegend up to the Tambach Fm. (Lützner, 1987; Sumida *et al.*, 1996);

g. all the formations of the Lodève, Saint-Affrique and Provençal (Demathieu & Gand, 1992) basins listed in Gand (1987) and Gand *et al.* (1995) and, generally, from all the French basins;

h. from the Aksautskian and Kinyrtchdskian formations of the northern Caucasus, Russia (few and poorly preserved specimens, Lucas *et al.*, 1999).

2. Possibly the same evolutionary stage, as shown by our Upper Permian Ichnoassociation, can be ascribed to the specimen of *Paradoxichnium problematicum* Müller, 1959 from the "Zechstein bei Gera" (Müller, 1959).

3. The same evolutionary stage, as shown by our Lower Triassic Ichnoassociation, could be ascribed to:

a. the Solling and Kronach ichnofaunas from the Buntsandstein Fm. (Germany) (Demathieu & Leitz, 1982; Demathieu, 1984);

b. the Solliès – Ville and Sanary ichnofaunas from the Buntsandstein Fm. (Provence, France) (Charles, 1949) and the "Werfénien" Fm. (Ellemberger, 1965);

c. the Lodève and Dio ichnofaunas from the Buntsandstein Fm. (Massif Central – France) (Gervais, 1857; Orszag-Sperber, 1966);

d. the Jaegerthal (Schimper, 1850), Saint-Valbert (Daubrée, 1857) and Granges la Ville (Buffard, 1966) ichnofaunas from the Buntsandstein Fm. of Vosges (Germany-Belgium-France);

e. the Roberts Fm. Ichnoassociation (Spathian in age) from the Alpes Maritimes of France (Demathieu, 1977).

4. We can consider the basal Middle Triassic (Lower and Middle Anisian) Ichnoassociation of the Southern Alps as at the same evolutionary stage as the ichnoassociations from:

a. Trémonzey/Vosge (France) from the Upper Buntsandstein Fm. (Demathieu & Durand, 1975);

b. Winterswijk (Holland) (Demathieu & Oosterink, 1983)

c. Lodève (France) (Demathieu, 1984);

d. the Holbrook Mb. of the Moenkopi Fm. (Peabody, 1948, Hunt & Lucas, 1993);

e. Fm. de Rimplas (Grès de Gonfaron) and Fromagine Fm. from the Alpes Maritimes of France (Demathieu, 1977).

5. The Middle Triassic (Upper Anisian) Ichnoassociation was at the same evolutionary stage as the ichnoassociations from the Muschelkalk of the Massif Central, Vosges (Granges la Ville), and Provence (Sanary) (Ellemberger, 1965; Buffard, 1966; Courel *et al.*, 1968; Courel & Demathieu, 1973).

6. We can consider our Upper Triassic (Carnian – Norian) Ichnoassociation as at the same evolutionary stage of the ichnoassociations coming from:

a. the Dokum, Lockatong, Stockton, Passaic (= Lower Brunswick) and Wolfville formations of the Newark Supergroup (North America) (Olsen, 1980, 1983);

b. the Middle Keuper formations of Valaise, Languedoc and Germany (Europe) (Haubold, 1984, 1986);

c. the Trias of Vieux Emosson in Switzerland (Demathieu & Weidmann, 1982);

d. the Molteno Formation, Zones A/1-4 of Lesotho (South Africa) (Ellenberger, 1972, 1974);

e. the Ipswich Coal of Queensland (Australia);

f. the Mercia Mudstone Group of South Wales (England) (Lockley *et al.*, 1996).

At this time the analysis is still in progress, but it seems that our results could usefully be applied to most of the published material.

FAUNAL UNITS AND ICHNOFAUNAL UNITS

In Fig. 4 the data for northern Italian ichnofaunas are plotted on a geochronological scheme. According to our analyses and all the available calibration data, it seems well proven that six groups of footprints can distinguish six discrete intervals in the Permian-Triassic periods. Each interval is characterised by different taxa, showing different and successive evolutionary levels.

At this point we have only to discuss:

1. with which type of unit these intervals can be associated;
2. their validity for correlation to basin and extrabasin level;
3. if they are suitable for defining the ages of the deposits.

It seems clear that reptile footprints can be used to distinguish more or less different associations, characterised by different evolutionary stages. As they, more or less correspond to Oppel's zones, and correspond to evolutionary units, they can be established as faunal units.

The evolutionary units or faunal units (FUs) are characterised by taxa, in general unique to the unit; moreover additional taxa may occur in more than one unit. The FUs can be named on the basis of characterising taxa or better by the most representative localities.

The FUs thus represent a co-evolved association of a variable number of coeval taxa. We know that within an association (and in light of the "Red Queen" theory), species change in co-evolutionary equilibrium, and thus the presence of one or a few of them could be sufficient to mark a well-defined evolutionary stage.

The FUs present optimal tools for continental deposits because they are based on associations and do not need markers, and because, from a theoretical point of view, they represent units that avoid all the problems of continuity and superimposition. They can be used in cases of single or punctuated finds. In practice they represent defining moments in biological evolution, and so, as associations, they are unrepeatable moments in time. The definition of FUs, even if requering experience and knowledge of the faunas, is possible and relatively simple, being based on sets of taxa borne of co-evolutionary processes and in ecologically characterised associations.

Thus each FU also represents the time interval of existence of the selected members of the association and thus the interval of persistence of such biological equilibria. Consequently a FU can easily be transformed into a faunal age (FA), or biochronological unit; each FA being typified by a find locality. A sequence of successive steps in evolution is so transformed into a sequence of FAs. It is obvious that without a well-fixed calibration a sequence of FAs just represents a sequence of time intervals, but after calibration they change into useful tools in the subdivision of sedimentary sequences and in correlation of non-superimposed deposits.

In the study of Tertiary-Quaternary continental deposits, which present similar problems, time subdivisions based on the concept of faunal ages are used, in their turn tied to evolutionary mammal associations (Walsh, 1998). This method is well-enough established (Wyss *et al.*, 1996; Pajak *et al.*, 1996; Walsh, 1998) that scales were organised into NALMAs (North American Land Mammals Ages) and SALMAs (South American Land Mammals Ages). The same use of evolutionary units was already implicit, although more or less hidden, within the concepts of "Ages-Reptiles" of Bonaparte (1973), "Reptiles Dynasties" of Bakker (1977) and within the "Empires" of Anderson & Cruikshank (1978). On each occasion these subdivisions were based on large collections of bone remains from sediments that did not have superimposed sections. Also the very recent Triassic stratigraphy (Lucas, 1999) is a suite of temporally successive bone-based assemblage zones.

From the analysis of the ichnological database it seems demonstrated that, on the same principles, it is possible that ichnoassociations (frequently classified just at ichnogenus level) can be used as FUs or FAs over the time scales of geological phenomena.

At the moment our ichnoassociations are still lacking in boundary events, but are well differentiated from an evolutionary point of view and well constrained from the chronometric point of view. We believe that, in the future, Land Ichnofaunal Units (LIUs) and the consistent Land Ichnofaunal Ages (LIAs) could be the most useful tools for subdividing and correlating Permian-Triassic continental sediments.

CONCLUSIONS

The present situation shows a general lack of stratigraphic markers, and the difficulty of using the traditional systems for assessing the age of Upper Paleozoic and Lower Mesozoic continental deposits. The only real alternative seems to be the use of non-formalised associations of fossils, each of them with a biochronological value and thus independent of the need for direct superposition and continuity.

For the Permian and Triassic continental deposits of the Alpine region, tetrapod footprints seem the best fossils for this purpose. They can be used to establish evolutionary units, once problems in the philosophy, systematics and nomenclature of the ichnofossils are overcome.

ICHNOTAXA	A	S	A	K	K	T	I	O	A	L	C	N/R	H	S	P
***Anomoepus* sp.**													□		
Ornitischia ind.												□	□		
***Parabrontopodus* sp.**													□		
Sauropoda ind.													□	□	
Prosauropoda ind.												□			
***Eubrontes* sp.**												□	□	□	
***Anchisauripus* sp.**											□				
***Grallator* sp.**												□	□		
Ceratosauria ind.											□	□	□	□	
"Phytosauria" ind.											□				
Therapsida ind.									□		□				
Procolophonida ind.									□						
Chelonomorpha ind.									□						
Isochir. delicatum									□						
***Isochirotheriun* sp.**									□						
Brachichir.* aff. *parvum									□						
***Brachichirotherium* sp.**									□		□				
Chirotherium rex									□						
Chirotherium barthi									□						
***Chirotherium* sp.**									□		□				
***Parasynaptichnium* sp.**									□						
***Synaptichnium* sp.**									□						
Chirotherida ind.						□	□		□		□				
Rhynch. tirolicus									□						
Rhynch. schochardti							□								
***Rhynchosauroides* sp.**						□	□		□						
Rhynch.* aff. *palmatus						□									
Rhynchosauroides pallinii						□									
***Dicynodontipus* sp.**						□									
Ichnioth.* aff. *cottae						□									
Ichniotherium accordii						□									
Pachypes dolomiticus						□									
Dromopus didactylus				□											
Dromopus lacertoides				□											
Amphisauropus latus				□											
Ichniotherium cottae				□											
Varanopus curvidactylus				□											
?Camunipes cassinisi				□											
***Batrachichnus* sp.**				□											

Fig. 4 – Stratigraphical distribution of the main groups of footprints for the Permo-Triassic sediments of northern Italy. Some clusters representing evolutionary levels can be distinguished. PERMIAN: A - Asselian; S - Sakmarian; A - Artinskian; K - Kungurian; K - Kazanian; T - Tatarian. TRIASSIC: I - Induan; O - Olenekian; A - Anisian; L - Ladinian; C - Carnian; N - Norian; R - Raethian. JURASSIC: H - Hettangian; S - Sinemurian; P - Pliensbachian.

The use of a sequence of footprint-based evolutionary units (LIUs) seems at the moment the most suitable system for timing and correlation of the continental sediments. A sequence of LIUs and the corresponding LIAs, still to be established after calibration, could be a powerful tool to apply to the problem of continental stratigraphy.

Taking into consideration the still general zoological meaning of the ichnogenera, it is obvious that the sensitivity of such units will be poor with respect to the units used in marine stratigraphy, since those biozones related to body fossils will carry greater confidence and will be considered more suitable where they can be used.

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CORRELATION OF THE UPPER PERMIAN SPOROMORPH COMPLEXES OF THE SOUTHERN ITALIAN ALPS WITH THE TATARIAN COMPLEXES OF THE STRATOTYPE REGION

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Key words – Palynology; Upper Permian; correlation; Southern Alps.

Abstract – The microfloras of the sub-Angarian province of the Tatarian stage of Late Permian age, in the type region, have been discussed and compared with Dzhulfian and Changxingian microfloras of the Southern Alps. A moderate taxonomic similarity, as well as different vegetation composition in the two regions, is indicated for the Urzhumsky Horizon (lower Tatarian). The occurrence of *Lunatisporites* and *Klausipollenites schaubergeri* pollen grains, and the smaller number of Costati, enhance the similarity between the Severodvinsky Horizon and the Val Gardena Sandstone 1st cycle microfloras. Based on the presence of *Protohaploxypinus microcorpus* in the Vjatsky Horizon and the very high degree of taxonomic similarity between the assemblages, a correlation is proposed between the microfloras of this late Tatarian horizon and those of the Val Gardena Sandstone and Bellerophon Formation 2nd and 3rd cycles. The appearance of *Lunatisporites noviaulensis*, contained in the topmost levels of the Vjatsky Horizon, and the absence of *Lueckisporites parvus* and *Tympanicysta* from the Tatarian type section, which are observed stratigraphically later in the Southern Alps, might suggest a very limited extent for the stratigraphic gap between the Tatarian and the Triassic, probably coinciding with only part of the Dorashamian (or Changxingian). The strong taxonomic similarity also indicates the rapid dispersion of the plants brought about by territorial continuity and by the similar paleoenvironmental and paleoclimatic conditions in the two regions. The two vegetations formed part of a single Euro-Cisuralian paleobotanical province.

Parole chiave – Palinologia; Permiano superiore; correlazione; Alpi Meridionali.

Riassunto –Vengono discusse e comparate le microflore di età Tatariana della regione tipo, con quelle della regione sudalpina di età Dzhulfiano e Changsingiano.
Si evidenzia una discreta similarità tassonomica tra le microflore delle due regioni in corrispondenza dell' orizzonte di Urzhumsky (Tatariano inferiore); la comparsa di granuli pollinici di *Lunatisporites* e la riduzione dei Costati aumenta la similarità tra le microflore dell'orizzonte di Severodvinsky e quelle di una parte del I ciclo delle Arenarie di Val Gardena; con la comparsa di *Protohaploxypinus microcorpus* nell'orizzonte di Vjatsky, e la alta similarità tassonomica che si instaura tra le associazioni si propone la correlazione tra le microflore di questo orizzonte (Tatariano superiore) con le microflore del I, II e III ciclo delle Arenarie di Val Gardena e della Formazione a Bellerophon. La assenza di *Lueckisporites parvus* e la non documentata presenza dell'evento a funghi (*Tympanicysta*), che nell'area sudalpina seguono stratigraficamente la comparsa di *Lunatisporites noviaulensis*, presente nel Tatariano terminale, potrebbero suggerire che l'entità della lacuna stratigrafica tra il Tatariano e il Trias sarebbe molto ridotta e verosimilmente corrispondente solo a parte del Dorashamiano (o Changsingiano).
La elevata similarità tassonomica evidenzia inoltre la rapida dispersione dei vegetali favorita dalla continuità territoriale e dalle simili condizioni paleoambientali e paleoclimatiche delle due regioni. Le due vegetazioni erano parte di una unica Provincia paleobotanica euro-cisuraliana.

INTRODUCTION

This study intends to contribute to the fervid debate within the Permian Working Group and the Subcommission of Permian Stratigraphy concerning the biostratigraphical and chronostratigraphical correlations of the Upper Permian.

Although understandably it is difficult to establish a correlation between the Upper Permian timescales of Europe and North America, because of the significant differences in sedimentary environments of their respective successions, ranging from continental to paralic in Western and Eastern Europe to deep marine in North America, we believe an attempt can be made to correlate the sporomorph assemblages of the Tatarian in the Volga–Urals type area, to those observed in the Val Gardena Sandstone (VGS) and the Bellerophon Formation in the Italian Alps, through the recognition of paleobotanical events such as the appearance of characteristic components.

The discussion that evolves herein takes into consideration a number of assumptions, namely:
• similar sedimentary environments in the two regions,

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from continental to paralic and shallow marine;
• position of the two regions roughly within the same latitude band (Ziegler, 1990) and the same paleoclimatic conditions;
• territorial continuity between Late Hercynian Europe and the East European Platform, ensuring rapid dispersion of the plants.

The Tatarian stage crops out extensively in the type area following the Kazanian, after a stratigraphic gap. Lithologically it is represented by sandstones, variegated clays, marls, dolostones and limestones, frequently crossed by lens-shaped bodies of fluvial sand, indicating prevalently continental, lagoonal and shallow-marine sedimentation. The fossils also belong to continental and lagoonal fauna such as pelecypods, conchostracans, ostracods and vertebrates.

In the type area the Tatarian stage is divided into three horizons: Urzhumsky, Severodvinsky and Vyatska. The most completely exposed portions of the latter horizon are to be found in the River Vetluga basin, where it has been further subdivided into three members (Borozdina & Olferiev, 1970; Olferiev, 1974).

The Tatarian begins with a geomagnetic polarity reversal (Illawarra-Khama Reversal) that continues from the Lower Permian (Burov & Esaulova, 1996). In the stratotype Cisuralian area, the Illawarra event distinguishes the Lower Tatarian (Urzhumskian) from the Upper Tatarian (Severodvinskian and Vjatskian) (Burov & Esaulova, 1996), and in Asiatic Russia the event can be traced up to the Tungusska and Pechora Basin, where it is possible to correlate Upper Permian sediments belonging to different paleobotanical environments.

SPOROMORPH ASSOCIATIONS OF THE TATARIAN IN THE TYPE AREA

Tatarian macro- and microfloras contained in the marine, transitional and continental sediments of the type region have been well identified by Kuntzel (1965), Gomankov & Meyen (1986), Esaulova (1995) and Koloda & Kanev (1996). Thus microfloral associations can be carefully compared on the basis of the widely published data. Biostratigraphical investigations conducted in the Volga–Urals region (Gusev *et al.*, 1993) based on brachiopods, foraminifers, ostracods, pelecypods, radiolarians, corals, bryozoans, macroflora, microflora and continental vertebrates have shown that the most significant changes within the major faunistic groups, along with flora renewal, took place between the Lower and Upper Tatarian. During this interval there was a transition from the floristic *Phylladoderma* complex to the *Tatarina* complex. This passage is believed to be a potentially correlative event between the different paleobotanical provinces of the Russian Federation because it is also marked by a paleomagnetic event.

The Urzhumsky Horizon (lower Tatarian) is characterised by miospore associations in which the striate and costate morphogroups are dominant and the disacciatriletes subdominant. Koloda & Kanev (1996) distinguish three taxa groups (important, common and rare) which, for greater clarity, are shown in table 1.

Roughly 50% of the Urzhumskian xerophytic flora species is present in the sporomorph associations of the Val Gardena Sandstone in the Italian Alps (Pittau, in Massari *et al.*, 1988, 1994; Pittau, in Cassinis, Cortesogno *et al.*,

Main taxa	Common taxa	Rare taxa
Alisporites splendens	*Scheuringipollenites ovatus*	*Limitisporites moersensis*
A. nuthallensis	*Protohaploxypinus jacobii*	*Gardenasporites heisseli*
Vesicaspora schemeli	*P. samoilovitchii*	*Vitreisporites signatus*
Protohaploxypinus amplus	*Ventralvittatina rotunda*	*Falcisporites zapfei*
P. perfectus		*Platysaccus papilionis*
Striatolebachiites spp.		*Gigantosporites* sp.
Vittatina costabilis		*Protohaploxypinus minor*
Vittatina subsaccata		*Striatopodocarpites antiquus*
Weylandites striatus		*Striatoabieites wilsonii*
		S. jansonii
		S. multistriatus
		S. richteri
		Lueckisporites virkkiae
		Representatives of *Weylandites*, *Vittatina* and *Fusacolpites*

Table 1

1999), with the exception of the representatives of *Fusacolpites*, *Striatolebachiites*, *Weylandites*, certain species of *Vittatina* and of *Protohaploxypinus*, specifically *P. amplus* and *P. perfectus*. The latter is a key taxon of the Kazanian microflora (Utting *et al.*, 1997) in the type region and extends up to the lower Tatarian. *P. amplus* has perhaps a broader stratigraphic range, Kungurian–lower Tatarian (Samoilovich, 1953; Varjuchina, 1971; Molin & Koloda, 1972; Foster, 1979; Koloda & Kanev, 1996), and its geographical distribution is also very wide. The two species have never been reported in the Val Gardena Sandstone or in the Bellerophon Formation. On the other hand, they have recently been recorded in the Tregiovo Formation (Pittau in Cassinis, Cortesogno *et al.*, 1999) of the southern Alpine basin, but are still not accurately dated – they may be Kazanian (Pittau in Cassinis, Cortesogno *et al.*, 1999) or Kungurian-Ufimian (Cassinis & Doubinger, 1991), but certainly older than the VGS (Cassinis *et al.*, 1999 and in press). Moreover, in the Urzhumsky microflora, significant characterising components of the younger flora of the southern Alpine and European Permian are missing.

The overlying Severodvinsky horizon is again characterised by xerophytic flora. Compared with the sporomorph associations in the Urzhumsky horizon, the following can be identified:

• a significant decrease in Costati and increase in Disaccites;
• a larger number of *Leuckisporites virkkiae* and, locally, of *Gigantosporites*;
• the appearance of species of *Lunatisporites* (= *Taeniaesporites*), *Scutasporites* and *Klausipollenites shaubergeri*.

The appearance of these pollen grains, especially *Lunatisporites* and *K. schaubergeri*, which are characteristic constituents of the Zechstein and southern Alpine Upper Permian sporomorph associations, is important because it enhances the similarity of the Cisuralian flora with the European flora for this stratigraphic interval.

The change in flora "from *Phylladoderma* to *Tatarina*" reported by the Russian authors appears therefore to translate, from a palynological point of view, into the points outlined above.

The palynological associations in the Vjatsky horizon testify once again to a xerophytic flora with dominant disaccates. *Vitreisporites*, represented by various species, is an abundant component in the associations together with *Leuckisporites virkkiae* and different species of *Lunatisporites* (*Taeniasporites labdacus*, *Taeniaesporites* sp.). *Protohaploxypinus microcorpus* appears along with another, albeit rare, element (*incertae sedis*) reported in the Southern Alps, *Inaperturopollenites nebulosus*. The presence of *Lunatisporites noviaulensis* (*noviaulensis/pellucidus* complex) has been reported by Foster & Jones (1994) in the upper Tatarian strata.

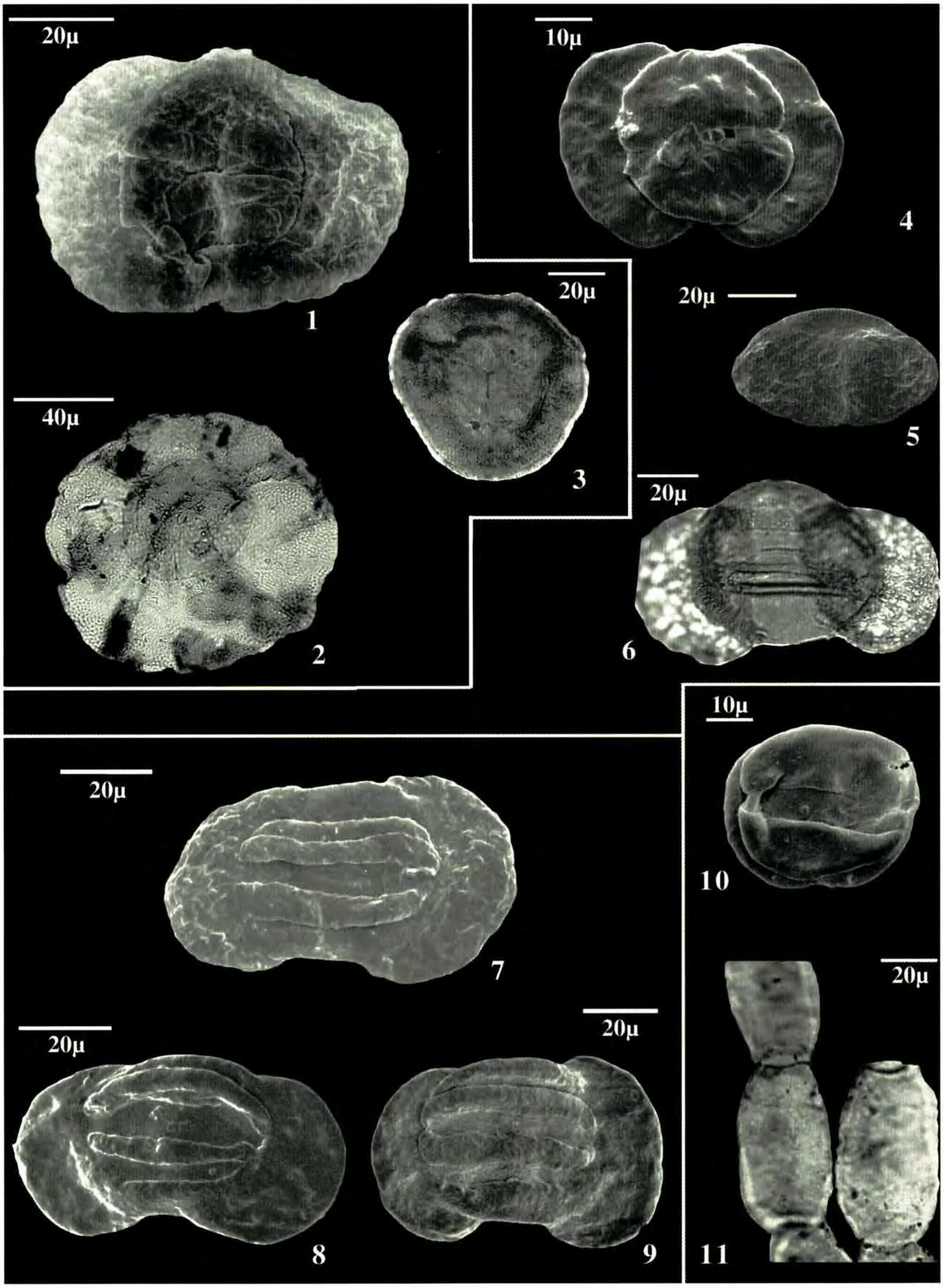
Contextually neither *Tympanicysta* nor *Leuckisporites parvus* have been reported in this stratigraphic interval in the Tatarian type section.

In the subsurface successions of the Moscow syneclise, European Russian Platform, the sporomorph assemblages (Lozovsky & Yaroshenko, 1994) attributed to the Vjatsky horizon are strongly hygrophytic. In the upper member of the horizon (Molomsky), in addition to about 30% of spores and a significant amount of *Ephedripites* (20%), the sporadic presence of *Tympanicysta stoschiana* is also reported. The presence of *Tympanicysta* has also been recorded from the latest Permian to lowermost Triassic basal Vetlugian deposits of Nedubrovo in the Vologda region of the Russian Platform (Krassilov *et al.*, 1999). Through these findings it is possible to define better the upper limit of the Tatarian and the consequent stratigraphic gap, which seems to be very small. Outside this geological district, in the Pechora Basin, *Tympanicysta* has been observed in the Induan (Foster & Yaroshenko, in Foster & Jones, 1994).

COMPARISON WITH THE VAL GARDENA SANDSTONE AND THE BELLEROPHON FORMATION IN THE ITALIAN ALPS

The quantitative and compositional evolution of the microflora in the southern Alpine basin has been addressed by the author in several papers (Conti *et al.*, 1986; Massari *et al.*, 1988, 1994; Conti *et al.*, 1997; Cassinis *et al.*, 1999), in an endeavour to recognise microfloristic events useful for establishing interbasin correlations, as distinguished from useful events for establishing international correlations (Plate 1).

The restricted stratigraphical range of *Protohaploxypinus microcorpus* together with its wide geographical distribution (see Pittau in Massari *et al.*, 1988) make this a very important taxon for correlation. Furthermore, if its FAD is accurately determined in a continuous succession and possibly in the stratotype of a chronostratigraphic unit, its correlating potential and reliability is enhanced considerably. Pittau (Conti *et al.*, 1986; in Massari *et al.*, 1988) attempted to correlate the Val Gardena Sandstone and Bellerophon Formations with the marine stages using key taxa such as *Endosporites hexareticulatus* (= *Playfordiaspora crenulata*, *sensu* Foster, 1979), *Protohaploxypinus microcorpus* and *Densoisporites playfordi*, to correlate the Val Gardena Sandstone (VGS) with the *P. microcorpus* palynozone (including *E. reticulatus* and *P. microcorpus*) in Australia (Foster, 1982) with unit 4 of the Chhidru Formation. At the time a poorly defined Dzhulfian age was assigned. Now that there appears to be general consensus (Foster & Jones, 1994) in attributing unit 4 of the Chhir-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dru Formation to the late Dzhulfian and part of the Dorashamian, this assignment can be extended to the VGS and to part of the Bellerophon Formation.

P. microcorpus appears in the lower portion of the VGS and its FO can be observed in the section with the longest stratigraphical interval relative to the fossil record, namely 20 m above the base of the formation in the Blätterback section (Massari *et al.*, 1988, 1994); it has never been observed in older successions in the Southern Alps, such as the Tregiovo Formation (Pittau in Cassinis *et al.*, 1999; Cassinis & Doubinger, 1991; Barth & Mohr, 1994).

Bearing in mind that this species occurs within the continuous sedimentary succession deposited in continental–paralic environments in the Tatarian type section, and that it has not yet been reported in older strata, the base of the Vjatsky horizon could be taken as the spike (?FAD) for this first appearance, or alternatively, level 2 of the Blätterback Formation, situated 20 m above the base of the VGS Formation, in the first sedimentary cycle.

CORRELATIONS BETWEEN TATARIAN AND SOUTHERN ITALIAN ALPINE SPOROMORPH ASSEMBLAGES

In the light of the above observations, the sporomorph assemblages observed in the Urzhumsky horizon are believed to be older than those of the VGS because they precede the appearance of *Lunatisporites* and *Klausipollenites schaubergeri*. On the other hand, the presence of characterising Kazanian flora species like *Protohaploxypinus amplus* and *P. perfectus*, combined with abundant costates pollen grains, indicate that they are again older than the southern Alpine and Zechstein floras. In conclusion it may be assumed that the Urzhumsky microfloras are older than the VGS and perhaps younger than the Tregiovo microfloras, although this hypothesis needs to be corroborated by further observations.

In the Severodvinsky horizon the palynoflora and plant communities exhibit similarities in the two regions: the floras concerned are meso-xerophytic containing few hygrophytic species, reflecting the extremely low occurrence of spores. The similarity in this stratigraphic interval is accentuated by the presence of *Lunatisporites* (=*Taeniaesporites*), *Klausipollenites schaubergeri* and *Scutasporites*, which, with the increased number of *Luekisporites virkkiae*, make the vegetational composition of the Cisuralian flora more similar to the VGS.

Taking the appearance of *P. microcorpus* as a basis for the correlation, the Severodvinsky microfloras can be correlated with the lower part of the Val Gardena Sandstone, coinciding with part of the first sedimentary cycle (Massari *et al.*, 1994).

The occurrence of *Protohaploxypinus microcorpus* in the Vjatska horizon of the Tatarian type section allows correlation with this portion with the Val Gardena Sandstone, and with part of the Bellerophon Fm.

In fact, the absence of *Lueckisporites parvus* and secondly of *Tympanicysta* can be considered an element of correlation between the two areas, indicating that the Vjatska horizon in the type section does not comprise the entire VGS and the Bellerophon period of deposition, but only a portion of the first cycle, all of the second and part of the third: that is, up to the appearance of *L. parvus* and the fungal event.

The presence of *Lunatisporites noviaulensis*, another important species because of its vast geographical distribution and the fact that its occurrence is stratigraphically confined to the uppermost Permian, is reported by Foster & Jones (1994) in the upper layers of the Vjatska horizon. In the Southern Alps this species occurs prior to the appearance of *L. parvus* and *Tympanicysta*, so again its presence does not contradict the existence of a stratigraphic gap between the Tatarian in the type section and the Triassic.

The comparison suggests a small stratigraphic gap, corresponding to the IV and V sedimentary cycles of the Southern Alpine basin (Massari *et al.*, 1994), which in chronostratigraphical terms should coincide with part of the Changxingian and may be even shorter in other parts of the Moscow syneclise.

Thus, on the basis of the assumptions discussed above and the criteria outlined here, a correlation is suggested between the upper Tatarian and the Val Gardena Sandstone and Bellerophon Formations of the Southern Alps, schematically represented in Fig. 1, and updated in the chronostratigraphical references to the marine stages.

A final consideration concerns the position in temporal terms of the two southern Alpine formations with respect to the paleomagnetic reversal event which was long sought in the Dolomitic basin. According to this correlation hypothesis, the Illawarra Event could be assigned to between the VGS and the Tregiovo Formation.

Plate 1

1. *Protohaploxypinus microcorpus* (Schaarschmidt) Clarke
2. *Endosporites hexareticulatus* Klaus
3. *Densoisporites playfordii* (Balme) Dettman
4. *Lueckisporites virkkiae* Potonié & Klaus
5. *Klausipollenites schaubergeri* Klaus
6. *Lunatisporites labdacus* (Klaus) Pittau
7. *Lunatisporites pellucidus* (Gobin) Helby
8. *Lunatisporites noviaulensis* (Leschik) Scheuring
9. *Lunatisporites noviaulensis* (Leschik) Scheuring
10. *Lueckisporites parvus* Klaus
11. *Tympanicysta stoschiana* Balme

CONCLUSIONS

Comparison of the sporomorph assemblages in the Val Gardena Sandstones and Bellerophon Formations, described in detail by Klaus (1963), Pittau in Conti *et al.* (1986, 1997), Massari *et al.* (1988, 1994), and Cassinis *et al.* (1999), with those of the Tatarian type section described by Koloda & Kanev (1996), has allowed us to advance correlation hypotheses that will certainly have to be substantiated by other data, but that can represent a starting point for further investigations and discussion.

The starting assumption is the territorial contiguity between the two regions and the similar paleoenvironmental and climatic conditions established at the end of the Paleozoic, in addition to the similarity of the microfloras, more so in the Severodvinsky and Vjatsky horizons.

The key features of the correlation are the appearance of *Lunatisporites* and *Klausipollenites schaubergeri* of *Protohaploxypinus microcorpus* and *Lunatisporites noviaulensis*. The absence of *Lueckisporites parvus* and the fungal event are considered correlational elements that rule out the presence of part of the Changxingian in the Tatarian type section, although these two points warrant further investigation.

The correlation proposed here suggests a correspondence of microfloras of the Severodvinsky and Vjatsky horizons with those of the VGS and Bellerophon Formation of the I, II and III cycles (Massari, 1994). The gap claimed to exist between the Tartarian of the type section and the Triassic is believed to coincide with the IV and V sedimentary cycles of the southern Alpine succession, specifically to part of the Changxingian, but there are elements in favour of the existence of a smaller gap in other parts of the Moscow basin (Lozovsky & Yaroshenko, 1994; Krassilov *et al.*, 1999).

Correspondence with the marine stages (Late Dzhulfian and Changxingian) is based on correlation with unit 4 of the Chhidru Formation and on the occurrence of foraminifers in the upper part of the Bellerophon Fm., as pointed out previously (Massari *et al.*, 1988; Conti *et al.*, 1995).

As the scientific debate on correlations of marine and continental scales is always lively and topical in the Working Groups, it is hoped that the subject dealt with in this paper will stimulate further investigation and speculation.

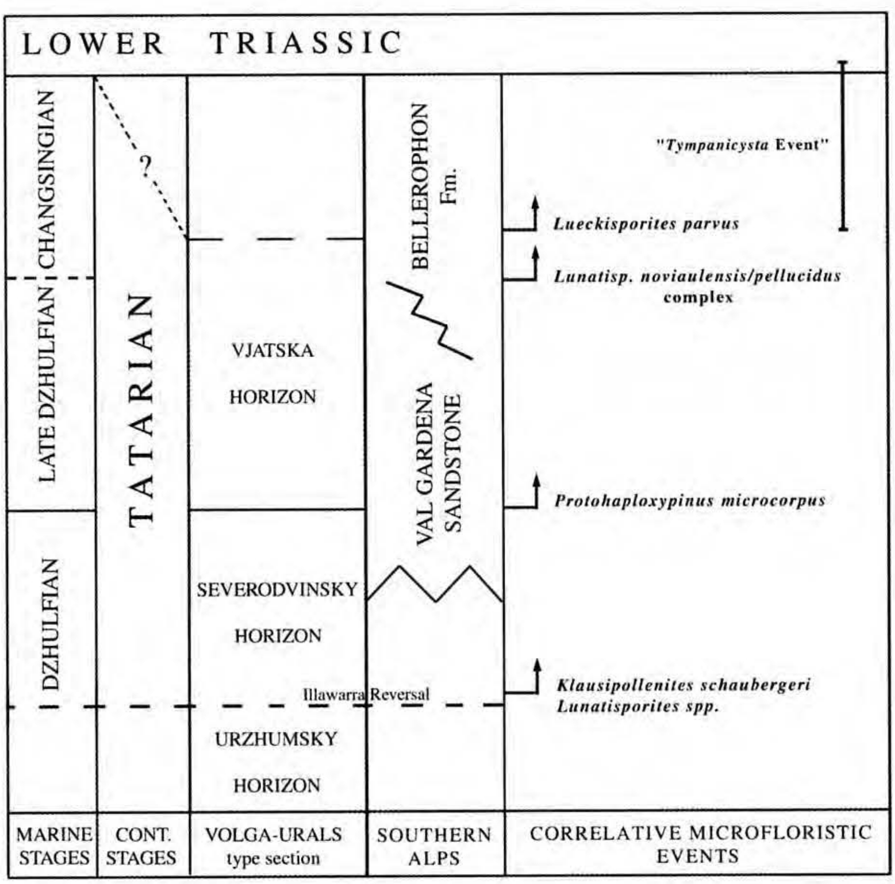

Fig. 1 – Correlation of the Tatarian with Southern Alps continental to shallow marine successions and the marine stages.

List of the sporomorphs cited in the text

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Alisporites splendens (Leschik) Foster, 1979
Densoisporites playfordi (Balme) Dettmann, 1963
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Ephedripites sp.
Falcisporites zapfei Klaus, 1963
Fusacolpites sp.
Gardenasporites heisseli Klaus, 1963
Gigantosporites sp.
Inaperturopollenites nebulosus Balme, 1970
Klausipollenites schaubergeri (Potonié et Klaus) Jansonius, 1962
Limitisporites moersensis (Grebe) Klaus, 1963
Lueckisporites parvus Klaus, 1963
Lueckisporites virkkiae Potonié et Klaus, 1954
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Scheuringipollenites ovatus (Balme et Hennelly) Foster, 1975
Scutasporites sp.
Striatoabieites jansonii (Klaus) Hart, 1964
Striatoabieites multistriatus (Balme et Hennelly) Hart, 1964
Striatoabieites richteri (Klaus) Hart, 1964
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Striatolebachiites spp.
Striatopodocarpites antiquus (Leschik) Potonié, 1958
Tympanicysta stoschiana Balme, 1979
Ventralvittatina sp.
Vitreisporites signatus Leschik, 1957
Vittatina costabilis Wilson, 1962
Vittatina sp.
Vittatina subsaccata Samoilovich ex Wilson, 1962
Weylandites sp.
Weylandites striatus (Luber) Utting, 1994

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PERMIAN FLORAS OF THE SOUTHERN ALPS

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Key words – Megaflora; Cuticles; Permian; Southern Alps; Italy.

Abstract – Permian plant megafossils are long since known from various parts of the Southern Alps. These fossils were traditionally mainly used for age estimates of continental formations. However, floras with well preserved cuticles on the basis of which natural species, genera and families could be recognised, also contributed towards a much better understanding of late Palaeozoic gymnosperms.
A short introduction is followed by an updated synthesis of the plant fossils discovered until now in the region. The floras from the Collio Formation in the upper Val Trompia (Brescia) and the Tregiovo Formation in the upper Val di Non (Trento-Bolzano), and from the overlying Upper Permian Val Gardena red beds, east of the Adige Valley are discussed.

Parole chiave – Macroflora; Cuticole; Permiano; Alpi Meridionali; Italia.

Riassunto – La presenza di piante permiane è da lungo tempo nota in vari settori della Alpi Meridionali. Tradizionalmente, questi fossili furono essenzialmente utilizzati per stimare l'età delle formazioni continentali. Tuttavia, flore con ben preservate cuticole in base alle quali specie naturali, generi e famiglie potevano essere riconosciute, contribuirono pure ad una assai migliore conoscenza delle gimnosperme tardo-paleozoiche.
Dopo una breve introduzione segue una aggiornata sintesi delle piante fossili finora rinvenute nell'area considerata. In particolare, vengono discusse le flore della Formazione di Collio in alta Val Trompia (Brescia), della Formazione di Tregiovo nell'alta Val di Non (Trento-Bolzano), nonché quelle sovrastanti, tardo-permiane, delle Arenarie rosse di Val Gardena, presenti ad est della Val d'Adige.

THE PRE-VERRUCANO LOMBARDO / PRE-VAL GARDENA FLORAS

Plant fossils from the Collio and Tregiovo formations underlying the Verrucano Lombardo and Val Gardena formations were first described more than a century ago (*e.g.*, Geinitz, 1869). The age of these floras has been discussed repeatedly and age assessments varied from Late Carboniferous to "Middle" Permian. The most recent account on these megafloras was published by Remy & Remy (1978) who dated these floras as respectively "middle to higher Saxonian" and "higher Saxonian". In recent years a third flora was described from the Ponteranica Conglomerate in the Orobic Alps (Kerp *et al.*, 1996). None of these localities has yielded a really rich and diverse flora. Conifers are the dominant elements in all these three floras. However, unlike the situation in the Val Gardena flora (see below), precise identifications are strongly hampered by the lack of cuticles, which are of essential importance for the recognition of late Palaeozoic conifer taxa; morphologically very similar species of conifers can often only be distinguished on the basis of their epidermal characteristics.

The Collio flora

The flora of the Collio Formation in the upper Val Trompia consists according to Remy & Remy (1978) of six taxa and a number of unidentifiable conifer (and ginkgophyte?) remains: *Sphenopteris suessii* (Pl. I, Figs 6, 7), *S. kukukiana* (Pl. I, Fig.12), *S. patens*, *"Sphenopteris"* cf. *interrupte-pinnata*, *Hermitia* (al. *Walchia*) *geinitzii* (Pl. I, Figs 3, 4) and *Walchiostrobus* sp.

Two of these taxa are only known from the Collio Formation (*S. suessii*, *H. geinitzii*) whereas *S. kukukiana* and *S. patens* were originally described from the German Zechstein. The single specimen illustrated as *"Sphenopteris"* cf. *interrupte-pinnata* (Geinitz, 1869; Remy & Remy, 1978) is very fragmentary preserved; this taxon was originally described from the Permian Copper Sandstone of the Ural Mountains (Russia). None of the

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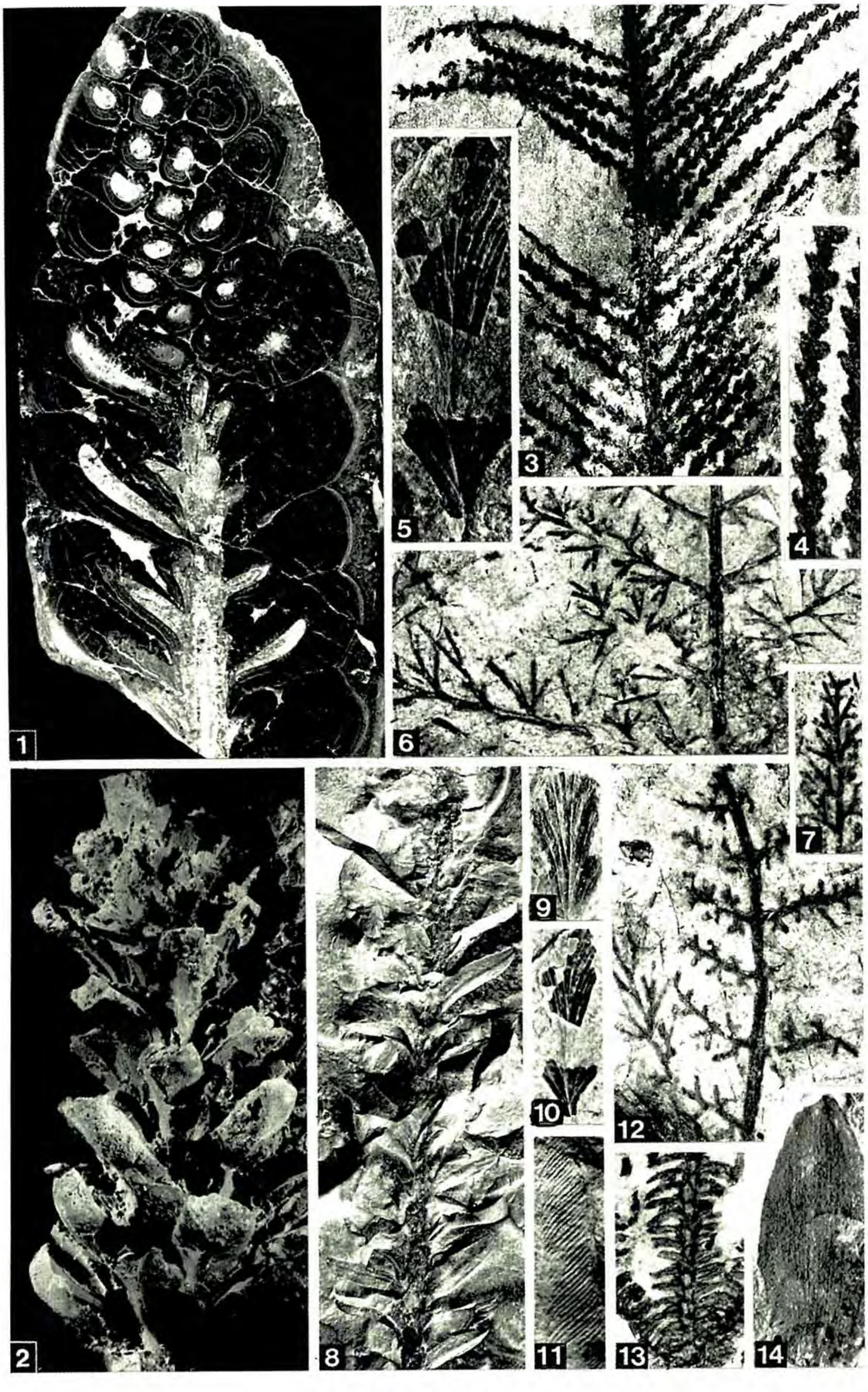
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sphenopterids is known from the classical Rotliegend. *Walchiostrobus* is a conifer cone originally described by Sordelli (1896) as *Curionia triumpilina* and assigned to *Walchiostrobus* by Florin (1940). Conifer foliage types are reminiscent of *Hermitia* (al. *Walchia*) *germanica* (Pl. I, 13), *H. gallica*, *Culmitzschia* (al. *Lebachia*) *laxifolia* and *Otovicia* (al. *Walchia*) *hypnoides*. These foliage types are normally found in the typical Rotliegend. However, forms similar (or identical) to the latter species are also known from the lower Upper Permian of the Arabian Peninsula. As already mentioned, these taxonomic assignments are only tentative because no cuticles are known.

The Tregiovo flora

According to Remy & Remy (1978), the flora of the Tregiovo Formation in the upper Val di Non, between the Bolzano and Trento provinces includes, apart from various conifer remains, *Lodevia* (al. *Callipteris*) cf. *nicklesii* (Pl. I, Figs 5, 9, 10) and *Lesleya* (al. *Taeniopteris*) *eckardtii* (Pl. I, Fig. 11). Of the first species, that was originally described from the Lower Permian of Lodève (Southern France), only a few incomplete pinnules have been found, whereas the second record refers to a single small fragment only; this latter taxon is a typical Zechstein element. More common are conifer remains. A variety of forms, ranging from very small-leaved types attributed to *Otovicia hypnoides* to large-leaved ones attributed to *Ullmannia frumentaria* (Pl. I, Fig. 8) has been recorded. Although also here identifications are still tentative due to the lack of cuticles, morphological similarities with these essentially Early respectively Late Permian species are striking. Other conifer taxa from the Tregiovo Formation have been compared with *Walchia piniformis*, "*Walchia*" *stricta* and *Quadrocladus*. Isolated large, broad, apparently non-decurrent leaves showing prominent parallel lines (stomatal rows?), previously assigned to *Lepeophyllum* or *Culmitzschia* (Remy & Remy, 1978) might well represent isolated *Ortiseia leonardii* leaves, well-known from the Val Gardena Formation (see below).

The Ponteranica flora

The Ponteranica Conglomerate near Gerola Alta in the western Orobic Alps has yielded a conifer species that was described as *Cassinisia orobica* (Kerp *et al.*, 1996). This conifer has very robust axes and with up to 2.5 cm long leaves. Most remarkable is their three-dimensional preservation in stromatolites (Pl. I, Fig. 1; Freytet *et al.*, 1996). Naturally formed casts are exact lifetime replicas of these Permian conifers (Pl. I, Fig. 2).

The age of the Collio and Tregiovo Formations

The age of the Collio and the presumably time-equivalent Tregiovo Formation has been discussed for more than a century. Initially these discussions were based exclusively on the megafloras, in later years palynological data have become available. Although both megafloras are rather poor in species and more material still should be studied, they both seem to include taxa which are so far only from the typical Zechstein. However, it should be noted that the ranges of taxa are still imperfectly known as most of the western and middle European Permian is either missing or unfossiliferous. Therefore, stratigraphical ranges of megaplant taxa in this particular time interval are very still poorly known. The most recent palynological data suggest a late Artinskian -?early Ufimian, respectively a Kungurian -Ufimian age for the Collio and the slightly younger Tregiovo Formation (Cassinis & Doubinger, 1991a, b; Barth & Mohr, 1994), which is in accordance with the ages estimated by Remy & Remy (1978) on the basis of the macrofloras. The palynological assemblages from the Tregiovo Formation seem to be younger than those from the marine Amanda Formation recognized in boreholes in the northern Adriatic Sea and probably corresponding to the *Neoschwagerina* (fusulinid) Zone (Sartorio & Rozza, 1991).

THE VAL GARDENA FLORA

Notably in the western Dolomites and in the Vicentinian Alps, grey interbeds in the Val Gardena Formation are sometimes rich in plant fragments of variable size and preservation. Although regularly reported since the second part of the 19th century, larger well-preserved plant fossils are rare. According to early identifications, the fossil as-

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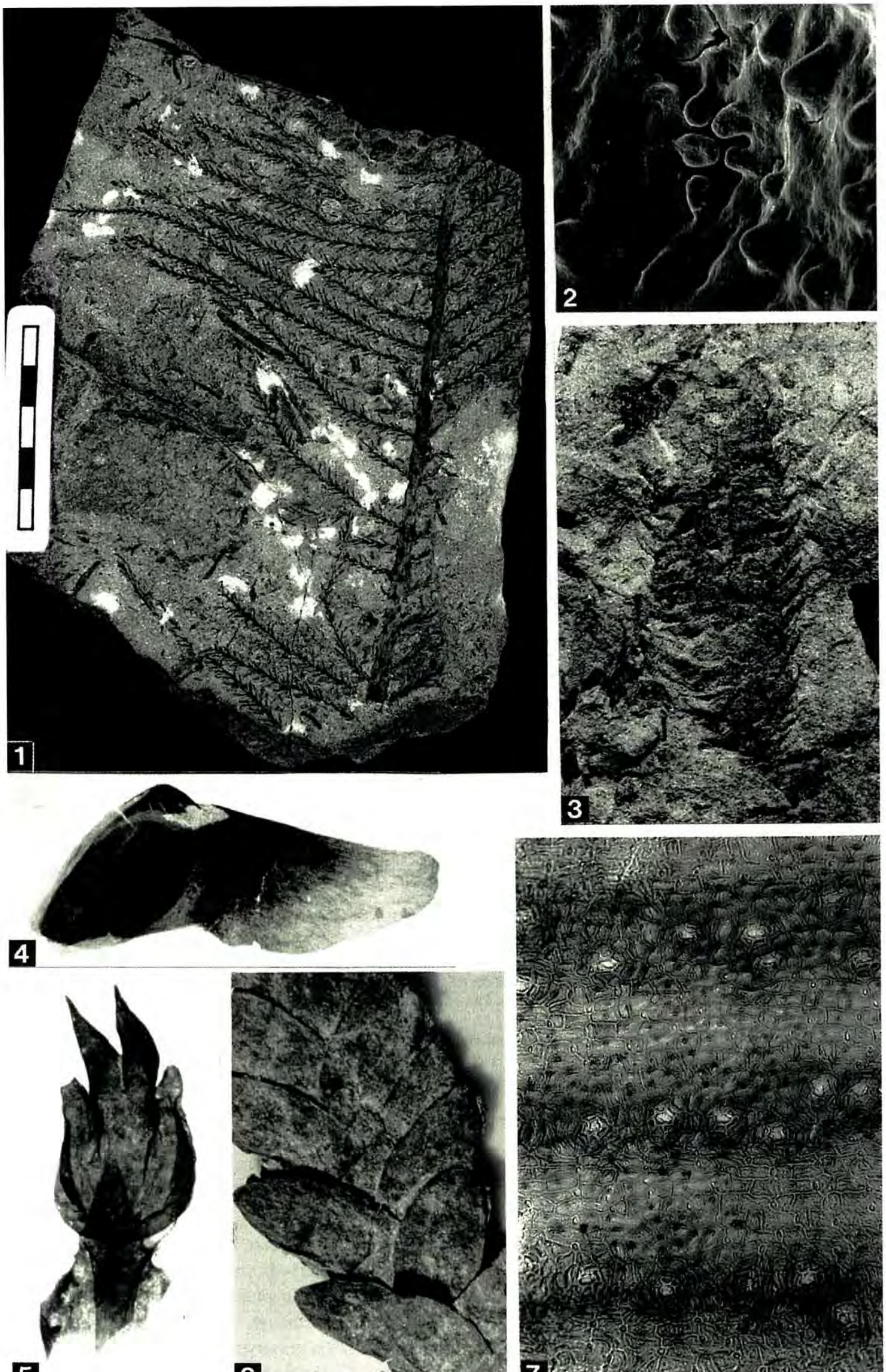
Plate I

1. *Cassinisia orobica*, polished section of a stromatolite showing a conifer in the centre. Pizzo Tre Signori, western Orobic Alps; Ponteranica Conglomerate. x 1.
2. *Cassinisia orobica*, natural siltstone cast. Pizzo Tre Signori, western Orobic Alps; Ponteranica Conglomerate. x 1.8.
3. *Hermitia* (al. *Walchia*) *geinitzii*. Monte Colombine; Collio Fm. x 1.
4. Detail of Fig. 3. x 2.
5. *Lodevia* (al. *Callipteris*) *nicklesii*. Exact locality unknown, between Trento and Bolzano; Tregiovo Fm. x 2.
6. *Sphenopteris suessii*. Monte Colombine; Collio Fm. x 1.
7. Idem. x 1.
8. *Ullmannia frumentaria*. ?Tregiovo; Tregiovo Fm. x 1.
9. *Lodevia* (al. *Callipteris*) *nicklesii*. Exact locality unknown, between Trento and Bolzano; Tregiovo Fm. x 1.
10. Same specimen as Fig. 5. x 1.
11. *Leslya* (al. *Taeniopteris*) *eckardtii*. ?Tregiovo; Tregiovo Fm. x 1.
12. *Sphenopteris kukukiana*. Monte Colombine; Collio Fm. x 1.
13. *Hermitia* (al. *Walchia*) cf. *germanica*. Monte Dasdana; Collio Fm. x 1.
14. Isolated conifer leaf, showing similarities with *Ortiseia leonardii* leaves. Exact locality unknown, between Trento and Bolzano; Tregiovo Fm. x 2.



semblages seemed to include coniferous foliage of both the Late Carboniferous-Early Permian genus *Walchia* and the Late Permian and Triassic genera *Ullmannia* and *Voltzia*. Based on foliage with preserved cuticle from the Western Dolomites, the only well-described element was the coniferous form-genus *Ortiseia* (Florin, 1964).

In contrast to the rare megafossils, bulk-macerated samples of siltstones that are rich in plant debris, may yield cuticle residues from which large numbers of small (usually <1 cm) but excellently preserved plant remains can be isolated (compare Pl. III) These cuticular fragments are dominated by conifer remains and include both vegetative and reproductive structures, such as isolated leaves, minute twigs, ovuliferous dwarf shoots, seeds and microsporophylls. In addition, one may note remains of pteridosperms. This material has been the basis for a detailed taxonomic study of the principal elements of the Val Gardena flora.

In gymnosperms the epidermal structure, as reflected in the plant cuticles, may vary according to family, genus and species, and hence proves to be of great potential in natural classification. Like in extant taxa, reproductive organs provide the most successful characters for delimiting natural genera and families. In absence of organic connection, cuticle analysis provides a reliable alternative for the necessary correlation of vegetative foliage and reproductive organs.

Through detailed comparative cuticle analysis it was possible to recognize three distinctive species within *Ortiseia* (Clement-Westerhof, 1984), viz. *O. leonardii* (Pl. II, Fig. 7) *O. visscheri* (Pl. II, Fig. 6) and *O. jonkeri* (Pl. II, Figs 1, 3; Pl. III, Figs 2, 5, 6, 8). The species could be described in terms of characteristics of vegetative structures as well as the organization of ovuliferous and polliniferous organs. The form-genus *Ortiseia* thus became "promoted" (Visscher *et al.*, 1986) to the status of a natural genus. It could be demonstrated that the ovules are characterized by the presence of a pollen/archegonial chamber, indicating the existence of zoidogamy among early conifers. The polliniferous organs yield distinctive pollen grains that correspond to the palynological form-genus *Nuskoisporites*. Size and wall organization indicates that the pollen grains represent prepollen, *i.e.* microspores of extinct zoidogamous seed plants that have not yet developed the capacity to produce a distal haustorial pollen tube (Poort *et al.*, 1997).The results of the work on *Ortiseia* formed a basic step towards a natural concept for the family Walchiaceae (Clement-Westerhof, 1984, 1988; Kerp *et al.*, 1990).

In the Val Gardena Formation, natural coniferous genera other than *Ortiseia*, include *Majonica* (*M. alpina*; Pl. II, Figs 4, 5; Pl. III, Fig. 3), *Dolomitia* (*D. cittertiae*) and *Pseudovoltzia* (*P. liebeana*; Pl. III, Fig. 7, *P. sjerpii*). These genera have been reconstructed with the aid of cuticle analysis on the basis of rich vegetatitive and fertile plant fragments (Clement-Westerhof, 1987). Particularly because of a common characteristic organization of ovuliferous dwarf-shoots, the three genera have been included in a separate family, the Majonicaceae. Polliniferous organs of *Majonica* yield pollen grains that belong to the palynological form-genus *Lueckisporites*.

As far as the coniferous remains are concerned, the detailed reconstruction of natural genera demonstrates that the composition of the flora of the Val Gardena Formation is different from what is assumed in earlier literature . The dominant conifers are *Ortiseia*, *Majonica*, *Dolomitia*, and *Pseudovoltzia*, rather than *Walchia*, *Ullmannia*, and *Voltzia*.

The most prominent pteridosperm fragments from the Val Gardena Formation correspond to the species formerly known from the Zechstein Basin of NW-Europe as *Callipteris martinsii*. On the basis of a detailed analysis of foliage and ovuliferous organs, this species is now included in the natural genus *Peltaspermum* (*P. martinsii*; Pl. II, Fig. 2; Pl. III. Figs 1, 4) of the family Peltaspermaceae (Poort & Kerp, 1990).

It should be realized that the above-mentioned taxa are only part of the flora of the Val Gardena Formation.The bulk-macerated plant material has yielded a variety of fertile fragments that have not yet been identified. Also palynological assemblages (*e.g.* Klaus, 1963; Massari *et al.*, 1994) testify to a much more diverse flora. Yet, because of their natural status, the plant taxa so far described from the Val Gardena Formation, are important links in the reconstruction of gymnosperm evolution.

It is now generally agreed that the age of the Val Gardena flora is Late Permian. All recognized species become extinct at or close to the Permian-Triassic junction. This extinction illustrates the profound effect of the Permian-Triassic biotic crisis on gymnosperm diversity in the Late Palaeozoic Euramerican floral realm (Visscher & Brugman, 1988; Visscher *et al.*, 1996; Poort *et al.*, 1997).

Plate II

1. *Ortiseia jonkeri*, lateral shoot system. Cortiana, Val Gardena Fm. x 0,65.
2. *Peltaspermum martinsii*, SEM outer leaf surface showing papillae and one stoma (compare Pl. III, 4). Butterloch, Val Gardena Fm. x 700.
3. *Ortiseia jonkeri*, polliniferous cone. Recoaro, Val Gardena Fm. x 2,2.
4. *Majonica alpina*, seed. Butterloch, Val Gardena Fm. x 4,4.
5. *Majonica alpina*, ovuliferous dwarf shoot, abaxial view. Butterloch, Val Gardena Fm. x 5,2.
6. *Ortiseia visscheri*, shoot ultimate order. Taubenleck, Val Gardena Fm. x 3,5.
7. *Ortiseia leonardii*, cuticle showing stomatal rows. Butterloch, Val Gardena Fm. x 130.

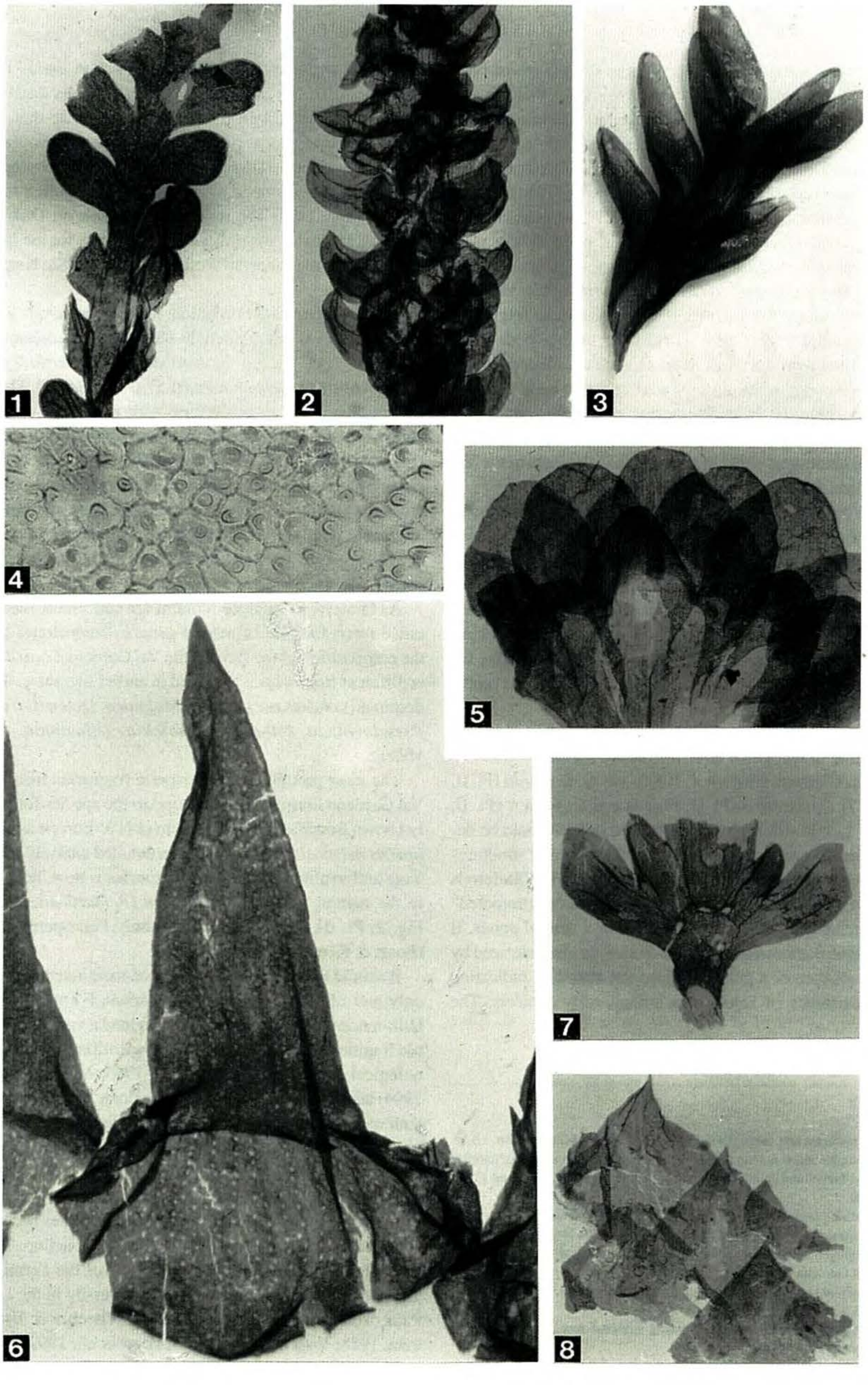

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Plate III

1. *Peltaspermum martinsii*, foliage. Butterloch, Val Gardena Fm. x 6.
2. *Ortiseia jonkeri*, shoot ultimate order. Taubenleck, Val Gardena Fm. x 10.
3. *Majonica alpina*, shoot. Butterloch, Val Gardena Fm. x 3.
4. *Peltaspermum martinsii*, leaf cuticle. Butterloch, Val Gardena Fm. x 270.
5. *Ortiseia jonkeri*, ovuliferous dwarf shoot. Butterloch, Val Gardena Fm. x 9.
6. *Ortiseia jonkeri*, leaf main axis. Butterloch, Val Gardena Fm. x 10.
7. *Pseudovoltzia liebeana*, ovuliferous dwarf shoot, abaxial view. Butterloch. Val Gardena Fm. x 4.
8. *Ortiseia jonkeri*, microsporophyls. Butterloch, Val Gardena Fm. x 10.

THE BELLEROPHON-WERFEN BOUNDARY IN THE WESTERN DOLOMITES (ITALY) - PETROGRAPHICAL STUDIES AND A NEW INTERPRETATION

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Key words – Bellerophon-Werfen Boundary; Depositional Evolution; Dolomites; Italy.

Abstract – Analysis of some detailed sections at the transition between the Bellerophon and Werfen Fms. in the western Dolomites has made possible a new interpretation of the formational boundary. In particular the Tesero (Fiemme Valley) and Seres (Badia Valley) sections were chosen, being representative of proximal and distal palaeogeographical conditions respectively. The boundary is located at the base of the thick bed containing the first oolitic levels and overlying a thin black organic-rich shaly layer. The criteria enabling us to reach this new interpretation are:
– a progressive and continuous evolution of the petrographical and sedimentary characteristics of the microfacies in the thick bed;
– a sharp morphological break on the outcrops.
The different phases of the depositional evolution in the studied interval can be correlated on a regional scale; these have already been recorded but their boundaries are in part reinterpreted here both in terms of location and meaning.

Parole chiave – Limite Bellerophon-Werfen; Evoluzione deposizionale; Dolomiti; Italia.

Riassunto – Lo studio di dettaglio di alcune sezioni al passaggio tra le formazioni a Bellerophon e di Werfen nelle Dolomiti occidentali ha permesso di proporre una nuova interpretazione per il limite formazionale. In particolare sono state scelte le sezioni di Tesero (Val di Fiemme) e di Seres (Val Badia), rispettivamente rappresentative di condizioni paleogeografiche prossimali e distali. Il limite viene ubicato alla base del banco che contiene i primi livelli oolitici, in corrispondenza di un sottile livello pelitico nero e bituminoso. I criteri che permettono di arrivare a questa nuova interpretazione sono i seguenti:
– una progressiva e continua evoluzione dei caratteri petrografici e sedimentari all'interno del banco;
– un netto stacco morfologico in affioramento.
Le differenti fasi dell'evoluzione sedimentaria nell'intervallo esaminato sono correlabili a scala regionale; queste fasi sono già note in letteratura, ma i loro limiti vengono qui in parte reinterpretati, sia per posizione che per significato.

INTRODUCTION

This paper deals with a new interpretation of the Bellerophon-Werfen boundary through the study of two representative sections in the Dolomites area: the Tesero section (Fiemme Valley) and the Seres section (Badia Valley) (Beretta, 1999) (Fig. 1).

This boundary, in previous studies, was generally discussed together with the Permian-Triassic chronostratigraphical boundary; the two boundaries are in fact very close and for many years were even considered to coincide.

For the Dolomites area, before the fundamental paper by Bosellini (1964), this boundary was not defined so clearly (Leonardi, 1935) because of the presence of transitional beds between two distinct lithological complexes (Upper Permian Bellerophon Fm. and Triassic "Werfenian"). Accordi (1958) located the boundary at the base of the *Claraia* marls in the Bletterbach-Butterloch section, lying above limestones and marly limestones; this author used the terms "fiemmazza" and "badiota" facies to label the two main lithological types of the Bellerophon Fm. The *Claraia* beds were also considered the base of the Werfen also in the Trento area (Venzo, 1955, 1962; Panizza, 1963), whereas the underlying thick oolitic limestones (about 20m) were assigned to the Upper Permian Bellerophon Fm.

Bosellini (1964) first recorded a mainly oolitic interval (from 20cm to 7-8m) in several sections of the Dolomites. This interval, initially defined "transitional" and then as the "Tesero Oolitic Horizon" (TOH), was considered the base of the Triassic Werfen Fm. The TOH always begins with a 15cm to 1m thick oolitic bed; its absolute base, in some sections, consists of facies representing the initial

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phase of oolite production, *i.e.* bio- and intraclastic limestones followed by "coated grain" limestones and then by the first small oolites.

For Assereto *et al.* (1973) the boundary was defined by a disconformity (or unconformity) at the base of the TOH. A hiatus was in fact recognised, based on biostratigraphical data: the *Comelicania* beds, present at the very top of the Bellerophon Fm. in the eastern Southern Alps, were correlated with the *Comelicania* beds of lower Dorashamian age found in the Transcaucasia, whereas the oolitic beds were considered of Lower Triassic age. Details on lithology at the boundary were not shown, but the Tesero section was included in the area where the TOH directly overlies the "fiemmazza" facies of the Bellerophon Fm.

Farabegoli & Viel (1982) considered the boundary as transgressive in the Trento-Val Sugana area. The TOH shows at the base, above bioturbated dolomitic mudstones of the Bellerophon Fm., thin black marls with volcanic pebbles and then fossiliferous mudstones interbedded with fine sandstones (80cm) before the appearance of an oolitic level (50cm).

Subsequently the criterion of locating the boundary at the base of the first oolitic bed was instead chosen for the Dolomites, and also for the Trento-Val Sugana area, by many authors, including: Ghetti & Neri (1983), Neri & Pasini (1985), Broglio Loriga *et al.* (1986), Noè (1987), Broglio Loriga *et al.* (1988) and Buggish & Noè (1988). All these authors define the transition between the two formations as very gradual.

More recently the Permian-Triassic succession was also interpreted in terms of sequence stratigraphy; the B-W boundary, *i.e.* the base of the first oolitic level, falls in the middle of a PAC (punctuated accretionary cycle) for Wignall & Hallam (1992), or in the middle of a transgressive system track bounded by erosional surfaces (the only remnant of a third-order cycle) for Noè & Buggish (1994). This boundary, for Massari *et al.* (1994), Massari & Neri (1997), Massari & Neri (in Cassinis *et al.*, 1998) and Neri (in Cassinis *et al.*, 1999), corresponds instead to the boundary between the two basal parasequences (CU cycles) of a third-order cycle.

The evolutionary facies trend and morphological characters led Cirilli *et al.* (1998) to locate the boundary, in the Seres section, at the base of a sequence in the middle of which the first oolites appear. Beretta (1999) confirmed the validity of this interpretation from other sections of the Dolomites as well, also using petrographical and geochemical data.

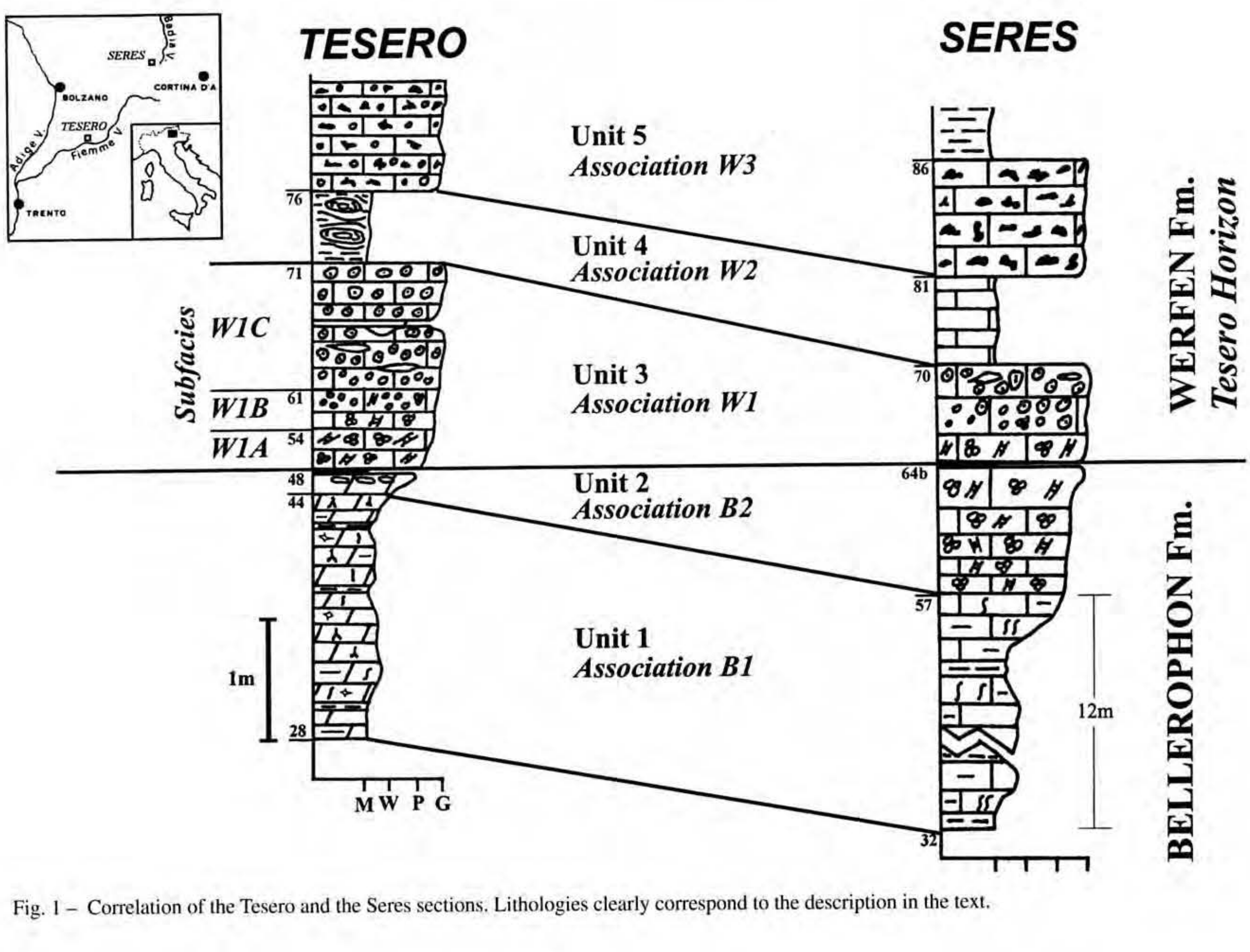

Fig. 1 – Correlation of the Tesero and the Seres sections. Lithologies clearly correspond to the description in the text.

LITHOSTRATIGRAPHY AND FACIES ASSOCIATIONS (FIG. 1)

Tesero Section: Lithostratigraphy

BELLEROPHON Fm.

Unit 1: Tes28 to Tes44 (2.40 m): The interval is dominated by grey-hazel dolomitic beds (2-10cm thick), marly and slightly vacuolar, with thin interlayers of bituminous clays. Bioturbations are frequent, with peculiar subvertical burrows, interpreted as root traces.

Unit 2: Tes45 to Tes48 (20cm): This consists of nodular, compact, grey-yellowish dolomitic beds, 2-5cm thick, showing a rapid transition to a marly black layer at the top; this level, rich in organic matter, is 7-10cm thick and more calcareous and nodular in the upper part.

WERFEN Fm.- TESERO HORIZON

The Tesero Horizon is about 8m thick here, but we will describe the lower 3m only. A sharp morphological boundary separates the TOH from the underlying unit.

Unit 3: Tes49 to Tes71 (2m): The base consists of dark grey, compact limestones in a 40cm thick bed, subdivided into six levels by stylolites or very thin pelites. Light grey limestones, in a 40cm layer resting on a wavy surface, overlie this interval. It is made up of five beds: the two lower ones are bioclastic whereas the upper ones (32cm) are oolitic; these beds are lenticular, separated by stylolites and with a clear low-angle lamination. The same characteristics are also present in the overlying grey oolitic limestones (60cm), with mudstone lenses and channels at the top. A 10cm thick packstone-wackestone separates these layers from a thick bed (50-80cm) of grey-hazel oolitic limestones at the top of the unit.

Unit 4: Tes72 to Tes76 (70cm to 1.1m): This unit shows different lithologies (and different thicknesses) with rapid lateral variations. Variously shaped microbialitic mud mounds are in fact laterally replaced by close interbeddings of grey-hazel mudstone (2-7cm thick) and marls (up to 3-4cm thick). Thin dark to yellowish marls and mudstones are sometimes interbedded in the mud mounds. Unit 4 is overlain by a very thick bed (up to 1m) of grey oolitic and/or intraclastic limestones, which forms the base of the middle and upper parts of the TOH.

Tesero Section: Petrographic-sedimentological analysis and facies associations

Detailed analysis has allowed us to identify different petrographical-sedimentary facies associations in the lithostratigraphic units.

BELLEROPHON Fm.

Association B1 (Tes32 to Tes44): Bioturbated mudstones and mudstone-wackestones prevail. The rare allochems are represented by bioclasts only (algae and bivalve fragments, gastropods and rare foraminifers). Thin microbialitic laminations are present in the upper part. Dolomite and calcite after anhydrite and gypsum pseudomorphs have been observed. The matrix consists of non-homogeneous dolomicrite, often pelletoidal.

Association B2 (Tes45 toTes48): This is mainly represented by wackestone-packstones characterised by an abundance of silt-sized peloids and algal fragments. Level Tes47 shows an increase in the terrigenous content and the presence of nodules of different lithologies: a) packstone-wackestones with echinoid and algae fragments, a high content of clay, organic matter, silty quartz grains, oxides and sulphides; this petrographic facies is also found in nodules at the top of the overlying black layer; b) unsorted detrital packstone-grainstone, with bioclasts (algae, bivalve and echinoid fragments, foraminifers), oomoldic oolites (Ø=100-200 microns), micritic oolites (with bioclastic nuclei or formed by micritic laminae only; Ø up to 500 microns); 10-15% of the constituents consist of terrigenous silty-sandy grains (quartz, muscovite and feldspar).

WERFEN Fm.- TESERO HORIZON

Association W1 - Three subfacies can be identified within this association.

Subfacies W1A (Tes49 to Tes54): This consists of packstones and rudstones, with packstone-grainstones at the top. Allochems are represented by bioclasts only (echinoid and algae fragments and foraminifers are more frequent). The rudstone levels are characterised by iso-oriented echinoid fragments, an increase of marly-clayey influx and a pseudo-flaser type structure. The bioclasts show, towards the top, a slight decrease in size and an increase in roundness; they are also more packed, better sorted and generally thinly coated. The matrix consists of non-homogeneous micrite. The echinoid fragments are always surrounded by a large syntaxial sparitic rim.

Subfacies W1B (Tes55 to Tes61): Interbedded packstones, grainstone-packstones and floatstones (0.5 to 2cm thick) prevail, although grainstones are rare. The bioclasts (mainly algae, then echinoid fragments, foraminifers, gastropods, bivalves and ostracods) prevail in the basal layers (up to Tes58) and in floatstones and detrital packstones; these are interbedded in the oolitic sediments, which gradually predominate from Tes58 onwards. At first oolites are oomoldic, small (Ø=100-200 microns) and surficial; numerous micritic peloids of similar size are present. Larger (Ø=0.4-0.5 mm) and complex oolites, spherical and mainly micritic, gradually replace the ever-present oomoldic ones towards the top. In the lower levels, as in the immediately underlying ones, bioclasts are thinly coated, more rounded and packed and better sorted. A few layers with finer bioclasts and marly clasts can also be noted. The matrix, where present, is micritic. The oomoldic oolites most probably derive from dissolution of aragonitic peloids and a further crystallisation of fine to medium equigranular sparite in the intragranular voids. These peloids originate from direct precipitation from oversaturated marine waters and characterise the initial phases of oolitic deposition. Three sparitic cementing generations can be observed here. The beds are subdivided into centimetre-thick layers by frequent and high amplitude columnar stylolites, mainly subparallel to the bedding surface. A second less evident stylolytic system overlaps the low-angle lamination. Similar characteristics can be noted in the overlying layers.

Subfacies W1C (Tes62 to Tes71): This is mainly composed of grainstones, with a few interbedded floatstones, mudstones and packstone-wackestones. Oolites are dominant among the allochems: grain-size (Ø from 0.5 to 1mm), complexity, proportion of aggregated and botryoidal forms, and sorting all increase towards the top. Micritic and oomoldic oolites and those with bioclastic/terrigenous nuclei are all present together; the former are no longer present at the top of the interval. Bioclasts are frequent in floatstones only, where a few irregular and millimetre-sized intraclasts (reworked oolitic grainstone) have also been observed. Some levels (Tes67 and Tes71) show a transition from oolitic grainstone at the base to oolitic and/or bioclastic floatstone, and to mudstone at the top. The micritic matrix, where present, is non-homogeneous.

Association W2 (Tes72 to Tes76): mudstones and mudstone-wackestones prevail. The few bioclasts are often transformed into microsparitic ghosts. Several strata show a microsparitic to finely sparitic, irregular, pervasive, anastomosing reticule, entrapping the pelletoidal and non-homogeneous micritic matrix. These characteristics are generally referred to the diagenesis of an original microbialitic/trombolitic texture (Bourque, 1997; Baud *et al.*, 1997).

Association W3 (Tes77 to Tes92): The basal part of the association only, where unsorted coarse grainstones and packstone-grainstones prevail, is described here. The allochems consist of bioclasts, small oomoldic oolites, large intraclasts and micritic algal peloids. The micritic matrix is mainly non-homogeneous or pelletoidal. Grainstones and packstone-grainstones show two sparitic cementing generations.

Seres section: Lithostratigraphy

BELLEROPHON Fm.

Unit 1: S32 to S57 (12m): This consists of interbedded grey-yellowish bioturbated marly limestones and marls. Bed thicknesses vary from 10-20cm to 60-70cm in both lithologies; the bedding shows irregular surfaces and nodular structures.

Unit 2: S58 to S64b (1.1m): This consists of compact greyblackish limestones and is characterised by the disappearance of bioturbation. Bed thickness varies from 10-20cm in the lower part to 30cm towards the top. The top of the unit consists of a 3-5cm black organic-rich marly layer that marks the boundary with the overlying Werfen Fm.

WERFEN FORMATION - TESERO HORIZON (4.90m)

We will describe here only the lower 2.60m of the TOH.

Unit 3: S65 to S70 (85cm): The base consists of a 50-cm-thick bed, subdivided into at least six layers by low-amplitude stylolites; the first 20cm are made up of dark limestones, the upper 30cm of greyish oolitic limestones. Two other oolitic beds follow, separated by high-amplitude stylolites and with scattered micritic lenses.

Unit 4: S71 to S81 (75cm): This consists of fine light-grey limestones, in 8-10cm thick beds, resting on an undulating surface.

Unit 5: S82 to S86 (1m): This consists of compact greyish coarse limestones, in irregular or lenticular beds, 20 to 35cm thick. This unit is overlain by light-grey micritic limestones, in thick to very thick beds (25-35cm to 1m).

Seres section: Petrographical-sedimentological analysis and facies associations

BELLEROPHON Fm.

Association B1 (S32 to S57): Bioturbated wackestones and packstones prevail, mudstones are subordinate and grainstones are rare. Bioclasts are dominant (bivalves and ostracods, then algae and miliolids among others). Oncolites can be observed in some layers. Some grainstones and packstones show iso-oriented bioclasts. A bioclastic silty-arenitic fraction is also common and characterises the biocalcarenitic levels. The micritic matrix is more uniform in the mudstone layers than in the coarser ones. The grainstones are characterised by eogenetic meteoric-phreatic cements.

Association B2 (S58 to S64b): Bioclastic packstones and packstone-grainstones prevail; large algae fragments, highly diversified foraminifers and algal peloids are the more frequent allochems. Packstones are commonly characterised by iso-orientation of the bioclasts, pseudo-flaser textures and the constant presence of a bioarenitic fraction. Bioclasts show a strong diagenetic compaction. The black marly layer at the top has a high organic fraction that almost completely obliterates the compositional and textural characteristics, similar to those of the underlying levels. The micritic matrix is generally scarce and non-homogeneous.

WERFEN FORMATION - TESERO HORIZON

Association W1 (S65a to S70): The base is composed of bioclastic packstone-wackestones and packstone-grainstones (S65a and b) similar to those described for the upper part of the Ass. B2. The bioclasts are mainly represented by highly diversified foraminifers, algae fragments and algal peloids. Level S65c is a bioclastic grainstone with few oomoldic oolites; the grains are well rounded here, less sorted and thinly coated. The upper part of the association consists of oolitic grainstone; bioclasts (foraminifers and gastropods) are rare and present mainly in the lower layers. Oolites are at first small (Ø = 0.2-0.3mm), oomoldic and surficial, micritic or with bioclastic nuclei; they are well sorted and, upwards, are gradually larger (Ø from 0.4 to 0.8-1mm) and more complex, with the micritic and oomoldic types prevailing; at the top, aggregated ooids and botryoidal forms are frequent, together with a few bioclasts and large calcarenitic intraclasts.

Association W2 (S71 to S81): This consists of mudstones, with closed and narrow undulating laminae; very rare ostracods are the only skeletal grains. Irregular, elongated and oriented micritic intraclasts are observed in the wackestones, infilling isolated microchannels in the central part of the interval. The terrigenous content (up to 3%) is formed by quartz grains (Ø 0.2-0.3mm) and by mica lamellae (up to 10mm). The matrix is micritic, brown and homogeneous.

Association W3 (S82 to S86): This is represented by intraclastic grainstones only. The intraclasts are microsparitic and have a rounded irregular shape; they are often fractured, sometimes iso-oriented. Skeletal grains (gastropods and ostracods) are quite rare.

CORRELATIONS AND DEPOSITIONAL EVOLUTION

The depositional environment across the boundary between the Bellerophon Fm. and the Werfen Fm. (Tesero Horizon) developed through the phases described below (Fig. 1).

Regressive (low energy) phase: Ass. B1 deposited, at Seres, in a shallow-water, low-energy, sheltered environment. Conditions of lowest energy and restricted water circulation are reached at the top of the association. Some levels, showing eogenetic meteoric-phreatic cements, could indicate a near intertidal environment. This association, in the central-eastern Dolomites, may be defined as "Low Energy Badiota Facies"; it represents a widespread regressive episode, already documented (Broglio Loriga *et al.*, 1988). At Tesero the characteristics of Ass. B1 point to inter/supratidal, clearly hypersaline conditions, then to a shallower and more restricted environment in relation to the underlying sediments. On the basis of these characteristics, the Ass. B1 of the two sections can be correlated to a good extent.

Transgressive phase - Black Level - Basal Oolitic Bed

Transgression and Black Level: The Seres section shows, from the very base of Ass. B2, a gradual and constant increase in the energy level: a strong current regime dominates the shallow waters and subtidal environment. At the top theAss. B2 shows a black organic-rich marly level. At Tesero the "fiemmazza" facies of the Bellerophon Fm. ends with about 10cm of dolomites showing a facies of higher hydrodynamic energy. This interval has only recently been recorded by one of the authors (Beretta, 1999). A black organic-rich marly layer is present at the top of this section too. Ass. B2 therefore show a good degree of correlation in both the sections. The black layer is found in many sections of the area and can be considered synchronous on a regional scale. The abundance and size of woody remains of continental origin, the palynological content (Cirilli *et al.*, 1998) and frequent terrigenous clasts indicate maximum proximity at the top of Ass. B2. This datum fits well, at least at Seres, with the interpretation of Ass. B2 as a CU and shallowing-up cycle. The Black Level is chosen as the boundary between the Bellerophon and the Werfen Fms; furthermore it does not represent a discontinuity in the evolution of the petrographic-sedimentary microfacies.

Basal Oolitic Bed: The lower part of this thick bed (Ass. W1) is always formed by bioclastic layers with rounded and coated grains, peloids and rare small oolites at the top. The oolitic layers in a strict sense follow with a gradual evolution; towards the top, the ooids increase in size and complexity. A reverse grading trend in oolitic sediments has been recorded from bars (Persian Gulf; Loreau & Purser, 1973), and from prograding shoals (Bahamas; Bathurst, 1975). A comparison with these modern areas enables us to interpret the Basal Oolitic Bed as an elementary CU cycle. The environment varies from subtidal to intertidal with even higher hydrodynamic levels. At Seres the succession is rather "clean". At Tesero, frequent muddy and/or bioclastic interbeddings are instead present among the oolitic beds; furthermore the top oolitic layer is slightly different in relation to the underlying ones, with poorer sorting and lower hydrodynamic levels. However, the frequent lateral facies variations lead us however to hypothesise, for the whole Oolitic Bed, an environment in which various constituents accumulated; these elements derived from surrounding areas of oolitic and/or bioclastic production (bars, dunes, less restricted platforms) in diverse and differentiated sedimentary conditions. This Bed shows a regional distribution, with thickness decreasing eastwards: 2m at Tesero (50cm of the "non-oolitic" lower part) and 85cm at Seres (20cm "non-oolitic"). This bed is considered to be a record of a very rapid transgression which lasted a few thousand years only and, as a consequence, was synchronous throughout the area.

Mudstones (low energy) and Microbasaltic phase: The basal oolitic phase was followed by a period of low hydrodynamism over the whole Dolomites; microbialitic or stromatolitic structures are in some places associated with

this phase (Ass. W2). At Tesero intertidal microbialitic bodies are interbedded with mudstones and pelitic layers, typical of tidal ponds. The transition between Ass. W1 and Ass. W2 is sharp, but at Tesero a depositional continuity seems to be present. This phase can be considered synchronous on a regional scale too.
Deepening phase – Storm Facies: An accurate analysis of this interval is beyond the scope of this study; we can simply note that it consists of sediments typical of muddy bottoms (below the fair-weather wave-base) with interbeddings of oolitic layers (also intra- and bioclastic); the latter do not show the typical reverse grading of the Basal Oolitic Bed, but are totally unsorted; this is interpreted as being due to basinwards storm sedimentation.

CONCLUSIONS AND INTERPRETATION OF THE FORMATIONAL BOUNDARY

The Basal Oolitic Bed, over the whole area of the Dolomites, shows a continuous evolution of the petrographical-sedimentary characteristics; this evolution, in our opinion, does not make it possible to locate formational or sequence boundaries inside the Bed. This event, if compared with present-day depositional areas, may actually represent an interval of a few thousand years.

For all these reasons, we consider it more appropriate to locate the Bellerophon-Werfen formational boundary at the base of the Basal Oolitic Bed.

If the said time-span is correct, the meaning of bio- and chronostratigraphical events in this interval should also be reconsidered, as the resolution of these dating methods is clearly lower.

This Bed, moreover, is part of an evolutionary trend that begins in the upper Bellerophon Fm. and continues in the Werfen Fm. without evident depositional discontinuities. This interpretation is in contrast with the more traditional location of the formational boundary and the more recent sequence stratigraphy interpretations (see Introduction). We can instead confirm that the interpretation proposed by Cirilli *et al.* (1998) for the Seres section is valid on a regional scale. Further confirmation can be drawn from other sections also; their description, with detailed petrographical (cathodoluminescence, SEM-EDS analyses), biostratigraphical and isotopic geochemistry data, will be the subject of a paper currently in preparation.

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THE FUNES/VILLNÖSS BASIN: AN EXAMPLE OF EARLY PERMIAN TECTONICS, MAGMATISM AND SEDIMENTATION IN THE EASTERN SOUTHERN ALPS (NE ITALY)

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Key words – Eastern Southern Alps; Permian; Athesian Volcanic District.

Abstract – The Funes Basin is an intermontane basin located in the northern Dolomites. At present, because of Neoalpine compressional tectonics, it is exposed for 2 km in a NE-SW direction and for about 3 km in an E-W direction. The basin is filled by up to 500 m thickness of andesitic lavas, which roughly thin eastwards and are replaced by up to 200 m of pyroclastic – mainly agglomerate – deposits. Stratigraphical, sedimentological and tectonic data indicate that the evolution of the Funes Basin was controlled by normal faults. No evidence of transcurrent tectonics was observed. An Early Permian age for the Funes Basin is suggested by the presence of lenses of Ponte Gardena Conglomerate within andesitic lavas and agglomerates, and by the calcalkaline cogenetic nature of the andesites of Funes and the Permian volcanics of the neighbouring Athesian Volcanic District.

Parole chiave – Alpi Meridionali orientali; Permiano; Distretto Vulcanico Atesino.

Riassunto – Il bacino di Funes è un bacino intramontano affiorante nella porzione settentrionale delle Dolomiti. Attualmente, a causa dell'intensa tettonica compressiva neoalpina, affiora solo per circa 2 km in direzione NE-SW e per circa 3 km in direzione E-W. Il bacino è riempito da lave andesitiche per uno spessore massimo di 500 m. Nella porzione orientale del bacino le lave sono rapidamente sostituite da potenti depositi di agglomerati cui si intercalano depositi alluvionali (per uno spessore massimo di circa 300 m). I dati sinora raccolti indicano che l'evoluzione del bacino di Funes è stata guidata da faglie normali; nessuna evidenza di tettonica trascorrente è stata sinora osservata. La presenza di lenti di Conglomerato di Ponte Gardena sia nelle lave che nei depositi piroclastici e l'affinità compositiva fra le andesiti di Funes e le vulcaniti del Distretto Vulcanico Atesino, suggeriscono che il bacino di Funes si sia evoluto durante il Permiano inferiore.

INTRODUCTION

During Late Carboniferous and Early Permian times, the central-eastern Southern Alps were the centre of intense volcanic activity linked to the formation of a series of intermontane basins and cauldrons (Massari, 1988; Cassinis *et al.*, 1997 and refs therein). The geometry, stratigraphy and evolution of such basins are important tools in deciphering the Late Paleozoic tectonic setting of the eastern Southern Alps. Up to now two contrasting scenarios have been proposed: i) a Late Variscan dextral transtensional regime (*e.g.* Cassinis & Perotti, 1994); and ii) opening of the Mesozoic Tethys (*e.g.* Selli, 1998).

The Funes Valley, located on the left side of the Isarco/Eisack river near Chiusa/Klausen, provides a good opportunity of understanding the structural relationships between the basal volcanic sequence of the " Athesian Volcanic District" (AVD), its coeval terrigenous volcaniclastic cover and the underlying Variscan metamorphic basement.

GEOLOGICAL SETTING

The Funes Valley is located in the westernmost part of the eastern Southern Alps (Fig.1), at the northern margin of the "Dolomitic sinclinorium" (Doglioni, 1987). The valley is located in the footwall of a Neogene north-vergent backthrust system (Val di Funes and Passo delle Erbe backthrusts according to Doglioni, 1987, and Ring & Richter, 1994).

This valley cross-cuts the Variscan metamorphic basement and its basal terrigenous and volcanic cover of Late Carboniferous to Permian age, showing their original structural and stratigraphical relationships. In the study area, the stratigraphic succession starts with the Variscan metamorphic rocks consisting of prevailing metapelitic and metapsammitic rocks of Early Paleozoic age (Bressanone Phyllite, Val Digon Fm. and Eores Quartzite), with interbedded acidic (Comelico Porfiroids, Caradocian) and basic (Gudon Fm., Early Silurian?) metavolcanic layers (Poli *et al.*, 1996).

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The metamorphic thermal peak is Visean in age (350 Ma; Del Moro *et al.*, 1980), and white mica cooling ages suggest a date of approx. 320 Ma for cooling below 350°C (Hammerschmidt & Stöckhert, 1987; Meli & Klötzli, 1997).

In the central-eastern Southern Alps, the deposition of clastic continental sequences of the "basal conglomerates" (Upper Carboniferous-Lower Permian) testifies the uplift and the erosion of the greenschist-facies Variscan basement as from the Late Carboniferous at least (Krainer, 1989).

During the Early Permian the magmatic activity began; the intrusion of the Bressanone Granodiorite (282 Ma±14; Del Moro & Visonà, 1988) and of coeval diorites (Luson, Chiusa) was followed by the deposition of the AVD sequence (D'Amico & Del Moro, 1988; Barth *et al.*, 1994; Di Battistini *et al.*, 1989; Rottura *et al.*, 1998). The deposition of the Ponte Gardena Conglomerate preceded and/or accompanied the AVD emplacement. Finally, the deposition of the Val Gardena Sandstone in Tatarian times (Massari *et al.*, 1994) sealed both the volcanics and the effects of the tectonism of Early Permian age.

THE FUNES BASIN

A geological sketch-map of the study area is shown in Fig. 2, while the tectonostratigraphic relationships are summarised in the sections of Fig. 3.

Abrupt lateral variations in the stratigraphic succession reflect intense tectonic activity during the deposition of the volcanosedimentary sequence, which reaches a maximum thickness of 500 m. The deepest portion of the volcanic sequence crops out in the Mittermühl locality where, however, Quaternary cover hides the underlying metamorphic basement. Here, andesite block-lavas, altered by hydrothermal fluids and intruded by basaltic-andesite dykes, crop out. In contrast, on Hauben/Mt. Cappello, at the top of the series, the andesite lavas have an amygdaloidal pyroxene-phyric texture. The irregularly-shaped elongated amygdales, usually filled with chalcedony, testify to both the magma high vapour-phase content of the magma, and the intensity of the late-stage magmatic activity.

Thin, discontinuous layers of quartz-rich, biotite-bearing tuffs with rhyodacitic lapilli are interbedded in the andesite lavas. Near Tiso/Teis, tuffs directly overlie the metamorphic basement, here strongly deformed and intruded by sets of dykes.

Andesites and basaltic-andesite dykes always intrude the andesite lavas. Petrographical and geochemical analyses on andesite dykes reveal their low- to high-K orogenic andesite nature (Table I and Fig. 4). The similarity with the neighbouring Chiusa epiplutonic diorite is also demonstrated. The same composition was obtained for the lava flows and block-lavas, as already found for all the Permian lavas of the AVD (Rottura *et al.*, 1998).

The andesite lavas reduce in thickness eastwards where they are replaced and covered by up to 200 m thick agglomerates and breccias. Along the northern margin of the basin, agglomerates and breccias directly overlie the metamorphic basement.

Agglomerates are chaotic and heterometric with clasts of prevailing andesitic composition, but clasts of rhyolites or rhyodacites as well as the underlying metamorphic basement can be locally found. Graded and cross-bedded metre-scale quartz-rich lenses of tuffaceous arenites or microrudites are sometimes present. Locally some reddish clay lenses were found. Bodies of

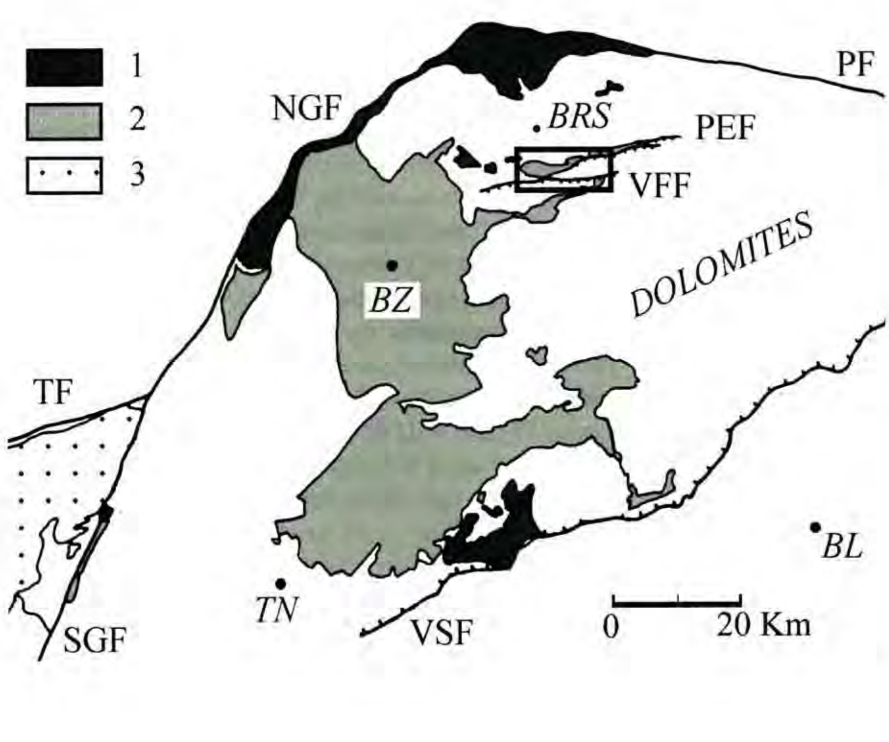

Fig. 1 – Location of the study area (rectangle) in the eastern Southern Alps. Distribution of the volcanic and plutonic rocks is shown. 1) Permian plutonics; 2) Permian volcanics of Athesian Volcanic District (AVD); 3) Oligocene Adamello pluton; NGF: North Giudicarie Fault; PEF: Passo delle Erbe Fault; PF: Pusteria Fault; SGF: South Giudicarie Fault; TF: Tonale Fault; VFF: Val di Funes Fault; VSF: Valsugana Fault. *BL*: Belluno; *BRS*: Bressanone/Brixen; *BZ*: Bolzano/Bozen; *TN*: Trento.

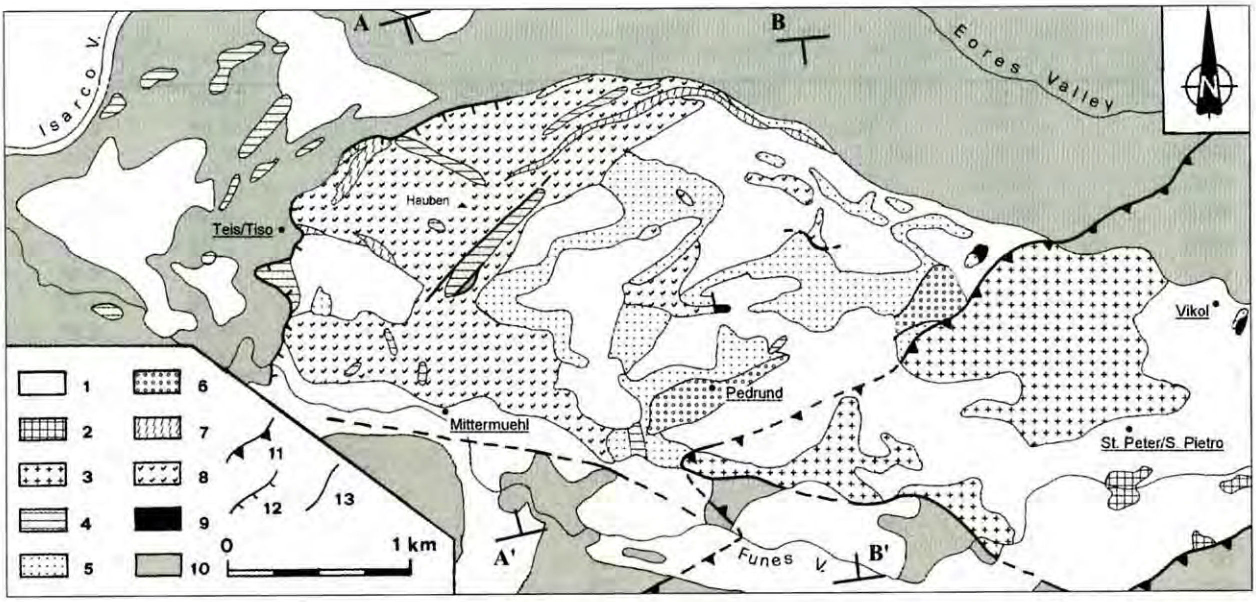

Fig. 2 – Geological sketch-map of the study area (location in Fig. 1). 1) Quaternary cover; 2) Val Gardena Sandstone; 3) rhyolites and rhyodacites; 4) andesitic to basaltic-andesitic dykes; 5) agglomerates; 6) Pedrund unit; 7) tuffs; 8) andesitic lavas; 9) Ponte Gardena Conglomerate; 10) metamorphic basement; 11) Neogene thrust; 12) Permian normal fault; 13) fault. A-A' and B-B': traces of geological sections in Fig. 3.

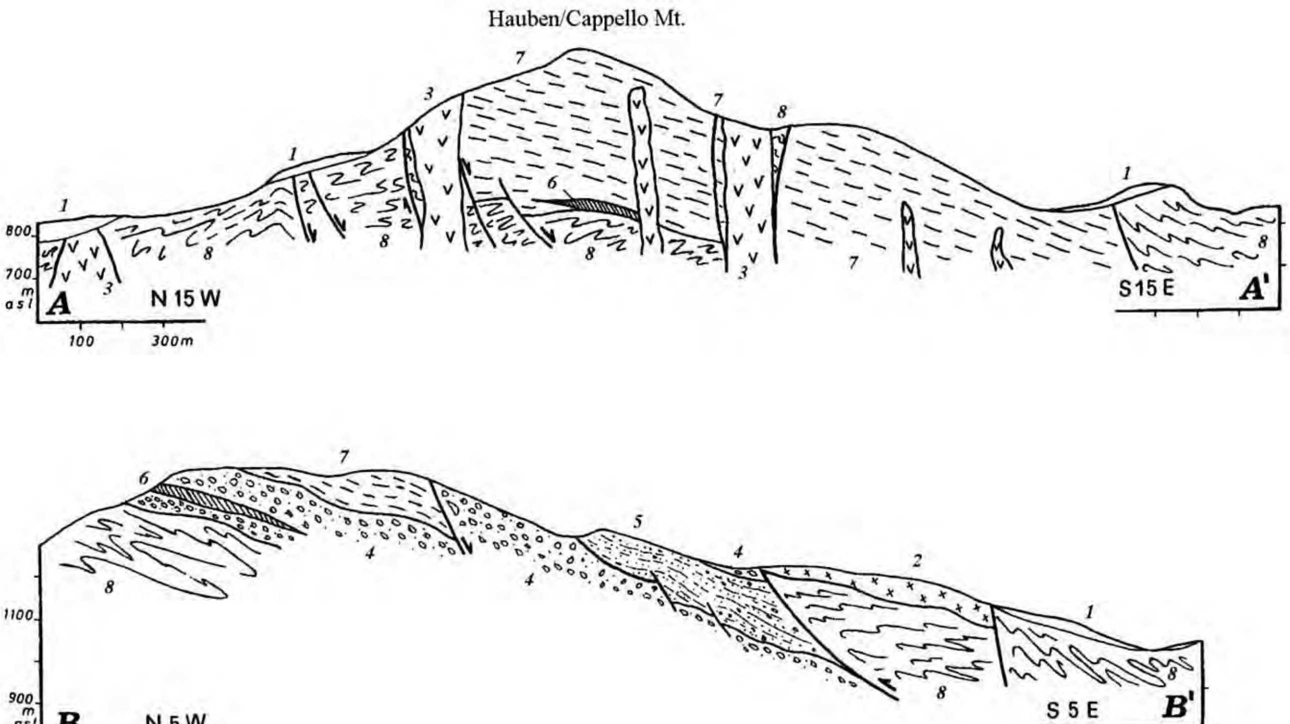

Fig. 3 – Geological cross-sections (*location in Fig. 2*). Legend: 1) Quaternary cover; 2) rhyolites and rhyodacites; 3) andesitic to basaltic-andesitic dykes; 4) agglomerates; 5) Pedrund unit; 6) tuffs; 7) andesitic lavas; 8) metamorphic basement.

Sample	TPZ6	TPZ19	TPZ30	TPZ48	TPZ53	TPZ63	TPZ54	86-4
Rock type	L 2	L 2	L 1	D 2	L 1	L 2	D 2	D 1
SiO2 (wt%)	55.45	55.76	56.69	55.41	56.36	53.70	53.38	56.91
TiO2	0.74	0.80	0.93	0.80	0.93	0.80	0.94	0.71
Al2O3	15.18	17.11	17.02	16.73	17.03	17.33	17.05	16.34
Fe2O3*	8.04	7.35	7.29	7.64	7.33	7.78	7.75	7.49
MnO	0.13	0.13	0.16	0.14	0.10	0.16	0.16	0.13
MgO	7.20	4.12	3.91	4.96	4.25	5.44	5.32	4.47
CaO	9.23	9.46	7.90	8.59	7.84	9.16	9.87	6.94
Na2O	1.68	1.61	1.90	1.67	1.94	1.74	1.52	1.73
K2O	0.45	0.63	2.30	0.48	2.20	0.35	0.37	2.21
P2O5	0.13	0.14	0.19	0.17	0.19	0.15	0.20	0.11
Total	98.23	97.11	98.29	96.59	98.17	96.61	96.56	97.04
LOI	1.74	2.90	1.61	3.37	1.73	3.79	3.41	3.11
Mg #	63.94	52.61	51.51	56.25	53.45	58.07	57.62	54.17
Cr (ppm)	640	248	220	206	222	300	234	277
Ni	80	27	45	34	45	32	28	18
Co	30	21	19	21	23	25	24	43
V	172	132	146	156	153	179	165	179
Cu	24	19	16	19	19	21	18	20
Pb	13	15	16	13	14	11	14	14
Zn	85	85	82	158	86	90	92	91
Rb	19	11	95	13	99	10	11	96
Ba	311	369	403	462	355	247	304	406
Sr	482	1387	312	839	311	775	935	199
Ta	0.76	0.83	0.97	0.82	0.92	0.85	0.92	0.98
Nb	9.3	9.9	28.7	10.0	11.8	10.3	11.8	8.9
Hf	3.87	4.22	4.32	4.00	3.82	3.74	4.02	3.47
Zr	148	160	172	158	171	149	172	134
Y	24	25	27	25	25	25	26	24
Th	8.94	9.73	10.1	10.4	10.3	9.32	9.42	9.00
U	1.73	1.89	2.04	1.85	1.9	1.66	1.81	1.85

Table I – Major and trace element contents for lavas (L) and dykes (D) of andesite (1) and basaltic andesite (2) composition in the study area.

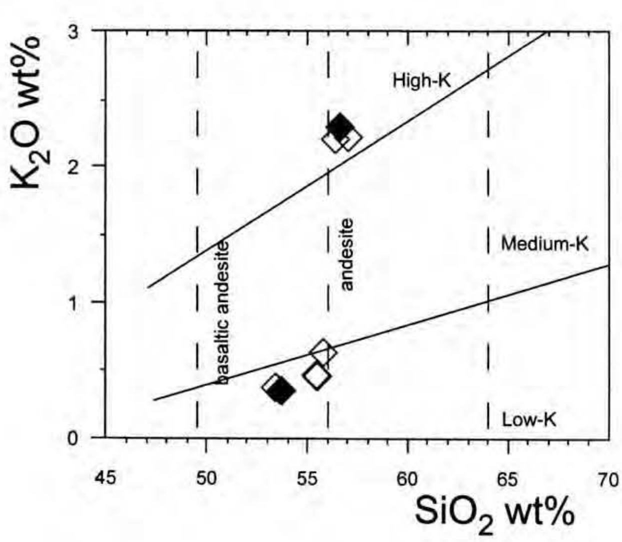

Fig. 4 – K_2O vs silica classification diagram (Rickwood, 1989) for lavas (open diamond) and dykes (filled diamond). According to Bargossi *et al.* (1981) and Di Battistini *et al.* (1989), lavas and dykes both show low to high K_2O contents.

chaotic andesitic breccias suggest a proximal location.

Decimetre to metre-scale lenses of conglomerates and breccias composed of fragments derived from the underlying low-grade basement (Ponte Gardena Conglomerate) are sometimes interbedded within the pyroclastic deposits.

Between Miglanz and St. Valentin/St. Valentino, alluvial terrigenous sediments (the *Pedrund unit*) replace the agglomerates. From the bottom to the top, thin layers of red to green arenites and red pelites pass into polygenic conglomerates (clasts of andesites, phyllites, phyllitic mica-schists, quartzites and scarce rhyolites) crop out. A clear stratification is emphasised by decimetre-thick units of microrudites and red arenites. Some caliche-rich horizons in pelites and graded or cross-bedded arenites are observed in this unit.

On top of the volcanosedimentary sequence, rhyolites, rhyodacites and rhyolitic ignimbrites have been emplaced tectonically. This volcanic unit overthrusts both the sedimentary units (*Pedrund unit* and agglomerates) and the metamorphic basement.

STRUCTURAL ANALYSIS

Compressional tectonics. The present structural setting of the study area is linked to the Neogene tectonic event (Doglioni, 1987; Castellarin *et al*., 1992; Ring & Richter, 1994). Low- to middle-angle NNW-vergent backthrusts, a general tilting to the SSE and the inversion of some inherited normal faults are ascribed to this event.

Moreover, along the Funes creek (Fig. 2), a subvertical WNW-ESE trending fault separates the metamorphic basement from the volcanosedimentary succession. According to Doglioni (1987) and Ring & Ricther (1994), it should correspond to a segment of the Funes Line, the major Neoalpine backthrust in the northern Dolomites. This fault is actually cut by the NE-SW trending NNW-vergent thrusts, and its fault zone shows evidence of a polyphase tectonic history and very different structural features with respect to the NNW-vergent backthrusts.

Extensional tectonics. The western edge of the basin is well preserved and exposed near Tiso. Here the metamorphic basement is characterised by: i) NE-SW and NW-SE trending andesitic and basaltic-andesite dikes (Fig. 5a), ii) N-S to NE-SW trending subvertical extensional fractures and sulphide- or barite-bearing veins; such veins also intrude andesitic-fault-breccias along NE-SW normal faults; iii) "fluidisation" breccias (*sensu* Reynolds, 1954). Locally a N 50°-trending dyke cross-cut by another trending N 160° was observed. In contrast, no dyke, vein or other extensional structure was observed in the rhyolite and rhyodacite lavas.

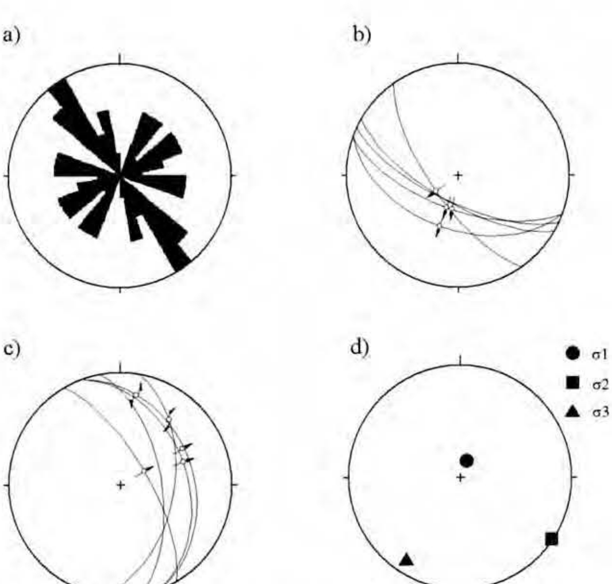

Fig. 5 – Selected structural data collected in the Funes Basin. a) Strike of andesitic dykes. Dataset: 37, interval: 10°, max: 13,51°; b) attitude and vector displacement of normal listric faults in the metamorphic basement. Dataset: 5; c) attitude and vector displacement of synsedimentary normal faults in the Pedrund unit; d) calculated stress field for Permian normal faults.

Also, the northern margin of the Funes Basin was tectonically active during the volcanism, as suggested by the presence of a set of WNW-ESE trending andesitic dykes, NW-SE trending and SW-dipping normal faults in the metamorphic basement (Fig. 5b), and a lens of agglomerates directly overlying the metamorphic basement. At least within the *Pedrund unit*, synsedimentary metre-scale faults trending NNW-SSE and dipping ENE were observed (Fig. 5c).

Both NW-SE and NNW-SSE normal faults are consistent with a pure extensional stress field with horizontal, NE-SW trending σ_3 and σ_1 subvertical (Fig. 5d).

FIRST RESULTS

These data indicate the presence in the northern Dolomites of an intermontane basin that at present is exposed for 2 km in a NE-SW direction and for about 3 km in an E-W direction. A Neoalpine NNW-vergent thrust at the eastern edge and a WNW-ESE-striking subvertical fault at the southern edge prevent a correct evaluation of the original width of the basin.

The Funes Basin is filled by up to 500 m thickness of andesitic lavas which roughly thin eastwards and are replaced by up to 200 m of pyroclastic – mainly agglomerate – deposits. A 100 m-thick, pelitic to ruditic alluvial sequence, showing NNW-SSE-trending synsedimentary normal faults, is interbedded within the upper part of the agglomerates.

East-dipping normal faults, andesitic dykes and hydrothermal veins, N-S to NE-SW trending, mark the well-preserved western edge of the basin.

Stratigraphic, sedimentological and tectonic data indicate that the evolution of the Funes Basin was controlled by normal faults. No evidence of transcurrent tectonics was observed.

An Early Permian age for the Funes Basin is suggested by the presence of lenses of Ponte Gardena Conglomerate within andesitic lavas and agglomerates and by the calcalcaline cogenetic nature of the Funes andesites and the Permian volcanics of the neighbouring Athesian Volcanic District.

In order to determine the precise age of formation of the Funes Basin, isotopic dating of zircons from the andesites is in progress.

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UPPER PALEOZOIC AND TRIASSIC CONTINENTAL DEPOSITS OF SARDINIA: A STRATIGRAPHIC SYNTHESIS

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Key words – stratigraphy; paleontology; Sardinia; Upper Paleozoic basins.

Abstract – Recently, a team of Italian and French geologists renewed field research on some (late) post-Variscan terrigenous and volcanic sequences in Sardinia.
From stratigraphical, paleontological, petrological and geochemical investigations, two main volcanic and sedimentary cycles appear so far confirmed.
In the older cycle (cycle 1), Late Carboniferous to Early Permian in age, a number of narrow, fault-bounded subsiding basins developed.
Mostly during the Early Permian ("Autunian") these troughs, which presently crop out in different parts of Sardinia, were characterised by both alluvial-to-lacustrine sediments, with differing lithofacies, and calcalkaline, mainly effusive, intermediate-to-acidic products. These latter volcanics were also distributed in large volumes outside the basin areas, covering a large part of the island.
The deposits pertaining to the second cycle (cycle 2) crop out in a restricted number of sites in northwestern as well as central to southwestern Sardinia.
They consist of fossil-barren alluvial, coarse-to-fine clastic red beds, followed upwards, after an unconformity, by the Buntsandstein, which is biostratigraphically related to late Olenekian(?)-early Anisian and late Anisian ages.
Furthermore, in Nurra, the presence of alkaline volcaniclastic products in the relatively older red beds allows the correlation of the second cycle with similar deposits cropping out in southern France and the Catalonian Pyrenees, generally ascribed to the late Early Permian to early Late Permian (*i.e.* to the "Saxono-Thuringian" *p.p.* of authors), with which Sardinia was linked during post-Variscan times.

Parole chiave – stratigrafia; palaeontologia; Sardegna; bacini tardo-paleozoici.

Riassunto – In questi ultimi anni, un gruppo di ricercatori italiani e francesi ha analizzato in dettaglio alcune delle più importanti successioni vulcaniche e sedimentarie (tardo-) postercíniche della Sardegna. Ciò ha permesso di acquisire nuovi dati stratigrafici, paleontologici e petrografici, confermando l'esistenza nell'isola di almeno due grandi cicli vulcanici e sedimentari. Durante il primo fra questi (ciclo 1), databile tra il Carbonifero superiore e il Permiano inferiore, una fase trascorrente interessò la catena varisica ormai sollevata, dando luogo alla formazione di piccoli e incisi bacini transtensivi. Soprattutto nel Permiano inferiore ("Autuniano"), queste fosse continentali, presenti in varie zone dell'isola, furono caratterizzate da una sedimentazione alluvio-lacustre, con facies molto differenti l'una dall'altra, e da prodotti magmatici, principalmente effusivi, ad affinità calcalcalina. Tali prodotti magmatici risultano ampiamente diffusi anche in zone extrabacinali di varie aree della Sardegna. Le successioni attinenti al secondo ciclo (ciclo 2) affiorano molto più limitatamente delle precedenti e precisamente nella zona nordoccidentale, sudoccidentale e centrale dell'isola. Esse sono caratterizzate da *red beds* alluvionali con caratteri sedimentologici e tessiturali diversi, con presenza di fossili solo nelle porzioni sommitali, che richiamano il Buntsandstein europeo e facenti passaggio all'ambiente marino del Muschelkalk. La correlazione di questi ultimi sedimenti con dati di sondaggio permetterebbe di assegnare loro un'età compresa tra l'Olenekiano?-Anisico inferiore e l'Anisico superiore. Intercalate nella parte basale dei *red beds* del secondo ciclo, ma solo nella Nurra, si trovano prodotti vulcanoclastici a carattere alcalino; ciò porterebbe a riferire l'insieme di questi depositi ad un'età permiana superiore, in base a confronti con le analoghe successioni della Francia meridionale e dei Pirenei catalani, cui la Sardegna era in quel tempo unita.

GENERAL FRAMEWORK

The aim of this paper is to show the current state of research on the Upper Carboniferous to Permian and Triassic sedimentary and volcanic sequences of Sardinia.

In this regard, the fieldwork recently carried out by a group of researchers on some Late Paleozoic basins of the island will be reviewed in the light of the relatively new data.

On the basis of stratigraphical, paleontological, petrographical and radiometric data, two major volcanosedimentary cycles can be recognised in the post-Variscan deposits of Sardinia. The older one (cycle 1) is Late Carboniferous to Early Permian *p.p.* in age, whereas the younger cycle

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(cycle 2) spans from Early Permian *p.p.* (post-"Autunian") to Middle Triassic (Anisian). The latter cycle was also defined as "Permo-Triassic" by some authors (*e.g.* Gasperi & Gelmini, 1980; Fontana *et al.*, 1982; Cassinis, 1996; etc.).

At the beginning of the first cycle, a (trans)tensional tectonic phase affected the Variscan basement, giving birth to a number of small intramontane, fault-bounded subsiding basins, of which the most remarkable crop out presently in various parts of the island (Fig. 1). They are the well-known basins of Lu Caparoni-Cala Viola (NW Sardinia), Seui-Seulo (Central Sardinia), Perdasdefogu, Escalaplano and Lago di Mulargia (SE Sardinia), Guardia Pisano and San Giorgio (SW Sardinia). Clear evidence of Upper Carboniferous deposits occurs only in the ?Westphalian D to Stephanian sediments of the last-mentioned basin (Cocozza, 1967; Del Rio, 1973; Fondi, 1979; Del Rio & Pittau, 1999).

Investigations seem to suggest an early structuration of the local basins in the Perdasdefogu and Seui areas, in a NE-SW trend. Subsequently, but probably still during the Early Permian, these troughs underwent major extensional tectonism, along with NW-SE to NNW-SSE oriented faulting, which generated a horst-and-graben framework. This structural trend, which is generally known as "Hercynian" in Sardinia, corresponds exactly to the major late-Hercynian (or late-Variscan) transcurrent faults (such as those of Posada and Mt. Grighini) and to the (late) post-Variscan dyke swarm cropping out in the western part of the island.

At that time, close to the Carboniferous-Permian boundary but mainly during the beginning of the Early Permian (Autunian of French authors), these troughs were characterised by both alluvial-to-lacustrine sedimentation and calcalkaline intermediate-to-acidic magmatism.

Due to the tectonic activity and the related subsidence, the Lower Permian volcanosedimentary infills of the various basins show thicknesses ranging from just a few metres to almost 400 m (see *e.g.* Broutin *et al.*, 1996). The repeated volcanic activity, mainly in the form of lava flows or shallow intrusions within the basin fill, and of pyroclastic flow and fall deposits originating from nearby volcanic centres, greatly contributed to the variations in basin sequences both in thickness and facies. This explains why precise correlation, even between successions cropping out in adjoining basins, has so far been hampered.

The rich paleontological record (macro- and microfloras, and rarely ostracods, stromatolites, algae and amphibians) yielded by these sediments allows the confirmation of an Early Permian (Autunian) age for most of these sequences.

The second cycle is, at present, clearly developed in northwestern Sardinia (Nurra), and probably in a few areas in central to southwestern regions of the island. It followed a stratigraphic gap of unknown duration and is represented by alluvial red beds, which are intercalated with tuffs and ignimbrites of alkaline composition in Nurra (markedly only at Monte Santa Giusta). This new volcanic and sedimentary period was seemingly connected with a more extensional regime and overlies the first cycle of Lower Permian deposits.

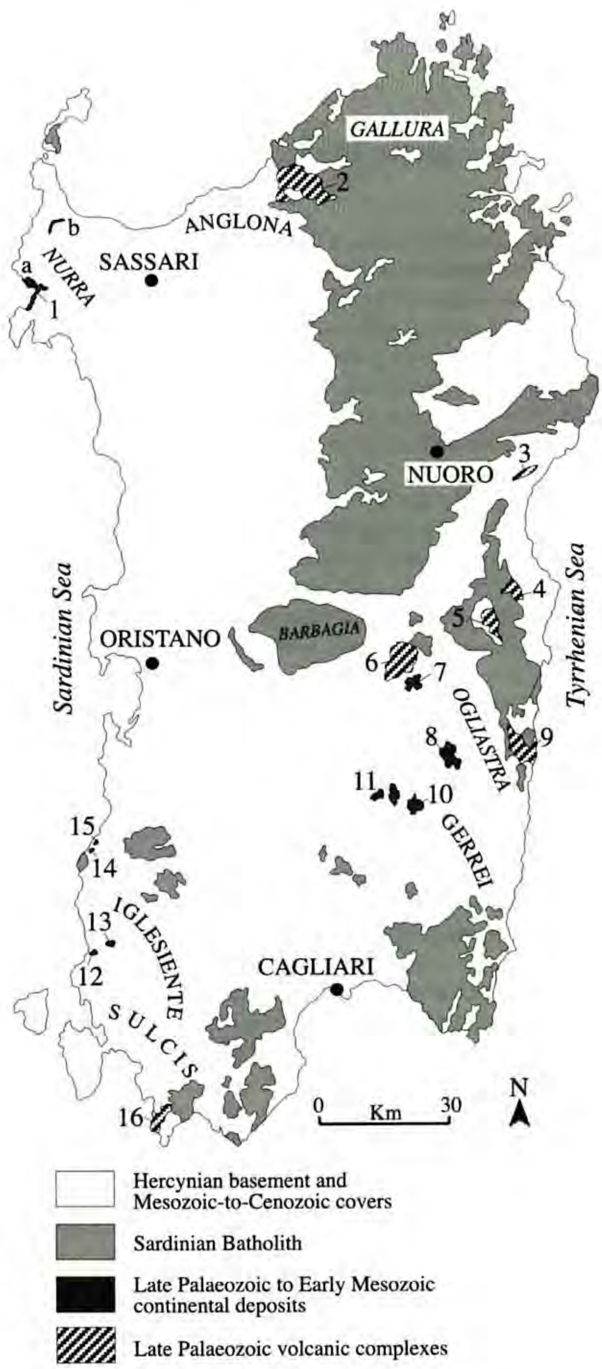

Fig. 1 – Location of the main late- to post-Hercynian sedimentary and volcanic sequences of Sardinia. 1. a: Lu Caparoni-Cala Viola Basin; b: Monte Santa Giusta. 2. Li Reni-Azzagulta. 3. Galtellì. 4. Baunei. 5. Villagrande-Strisaili. 6. Monte Perdedu-Monte Alastria. 7. Seui-Seulo. 8. Perdasdefogu. 9. Monte Ferru di Tertenia. 10. Escalaplano. 11. Lago di Mulargia. 12. Guardia Pisano. 13. San Giorgio. 14. Punta Acqua Durci. 15. Tuppa Niedda. 16. Teulada.

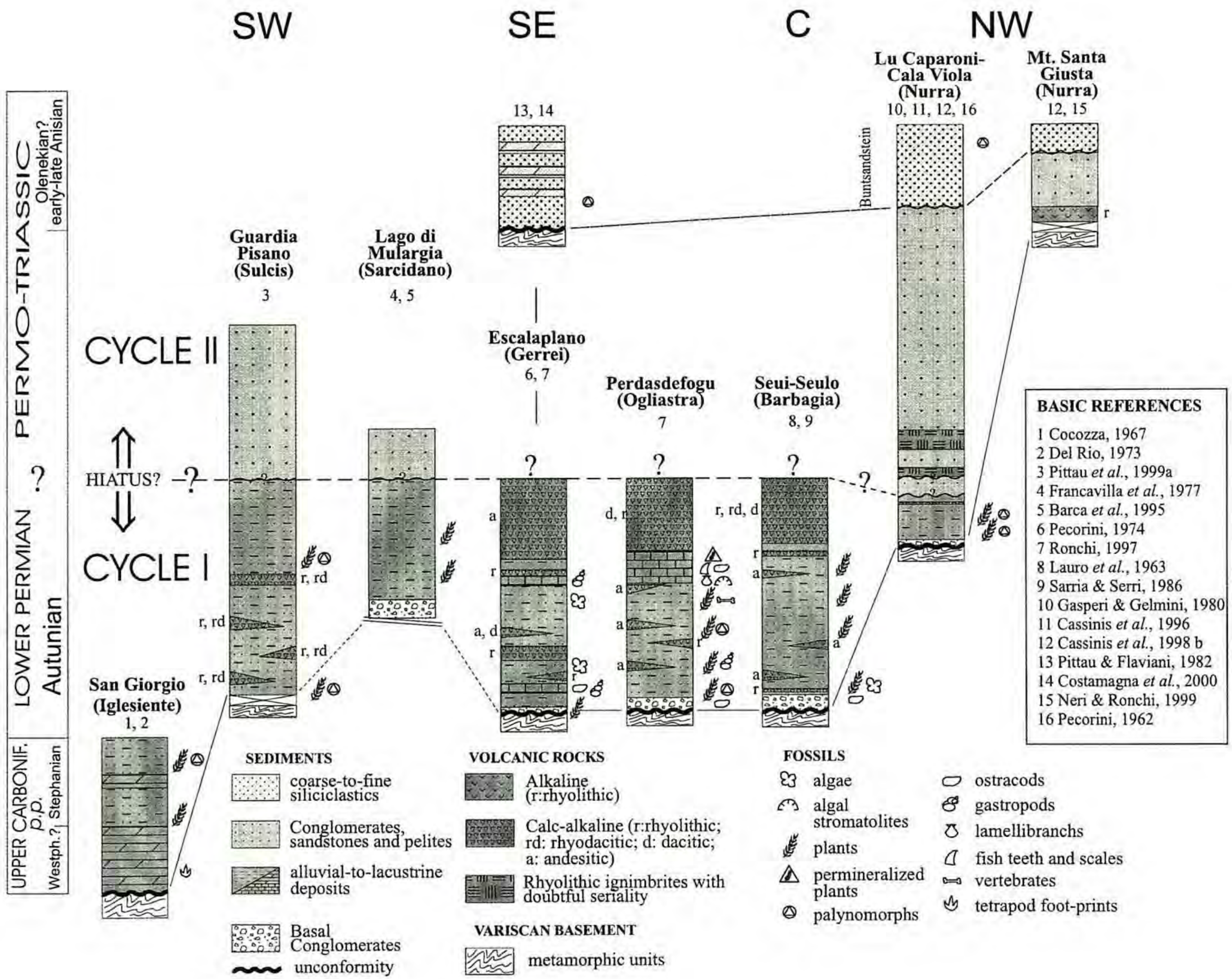

Fig. 2 – Correlation scheme among the Late Paleozoic volcano-sedimentary sequences of Sardinia. Vertical distances are not time- or thickness-related.

A clear unconformity between the two aforementioned cycles has not so far been observed at any of the previously cited basin localities. However, the abrupt facies changes in the two volcanosedimentary sequences suggest fundamental changes both in the structural regime and in the depositional environment. Furthermore, a marked angular unconformity between the Autunian and the "Saxonian-Thuringian" is recorded in various west-Mediterranean continental regions (*e.g.* Broutin *et al.*, 1994) from which the primary Corsica-Sardinia block moved subsequently towards the present-day Tyrrhenian.

According to some authors (*e.g.* Cassinis *et al.*, 1996, 1998a; Fontana *et al.*, this volume), the Lu Caparoni-Cala Viola Permian-Triassic detrital sequence ends with Buntsandstein-type deposits of early-to-late Anisian age (Pittau, 1999). These can be correlated by lithofacies and microfloristic associations with the Cugiareddu-drilled red beds (NW Sardinia) (Pittau Demelia & Flaviani, 1982) and with the southwest Campumari and southeast Escalaplano outcrops (Pittau Demelia & Del Rio, 1980; Pittau *et al.*, 1999b; Costamagna *et al.*, 2000).

STRATIGRAPHICAL CONSTRAINTS

Although the Late Paleozoic troughs of Sardinia developed in a similar intracontinental scenario, marked by a basin-and-swell landscape, their respective sequences greatly differ in thickness, extent and facial development. An extensional regime governed all these troughs, producing block-faulting and transtensional/transpressional movements. The environmental conditions of each basin were typical either of small lakes or of swamp ponds, from some centimetres to some metres in deep, and periodically exposed.

Hereafter, a brief stratigraphic synthesis of the most significant Upper Paleozoic basins of Sardinia are outlined. In Fig. 2, the investigated Upper Carboniferous to Middle Triassic successions of the island are shown.

First cycle

The first cycle is generally ascribed to "Autunian" times, apart from the (?)Westphalian C-Stephanian deposits which have been discovered only in southwestern Sardinia

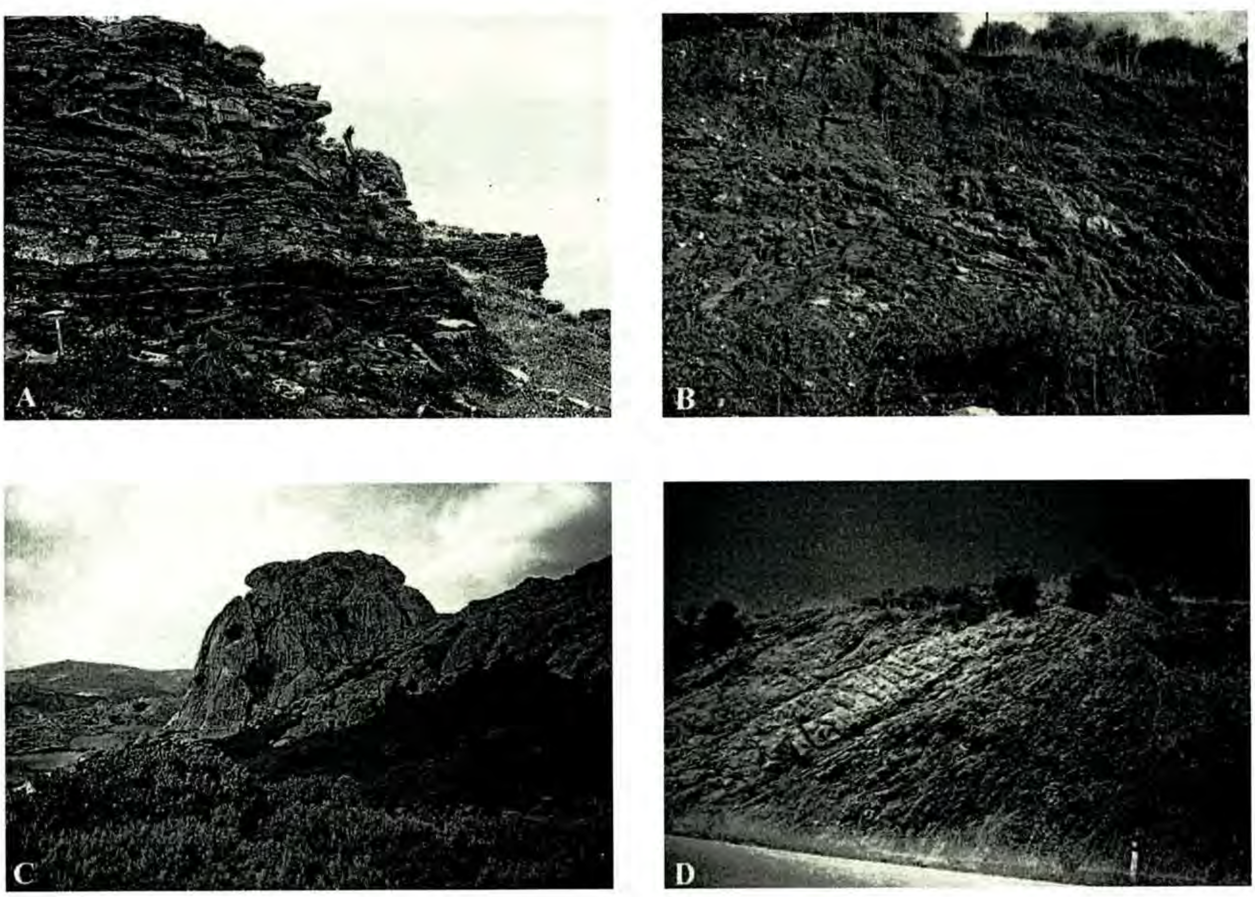

Fig. 3 – Representative examples of the first-cycle sequences in Sardinia. A: alternating dolostones, limestones and silicified carbonates (Perdasdefogu Basin, Ogliastra); B: coarse-to-fine red siliciclastics (Basal Conglomerate) disconformably overlying the Variscan basement (Escalaplano Basin, Gerrei); C: the subvolcanic rhyolitic dome of Monte Taddì (Seui Basin, Barbagia di Seulo); D: Lower Permian coarse to fine clastics in tectonic contact with the Variscan basement (Lago di Mulargia, Sarcidano).

(San Giorgio Basin, Iglesiente and Tuppa Niedda, Arburese). This cycle normally begins, in all the troughs, with more or less coarse basement breccias (Basal Conglomerates) unconformably overlying the Cambrian to Lower Carboniferous Variscan substratum (Fig. 3b, d). Alluvial-to-lacustrine, fossiliferous, mainly dark, shaly-to-sandy sediments follow, together with frequent volcanic (lavas) and volcaniclastic products (ignimbrites, pyroclastic fall deposits and breccias).

Locally, intercalations of conglomerates, lacustrine carbonates (up to 70 m in the Perdasdefogu Basin) with frequent silicification phenomena, anthracite deposits and pedological as well as diagenetic nodules occur (Ronchi, 1997 *cum bibl.*) (Fig. 3a). From most of these basins, *i.e.* Lu Caparoni-Cala Viola (Pecorini, 1962; Gasperi & Gelmini, 1980), Seui-Seulo (Lauro *et al.*, 1963; Sarria & Serri, 1987), Lago di Mulargia (Francavilla *et al.*, 1977; Barca *et al.*, 1995a), Guardia Pisano (Barca *et al.*, 1992; Pittau *et al.*, 1999a) and from other minor outcrops (Barca *et al.*, 1995b), the finer laminated sediments yielded abundant macrofloras and palynomorphs. In the Perdasdefogu Basin, other significant paleontological data such as permineralised plants, amphibians, algal stromatolites, ostracods and fish remains have also been highlighted. During the last century, many studies on the "Carboniferous-Permian" Sardinian macrofloras were performed (*e.g.* Comaschi Caria, 1959; Spano, 1976). Recently, according to Broutin (in Ronchi *et al.*, 1998; Broutin & Ronchi, 1999) the palynological assemblage of Perdasdefogu is qualitatively and quantitatively similar to that of Lu Caparoni Fm.; he relates these forms to the "A3" Biozone of Doubinger (1974) for the "Autunian" of Europe. Furthermore, according to Barca *et al.* (1992) and Pittau *et al.* (1999a), the well-preserved microflora of Guardia Pisano allow correlation with the Stephanian-Autunian of Western Europe, the Wolfcampian of the North American Midcontinent and the Ghzelian-Asselian of the Donetz and Ural basins.

In Sardinia, the first cycle generally ended with a major deposition of volcanic rocks, represented by rhyolitic and dacitic ignimbrites, lava flows and domes (Traversa, 1979; Lombardi *et al.*, 1974; Cortesogno *et al.*, 1998). Widespread magmatic basins also developed in various parts of Sardinia, where the sedimentary record appears reduced or missing. As in other western Mediterranean regions, the first-cycle fossil content shows that it ranged from the Late Carboniferous to the first part of the Early Permian; this age assessment is also confirmed by the calcalkaline volcanics found within and/or on top of these sedimentary sequences. However, we cannot exclude the possibility that this igneous activity continued up to the beginning of the Late Permian, as some radiometric dating may suggest (see *e.g.* Cozzupoli *et al.*, 1971; Edel *et al.*, 1981b; Tab. 1). This cycle is unconformably overlain by post-Autunian to Middle Triassic clastics in Nurra (and dubiously in Sarcidano and Sulcis) and by both Middle Jurassic and Eocene deposits in central to southeastern Sardinia (Barbagia, Ogliastra and Gerrei).

Second cycle

The second cycle consists of red azoic siliciclastic deposits (the so-called "Verrucano sardo" *Auct.*), which are mainly widespread in the Lu Caparoni-Cala Viola area (Nurra in northwestern Sardinia – Fig. 4a). In southeastern (Lago di Mulargia in Sarcidano – Fig. 4b) and southwestern Sardinia (Guardia Pisano in Sulcis) there are also similar, but less thick, alluvial coarse-grained deposits, which still require comparison with the Nurra sequence. Their chronostratigraphical attribution is hampered by the lack of fossils and the absence of reliable radiochronometric data (see Table 1).

Furthermore, the age of the Lu Caparoni-Cala Viola Basin (Nurra), alluvial to coastal plain sequence is still a matter of debate. In this area, the first-cycle Autunian plant-bearing sediments, *i.e.* the Lu Caparoni Formation *Auct.* is (disconformably?) overlain by mature white quartz-conglomerate fan deposits (unit 1 *sensu* Gasperi & Gelmini, 1980). The above units, presently both ascribed to the same cycle (Cassinis & Ronchi, 1997; Fontana *et al.*, this volume) are probably disconformably capped (without any clear evidence so far) by a fining-upwards sequence, up to some 200 m thick (units 2 and 3 of Gasperi & Gelmini, 1980), which is in turn followed by some 50 m of Buntsandstein-like deposits. This last portion (unit 4 *sensu* Gasperi & Gelmini, 1980), which can clearly be correlated to the Buntsandstein of Provence, begins with a metres-thick quartz conglomerate. In this context, in fact, significant boreholes in Permian and Triassic deposits were drilled a little to the east of Cala Viola (Lotti, 1931) and north of this area, near to Cugiareddu (Pomesano Cherchi, 1968), helping to unravel the chronostratigraphical and depositional problems. The whole fossil-barren red beds have historically been attributed to the second, Permian-Triassic cycle (Vardabasso, 1966; Gandin *et al.*, 1977; Gasperi & Gelmini, 1980; Fontana *et al.*, 1982).

The presence of some macroflora imprints (*Equisetum* cfr. *mougeotii*) led Pecorini (1962) to attribute the top of the aforementioned sequence to a doubtful Early Triassic age. Recently, Pittau (1999) ascribed these deposits to an ?Olenekian-early Anisian to late Anisian interval on the basis of lithostratigraphical correlations with similar sporomorph-bearing rocks in the Cugiareddu Well (see also Pittau Demelia & Flaviani, 1982). Concerning the stratigraphy as well as the petrography of sediments and volcanic rocks of the whole sequence, they have been analysed in detail by Cassinis *et al.* (1996, 1998b) and Fontana *et al.* (this volume).

At Monte Santa Giusta (Nurra), about 60 metres thickness of clastic deposits of uncertain age are poorly exposed at the base of the Muschelkalk and Keuper sequence (Fontana *et al.*, this volume). These authors correlates this succession with the Lu Caparoni-Cala Viola red beds (see also the inferred correlation made by Gasperi & Gelmini, 1980), explaining its reduced thickness as related to the structural height of the basement or to a slow subsidence rate. Acidic welded tuffs with alkaline affinities crop out at the base of the cited sediments (Lombardi *et al.*, 1974, Cassinis *et al.*, 1996; Cortesogno & Gaggero, 1999b). In a recent study, mainly dealing with the Muschelkalk and Keuper transgressive deposits, Carrillat *et al.* (1999) attributed the whole siliciclastic succession at the base of Monte Santa Giusta to a Lower Triassic to Anisian interval.

MAGMATISM

The voluminous late- to post-Variscan magmatic products, cropping out in Sardinia over more than 8000 km^2, have always been considered a decisive tool in the better understanding of the geological evolution of the island during the Late Paleozoic.

For this reason, studies dealing with this issue are numerous (*e.g.* Vardabasso, 1939; Traversa, 1979; Lombardi *et al.*, 1974; Atzori & Traversa, 1986; Cortesogno *et al.*, 1998). Also, geochronological and geochemical studies were carried out recently by various authors on the Permian-Triassic dykes, cutting through the Variscan intrusives and metamorphic basement all over the eastern part of the island (Atzori & Traversa, 1986; Vaccaro *et al.*, 1991; Traversa & Vaccaro, 1992).

The volcanic products pertaining to the first cycle show a calcalkaline affinity, which has been variously interpreted by some researchers as the consequence of a poorly constrained active margin with a volcanic arc, located in the Ligurian-Provencal area (Cabanis, in Broutin *et al.*, 1994, and references therein), and by others as due to partial melting processes at the mantle-crust interface (Cortesogno *et al.*, 1998). This magmatism affected all the Early Permian basins of Sardinia, although it consists of large volumes mainly distributed in extrabasinal areas, particularly on the eastern side of Sardinia. From north to south, it gave birth to broad volcanic complexes or calderas, like those of Li Reni-Azzagulta in Gallura (Traversa, 1979 *cum bibl.*; Del Moro *et al.*, 1996), Monte Perdedu-Monte Alastria in Barbagia di Seui (Cozzupoli & Lombardi, 1969; Cozzupoli *et al.*, 1971), Monte Ferru di Tertenia in Ogliastra, and Capo Teulada in Sulcis (Lombardi *et al.*, 1974).

Within the Lower Permian basins, acidic to intermediate volcaniclastic deposits as well as rhyolitic to rhyodacitic domes, lava flows and dykes are always associated with siliciclastic deposits (see the recent works of Cassinis *et al.*, 1998a; Cortesogno *et al.*, 1998; Cortesogno & Gaggero, 1999a). Commonly, the first-cycle volcanic activity began with acid explosives and major volumes of andesitic lavas with columnar joints and chaotic breccias (Seui, Escalaplano, Perdasdefogu), followed locally by dacitic lavas (Perdasdefogu) and dykes, and rhyolitic dykes and ignimbrites (Seui, Perdasdefogu). In the Seui Basin, spectacular updoming of dacitic to rhyolitic magmatic bodies also occurs (Calzia *et al.*, 1999) (Fig. 3c). In some sequences, like those of Escalaplano (Pecorini, 1974; Ronchi, 1997), the volcanic products largely dominate the strictly sedimentary units, but mostly they are intercalated with the sediments and occur as lens-shaped volcanic breccias, stratified cinerites and tuffs, hyaloclastites, lava plugs and sills. Stratigraphical and recent radiometric data generally place this volcanic activity during the Early Permian.

In contrast, an alkaline affinity was stated for the first

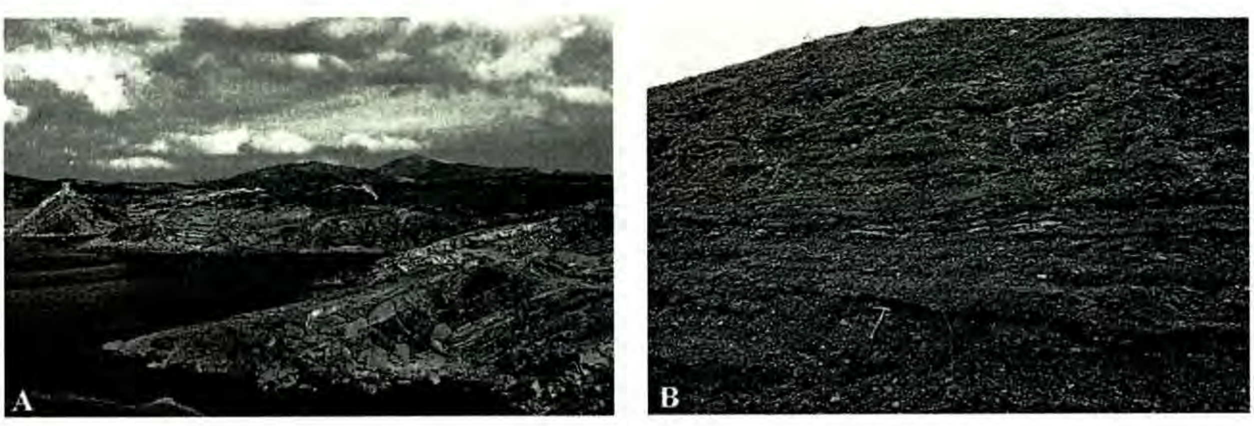

Fig. 4 – Two examples of the second-cycle continental deposits. A: the Permo-Triassic siliciclastic sequence near Cala Viola (Nurra); B: channelled red alluvial deposits near Lago di Mulargia (Sarcidano).

REGION	SITE	AGE	ISOTOPIC METHOD	ROCK TYPE	AUTHOR
NW SARDINIA (Nurra)	Mt. Santa Giusta	197+/- 6Ma	K-Ar (WR)	alkaline rhyolitic ignimbrite	Lombardi *et al.*, 1974
		204+/- 6Ma	"		
		297+/-9 Ma	K-Ar (Bt)	rhyolitic ignimbrite	Edel *et al.*, 1981
		244+/-9 Ma	K-Ar (Fs)		
		296+/-8 Ma	K-Ar (Bt)		
N-NE SARDINIA (Gallura-Anglona)	Azzagulta-Li Reni	247+/- 7Ma	K-Ar (WR)	rhyolitic ignimbrite	Edel *et al*, 1981
		211+/- 6Ma	K-Ar (Fs)		
		226+/- 6Ma	K-Ar (Fs)		
		258+/- 7Ma	K-Ar (Fs+Q)		
		267+/- 7Ma	K-Ar (Fs+Q)		
		262+/- 9Ma	K-Ar (Q+ Fs)		
		249+/- 7Ma	K-Ar (Fs+Q)		
		260+/- 9Ma	K-Ar (Fs+Q		
	Coghinas Valley-Aggius	286+/-3Ma	Rb-Sr (Ms)	arkosic sandstone	Del Moro *et al.*, 1996
		288+/-3Ma	Rb-Sr (Bt)	"	
		288+/-11Ma	Rb-Sr (WR)	rhyolitic ignimbrite	
		268+/-8 Ma	Rb-Sr (WR)	intermediate dyke	Vaccaro *et al.*, 1991
		289+/-9 Ma	"		
(Goceano)	Concas	273+/-9 Ma	Rb-Sr (WR)	dyke	Vaccaro *et al.*, 1991
		282+/-9 Ma	Rb-Sr (WR)		
		280+/-9 Ma	Rb-Sr (WR)		
		210+/-115 Ma	Rb-Sr (WR)		
	Mt. Lerno	281+/-10 Ma	Rb-Sr (WR)	dyke	Vaccaro *et al.*, 1991
(Baronie)		291+/-9 Ma	Rb-Sr (WR)	dyke	Vaccaro *et al.*, 1991
CENTRAL-E SARDINIA (Barbagia)	Sorgono	298+/-9 Ma	Rb-Sr (WR)	dyke	Vaccaro *et al.*, 1991
	Galtellì (Orosei)	268+/-10	K-Ar (WR)	rhyolitic ignimbrite	Cozzupoli *et al.*, 1984
		248+/-9	"		
		275+/-11	"		
		238+/-11	"		
		280+/-7Ma	"		
	Mt. Perdedu-Mt. Alastria	260+/-5Ma	K-Ar (WR)	quartzlatitic subvolcanite	Cozzupoli *et al.*, 1971
		212+/-5Ma	"		
	Seui-Seulo	255+/-5Ma	"	rhyolitic ignimbrite	
		250+/-5Ma	"		
		262+/-5Ma	"		
		255+/-6Ma	"	Trattalas Diorite	
		265+/-5Ma	"		
		220+/-4Ma	"	rhyolitic lava	
		225+/-5Ma	"		
		250+/-6Ma	"		
		255+/-8Ma	"		
		259+/-7Ma	K-Ar (Pl)	ignimbrite	Edel *et al.*, 1981
		261+/-8Ma	K-Ar (Fs)	ignimbritic tuff	
(N Ogliastra)	Talana-Villagrande Strisaili	301+/-8Ma	K-Ar (Bt+Cl)	dyke	Edel *et al.*, 1981
		281+/-10 Ma	K-Ar (Pl)	ignimbrite	
		308+/-11 Ma	K-Ar (Pl)	"	
		235+/-7 Ma	K-Ar (WR)	rhyolitic ignimbrite	Lombardi *et al.*, 1974
		228+/-5 ma	"		
		251+/-5 Ma	"		
		289+/-9 Ma	Rb-Sr (WR)	intermediate dyke	Vaccaro *et al.*, 1991
CENTRAL-SE SARDINIA (S Ogliastra)	Baunei	281+/-10 Ma	K-Ar (Pl)	ignimbrite	Edel *et al.*, 1981
		291+/-11 Ma	K-Ar (Pl)		
		137+/-9 Ma	K-Ar (Fs+Q)		
		290+/-5Ma	K-Ar (Cl+Bt)		
	Perdasdefogu	220+/-5Ma	K-Ar (WR)	quarzlatitic lava (dacite)	Lombardi *et al.*, 1974
		218+/-6Ma	"		
	Escalaplano	230+/-5Ma	K-Ar (WR)	latitic lava (andesite)	Lombardi *et al.*, 1974
		197+/-7Ma	"		
		295+/-8Ma	K-Ar (Bt)	andesite	Edel *et al.*, 1981
	Mt.Ferru di Tertenia	223+/-7Ma	K-Ar (WR)	rhyolitic ignimbrite	Lombardi *et al.*, 1974
		211+/-8Ma	K-Ar (Fs+Q)	ignimbrite	Edel *et al.*, 1981
		252+/-8Ma	K-Ar (Pl)		
SW SARDINIA (Sulcis)	Capo Teulada	260+/-5Ma	K-Ar (WR)	rhyodacitic lava	Lombardi *et al.*, 1974
		228+/-6Ma	"	quartzlatitic lava	
	Guardia Pisano	295+/-Ma	SHRIMP	rhyolitic lava	Pittau et al., 1999

Table 1 – Radiometric data for the Late Paleozoic volcanic rocks and dykes of Sardinia.

time by Lombardi *et al.* (1974) for the volcanic deposits intercalated within the second sedimentary cycle of Nurra.

This change to a typically anorogenic character has been related by some authors to a post-Variscan global plate reorganisation (Cortesogno *et al.*, 1998). Such straightforward affinity is partly confirmed by recent works, but only for Monte Santa Giusta volcaniclastics (Cassinis *et al.*, 1996; Cortesogno & Gaggero, 1999b). In contrast, the porphyric pebbles included in unit 2 of Lu Caparoni-Cala Viola Basin (Gasperi & Gelmini, 1980), as well as the pyroclastic rocks cropping out at the base or intercalated in the same part of the second cycle, show a calcalkaline petrographic affinity which could be «either primary or induced by important secondary depletion of mobile elements» (Cortesogno & Gaggero in Cassinis *et al.*, 1996 and 1998b).

GEOCHRONOLOGICAL DATA AND AGE PROBLEMS

Many isotopic age analyses have been performed on the Upper Paleozoic volcanic rocks (mostly using K/Ar and Rb/Sr methods) at various sites in Sardinia (Cozzupoli *et al.*, 1971, 1984; Lombardi *et al.*, 1974; Edel *et al.*, 1981; Del Moro *et al.*, 1996). These rocks have shown differing ages which span from the Late Carboniferous to the Late Triassic, and these discrepancies probably derive from alteration processes (Table 1).

Even if Sardinia seems not to have been strongly affected by Alpine orogenic events, the calculated ages spread over a time span of more than 150 Ma (see also the schematic tables in Fontana *et al.*, 1982; Beccaluva *et al.*, 1985; Gelmini, 1985). Furthermore, it appears very difficult to compare data gathered using different isotopic techniques and based on whole rock or single mineral analyses. In the opinion of some authors, the late- to post-Variscan volcanic activity developed in two different periods: 250-260 Ma and 210-230 Ma for Fontana *et al.* (1982), around 250 and 280 Ma for Gelmini (1985), and 340-275 Ma and 275-235 Ma for Broutin *et al.* (1994).

In contrast with these authors, Edel *et al.* (1981) stated that the Upper Paleozoic volcanic products are almost contemporaneous and referable to an older calcalkaline cycle (late Westphalian-early Stephanian) and a younger one with alkaline affinities (Stephanian).

In the light of the modern reliability criteria, quality data appear to be those gathered by Cozzupoli *et al.* (1984) on the Galtellì ignimbrites (284±15 Ma), Del Moro *et al.* (1996) on the Gallura rhyolites (288±11 Ma), and Pittau *et al.* (1999a) on the rhyolitic-rhyodacitic lavas intercalated in the Guardia Pisano sequence (295±5 Ma, SHRIMP). These measurements are in good agreement with macro- and microfloral associations, which point to a basal Permian age for the bulk of the first-cycle sequences. Geochronological Rb/Sr and Ar-Ar age determinations on whole rocks and mineral separates were also performed on a great number of Late Paleozoic dykes throughout the island (Vaccaro *et al.*, 1991).

For these authors and for Atzori and Traversa (1986), this type of magmatism in Sardinia was largely heterochronous, both preceding and following the Permian effusive activity. The dyke activity probably occurred during two phases: the first one, represented by calcalkaline, normal to high potassium and peraluminous anatectic lithotypes, developed during two different stages, at 298-289 Ma and 270 Ma; the second phase, which is represented almost exclusively by basaltic products and rare peraluminous rhyolitic dykes (subalkaline and transitional types are found in central-southern Sardinia, while alkaline types are restricted to the north), shows ages of approximately 230 Ma.

The geochronology of the second cycle is more doubtful and, conversely with respect to the preceding cycle, is not controlled by a paleontological record. As previously stated, the alkaline affinity of Monte Santa Giusta ignimbrites prompts comparison with the Late Permian ("Saxonian-Thuringian") activity developed in other western Mediterranean areas, like Corsica and Provence.

As shown, the evolution and timing of the Late Paleozoic to Middle Triassic sequences of Sardinia may not be properly unravelled until new isotopic dating measurements or paleontological discoveries are made.

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STRATIGRAPHIC ARCHITECTURE AND COMPOSITION OF THE PERMIAN AND TRIASSIC SILICICLASTIC SUCCESSION OF NURRA (NORTHWESTERN SARDINIA)

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Key words – Stratigraphy; red beds; arenite composition; Sardinia; Permian-Triassic.

Abstract – This work represents a preliminary revision of the terrigenous Permian and Triassic stratigraphic successions of the Nurra region (northwestern Sardinia). On the basis of litho-stratigraphical, sedimentological and petrographic evidence, supported by scattered and rare biostratigraphical data, the whole post-Variscan succession has been subdivided into the following units: I) the "Punta Lu Caparoni Formation" (PLC), made by the traditional alluvial-to-lacustrine deposits and upwards by a whitish succession which corresponds to the unit 1 of Gasperi & Gelmini, 1980; II) a volcanic-bearing unit of problematic geochemistry, scarcely cropping out in the Lu Caparoni-Cala Viola area and well-documented in the Monte Santa Giusta section, where shows an alkaline composition; III) a clastic unit of reddish conglomerates, sandstones and minor pelites, rich in volcanic-derived clasts; as in the Monte Santa Giusta, this unit represents a fining-upwards sequence, evolving from alluvial fan/braided river settings to meandering river and coastal plain deposits, and corresponds to units 2 and 3 of Gasperi & Gelmini (1980); IV) an upper clastic unit, separated from the underlying deposits by a marked disconformity. It comprises a basal quartz conglomerate grading upwards into reddish sandstones and minor pelites, showing alternating fluvial and tidal influx.
Scarce biostratigraphical data allow us to relate the base of the investigated siliciclastic succession to the Early Permian ("Autunian" Auct.) and the top to an Olenekian (?)-early Anisian to late Anisian time, but do not allow a chronostratigraphical classification, unit by unit, of the whole sequence and the evaluation of the duration of hiatuses at the main unconformities.

Parole chiave – Stratigrafia; *red beds*; composizione delle areniti; Sardegna; Permiano-Triassico.

Riassunto – Questo lavoro fornisce una revisione preliminare della successione permiana e triassica della Nurra dal punto di vista stratigrafico, sedimentologico e composizionale. Sulla base di nuovi dati di campagna, gli autori propongono di suddividere la successione in quattro unità principali: I) la "Formazione di Punta Lu Caparoni" (PLC), che oltre ai tradizionali depositi alluvio-lacustri comprende, per la prima volta, la sovrastante unità clastica, di colore biancastro, già denominata come unità 1 da Gasperi & Gelmini (1980); II) uno o più corpi vulcanici, rappresentati da ignimbriti e tufi a composizione riolitica, spesso molto alterati. Questi prodotti affiorano sia ad un livello stratigrafico imprecisato tra la PLC e la successione rossastra permiana e triassica di Lu Caparoni-Cala Viola e sia, benchè la correlazione tra i rispettivi episodi vulcanici risulti tuttora incerta, alla base della sezione di Monte Santa Giusta; III) la successione soprastante, che corrisponde alle unità 2 e 3 di Gasperi & Gelmini (1980), è costituita da conglomerati e arenarie fluviali, con abbondanti clasti di origine vulcanica. In base ai dati forniti dalla sezione di Monte Santa Giusta, per quanto condensata e lacunosa, si può ritenere che le unità 2 e 3 di Gasperi & Gelmini (1980) costituiscano un'unica sequenza fining-upward, con transizione da facies braided a facies meandriformi e quindi di piana costiera; IV) l'unità silicic1astica superiore (unità 4 di Gasperi & Gelmini, 1980) è discordante sulla successione precedente; la discordanza è marcata da un banco conglomeratico ricco in ciottoli di quarzo provenienti dal basamento metamorfico. Verso l'alto passa ad arenarie a stratificazione incrociata bipolare e successivamente a siltiti ed arenarie fini con influssi alternativamente fluviali e tidali.
Il limite con i soprastanti depositi del Muschelkalk è graduale e parzialmente visibile nella sezione di Monte Santa Giusta, mentre a Cala Viola la successione si trova in contatto tettonico con i sedimenti del Keuper.
I dati composizionali indicano contributi prevalenti dal basamento metamorfico e dalle vulcaniti permiane. Con la deposizione dell'unità sommitale (IV), arenarie e conglomerati mostrano un'elevata maturità tessiturale e composizionale, con un consistente incremento nella quantità di quarzo.
Sulla successione in esame sono disponibili solo rari dati cronostratigrafici. La parte inferiore della "Formazione di Punta Lu Caparoni" è datata al Permiano inferiore ("Autuniano" *Auct.*) in base al contenuto macrofloristico e palinologico; la parte superiore dell'unità IV a Cala Viola ha fornito flore mal conservate (*Equisetum* cfr. *mougeotii*). Le associazioni palinologiche rinvenute nel pozzo Cugiareddu, entro depositi clastici probabilmente correlabili con l'unità IV, indicherebbero la presenza dell'Olenekiano(?)-Anisico inferiore e dell'Anisico superiore.

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INTRODUCTION

In the Nurra region (northwestern Sardinia), a post-Variscan continental succession, represented by terrigenous deposits intercalated with volcaniclastic products, is discontinuously exposed (Fig. 1). It has been the subject of a number of studies: Lotti (1931), Oosterbaan (1936), Moretti (1959), Pecorini (1962), Vardabasso (1966), Gasperi & Gelmini (1980), Fontana *et al.* (1982), Gelmini (1985), Cassinis *et al.* (1996, 1998a, 1998b) and Neri *et al.* (1999). The present study concerns the stratigraphy, sedimentology and composition of such a red-bed succession, in the Alghero area, between Punta Lu Caparoni to the north and Cala Viola to the south (Fig. 1).

Gasperi & Gelmini (1980) provided the first lithostratigraphical classification of the succession, subsequently accepted by most researchers (*e.g.*, Cassinis *et al.*, 1996, 1998a, 1998b). According to their classification, the whole post-Variscan and sub-Muschelkalk (using the terminology of Owen, 1987) terrigenous succession has been divided into a thin, discontinuous formation (Punta Lu Caparoni Fm.), Early Permian in age, and in a terrigenous sequence, 200-300 m thick. The terrigenous sequence is in turn divided into four informal units: unit 1 comprises mainly quartz conglomerates and coarse whitish sandstones; unit 2 consists of reddish conglomerate and sandstones, with clasts of volcanic and metamorphic rocks; unit 3 consists of grey sandstones with tabular and trough cross-stratification, alternating with upward-increasing red siltstones; unit 4 begins with coarse quartz conglomerate and grades into fine to medium-grained sandstones.

Fig. 1 – Location map of the main outcrops of the Permian and Triassic deposits in the Nurra region.

The vertical transition from these siliciclastic deposits into the Middle Triassic carbonates (Muschelkalk) is not exposed over a great part of the investigated area, mainly due to tectonics, with the remarkable exception of Monte Santa Giusta (Carrillat *et al.*, 1999; Neri & Ronchi, 1999) and of the Cugiareddu borehole (Pomesano Cherchi, 1968).

The biochronostratigraphical data are poor, and limited to the basal and upper portions of the succession; the base (Punta Lu Caparoni Fm. *s.s.*) yields plant remains and palynomorphs of Early Permian ("Autunian") age. The upper portions of unit 4 (*sensu* Gasperi & Gelmini, 1980) contain a scattered and poorly preserved paleoflora (*Equisetum* cfr. *mougeotii*, reported by Pecorini, 1962) indicating a presumed Olenekian(?)-early Anisian to late Anisian, on the basis of two differing palynological associations of the top of the Permian and Triassic sequence, from the Cugiareddu well (Pittau Demelia & Flaviani, 1982; Pittau, 1999). A considerable time-span is encompassed by these two age-data points; we have tried to achieve new biostratigraphical data through a systematic sampling of the rare grey pelite horizons intercalated into the sequence, for palynological study, but without appreciable results. Thus, the stratigraphic gap associated with the unconformity between units 3 and 4 of Gasperi & Gelmini (1980) cannot be estimated.

An attempt has been made by Cassinis *et al.* (1996, 1998a) and Neri *et al.* (1999) to correlate the Permian and Triassic succession of Nurra to the "Two Tectonosedimentary Cycles" model proposed by the Italian IGCP 203 Group (1986) for the South-Alpine area. Unfortunately, the poor biostratigraphical data and the fact that boundaries between units 1, 2 and 3 have not been observed in the field, prevent such a correlation. We need more data on bounding unconformities (real or presumed) and their ages.

However, research recently carried out by the authors may lead to a new lithostratigraphical subdivision for the

Permian and Triassic succession of the Nurra region, as documented in the following sections (note that we use the numbers 1, 2, 3 and 4 to indicate units of Gasperi & Gelmini, 1980 and roman numerals I, II, III and IV for new units from this work). Since an important sector of the study area (*i.e.*, the steep sea-cliff south and west of Punta Lu Caparoni) has not yet been investigated, we do not propose new formation names here.

STRATIGRAPHY

Punta Lu Caparoni section

According to the lithostratigraphy of Gasperi & Gelmini (1980), this section, about 45 m thick, consists, of two units, separated by a marked disconformity. They are the Punta Lu Caparoni Fm. and unit 1 of the Permian and Triassic terrigenous succession.

The first unit (Figs 2 and 3) begins with a poorly sorted conglomerate, about 1 m thick, mainly consisting of angular clasts derived from the local phyllitic basement.

Dark-grey laminated shales, siltstones and fine-grained

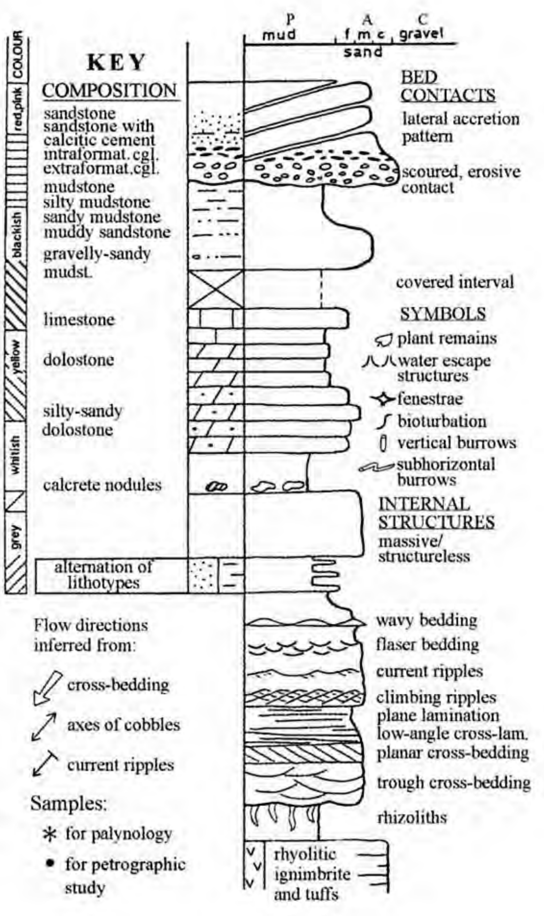


Fig. 2 – Legend of symbols for Figs 3, 5, 6 and 7.

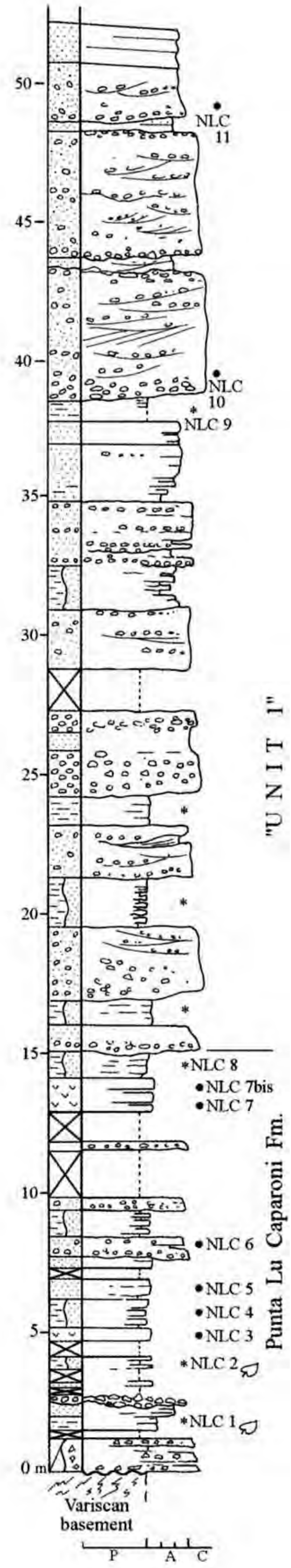
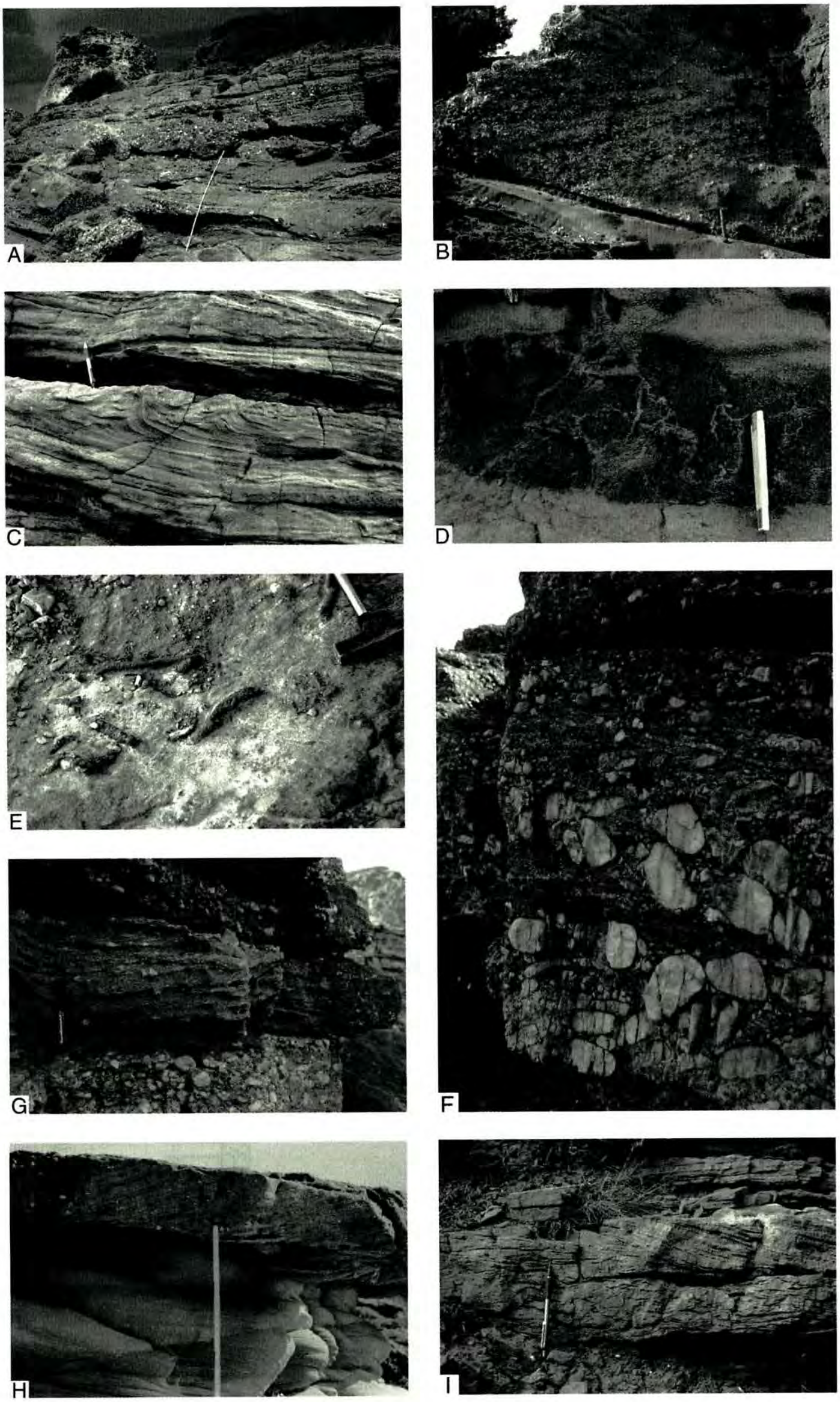


Fig. 3 – Stratigraphic section of Punta Lu Caparoni (symbols as in Fig. 2); Unit 1 and Lu Caparoni Fm. *sensu* Gasperi & Gelmini, 1980.


A
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sandstones are present upwards. Two main fossil-bearing horizons occur in this part of the section, about 5 m thick, yielding Autunian macrofloras and palynomorphs of Early Permian age (Pecorini, 1962; Broutin *et al.*, 1996, 1999).

The following tract of Punta Lu Caparoni Fm. (about 10 m thick) is discontinuously exposed. However, it does show some decimetre- to metre-scale conglomerate intercalations, with clasts of quartz and phyllites, alternating with sandstones and siltstones. Yellowish and grey massive fines, constituting decimetre- to metre-thick horizons with ferruginous crusts at the top, also occur: they are interpreted as kaolinised cinerites, and the crusts may represent pedogenic horizons.

Similar deposits also occur in the lower and middle part of the overlying unit 1 of Gasperi & Gelmini (1980), alternating with quartz-conglomerates and coarse sandstone bodies, increasing in thickness upwards. These coarse-grained bodies typically have scoured, erosional lower boundaries and marked lateral variations in thickness, mainly due to the deepness of scouring.

The strongly channelised bodies, separated by considerable amounts of fines, grade upwards into laterally-continuous conglomerate and coarse sandstone units, several metres thick, with trough cross-bedding. The uppermost part of the section is represented by a sandstone body showing a lateral accretion pattern.

Evidence of volcanic activity in these upper deposits is clearly represented by high percentage of embayed quartz and volcanic rock fragments, as well as by fine volcaniclastics.

Facies interpretation. A possible interpretation of such a vertical evolution of the facies pattern may be the following:

– A fluvial-lacustrine environment may be suggested for the lowermost part of the section, dominated by siltstones rich in plant remains. The basin was probably small in size; other deposits cropping out in the Nurra region and referred to as the Punta Lu Caparoni Fm. were probably laid down in separate, endoreic basins.

– Coarse sandstone to conglomerate bodies, progressively increasing in thickness, encroach on the sandy-pelite lacustrine deposits; each coarse-grained body has a disconformable base, so it is difficult to indicate a major-rank disconformity, separating the Punta Lu Caparoni Fm. *sensu* Gasperi & Gelmini (1980) from the overlying unit 1.

Finally, the tabular cross-bedded bodies may indicate the presence of a widely-distributed braided-river environment, possibly followed upwards by a meandering-river regime (Fig. 3). Unfortunately, we lack biostratigraphical data to support the continuity between the two above units.

Torre Bianca section

In the Torre Bianca area the deposits corresponding to unit 2 of Gasperi & Gelmini (1980) crop out with good exposures. A continuous tract more than 50 m thick can be observed, but its lower and upper boundaries are not exposed. Moreover, considerable tectonic deformation (faulting) separates this section from other outcrops in the area.

The section has not been measured in detail, but carefully examined and documented by photos and drawings of the main facies. Spectacular outcrops show the facies model of this tract, mainly dominated by braided river depositional settings (Figs 4A and 4B).

The main lithofacies are represented by conglomerates and conglomeratic coarse sandstones, with clasts deriving both from the "porphyric" volcanics and the metamorphic Variscan basement (quartz, minor phyllites). Depositional structures consist of metre-sized trough cross-bedding and, subordinately, tabular cross-bedding forming sets up to a metre thick; they probably record longitudinal bars (Fig. 4B).

Torre del Porticciolo and Cala Viola sections

The Torre del Porticciolo and Cala Viola sections (Figs 5 and 6) have been measured along the coastline between Cala del Turco and Cala Viola. Both sections may be subdivided into two units, separated by an impressive unconformity, and corresponding to units 3 and 4 of Gasperi & Gelmini (1980) respectively. Only the upper part of unit 3 crops out, and this makes defining the relationship with the underlying units problematic. Unit 4 crops out from the base, but its upper portion is cut by faults. In fact, this unit at Cala Viola is overlain, through a tectonic contact, by Keuper evaporites and dolomites.

The lower unit (unit 3) is represented by grey-green to

Fig. 4 – A: sedimentary facies in the Torre Bianca area. Cross-cutting conglomeratic bodies represent braided-river dominated setting; B: thick (about 1.20 m) conglomeratic body with planar, cross-bedding probably showing the frontal accretion of a longitudinal bar. Torre Bianca section; C: water escape structures on low-angle parallel lamination in the arenites of Torre del Porticciolo section. These deposits represent a laterally-accreted channel fill; D: reddish floodplain deposits containing rhizoliths in the upper part of Torre del Porticciolo section; E: cylindrical, featureless, variably oriented burrows on a stratigraphic surface in the uppermost part of a channel body, Cala Viola section; F: sheet-like massive conglomerate bank with well rounded, imbricated quartz cobbles underlying the base of the upper unit in the Torre del Porticciolo section; G: tabular cross-lamination in the granule-cobble-bearing arenites at the top of the conglomerate bank. Unit 4 in the Torre del Porticciolo section; H: planar cross-bedding in the lower part of unit 4 in the Torre del Porticciolo section; I: flaser and wavy-bedded body cut by a small cross-bedded (lateral accretion?) channel-fill. Upper part of the Cala Viola section.

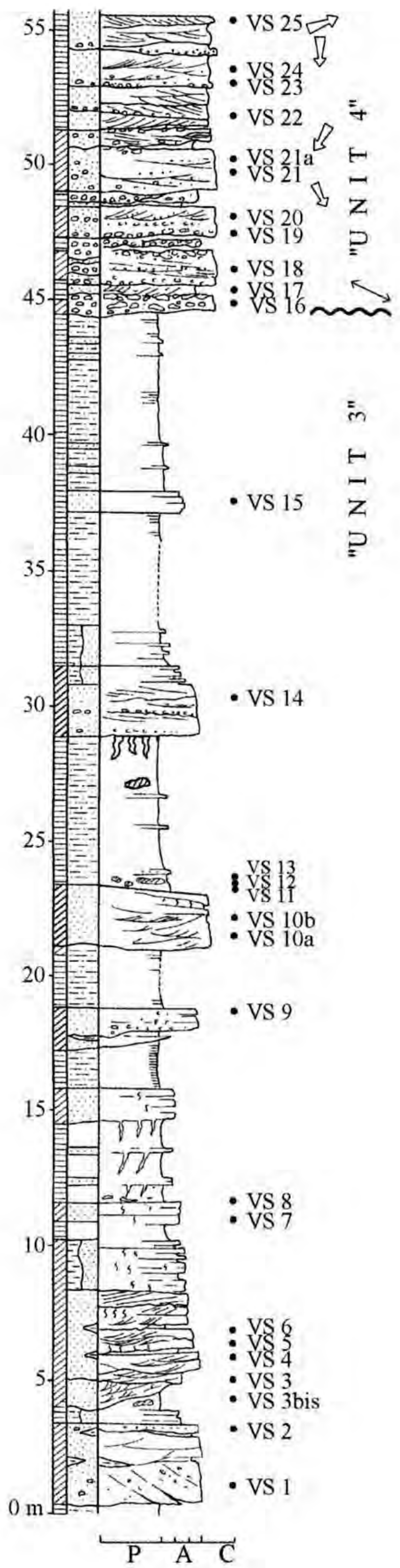

Fig. 5 – Stratigraphic section of Torre del Porticciolo (see Fig. 2 for legend); Units 3 and 4 *sensu* Gasperi & Gelmini, 1980.

brown channelised arenitic bodies decreasing in thickness and lateral continuity upwards in the section, interbedded with reddish pelites (Figs 5 and 6). Sandstones are commonly medium- to coarse-grained, with sparse quartz pebbles, and form lenticular channel-fills, with a marked erosional base cutting into structureless (with the exception of pedogenic features) reddish siltstones and pelites. Arenitic bodies show a complex internal organisation, with trough- and planar cross-bedding and minor, lens-shaped, intercalations of reddish pelites (Figs 5 and 6). From the outcropping base of the sections (in particular, in the Torre del Porticciolo area), the channel-fills show clinostratified sets indicating lateral accretion (Fig. 4C). Sets are 20-30 cm thick, dipping at 25°-30°, and are separated by thin red pelitic interlayers. The internal structures of sandstone beds within the channel-fills include trough cross-bedding and low-angle to parallel lamination, accompanied by climbing ripple cross-lamination and water-escape structures (Fig. 4C).

Sheet-like sandstones with scarce lateral persistence are also present, intercalated with the channel-fill bodies; they display planar cross-bedding and parallel lamination. These arenitic layers become more frequent in the upper part of the unit, where silty-pelitic bodies increase in thickness.

Colour mottling and other pedogenetic features such as rhizoliths and carbonate concretions are frequent in the pelitic intervals, and become more abundant in the thicker layers (Fig. 4D). Bioturbation, common both in red pelites and in fine- to medium-grained sandstones, is represented by horizontal and vertical tubes on the set surfaces (Fig. 4E).

Facies interpretation. The overall facies model may indicate a sinuous (meandering?) channel complex for the lower part of the section, at least in the Torre del Porticciolo area. The thick reddish pelite horizons intercalated between channel-fills are interpreted as flood plain deposits on which pedogenic transformations took place. Geometry and sedimentary structures suggest that sheet-like arenitic layers intercalated within siltstones are crevasse-splay deposits. The main feature of the upper part of the unit is the very low lateral continuity of arenitic bodies (ribbon-like channel-fills) and the prevalence of pelites.

The upper unit (unit 4) overlies unit 3 through a distinctly disconformable contact and a sharp change in facies. The lowermost part of the unit, 6-8 m thick, consists of a sheet-like conglomerate bank of regional extension, internally subdivided into decimetre- to metre-thick sets with erosional bases; conglomerates are usually matrix-supported, with well-rounded, imbricated (a-axis), cobble-grade clasts made almost exclusively of quartz (Fig. 4F). Planar and trough cross-bedded sets of microconglomerate and coarse sandstone are intercalated.

The conglomerate bank grades upwards into a unit, 5-6 m thick, of pebbly coarse sandstones; they show planar and trough cross-bedding, frequently with tangential basal contacts and opposite dips of the laminae (Figs 4G and 4H).

The upper part of unit 4 crops out only in the Cala Viola section (Fig. 6); it is represented by medium-grained reddish arenites, with subordinate pelites, forming metre-sized flaser-bedded sets, locally evolving upwards into horizontal to low-angle laminated or trough cross-bedded thin sandstone layers (Fig. 4I). Metre-thick fining-upwards channel-fills occur at different levels. The uppermost part of the succession contains numerous small, 30-50 cm thick, erosive sand bodies with little lateral persistence; they are frequently rich in tabular-shaped to rounded intraclasts.

Facies interpretation. Facies analysis suggests that the basal coarse-grained conglomerate sheet was deposited within a braided fluvial system, probably during a tecton-

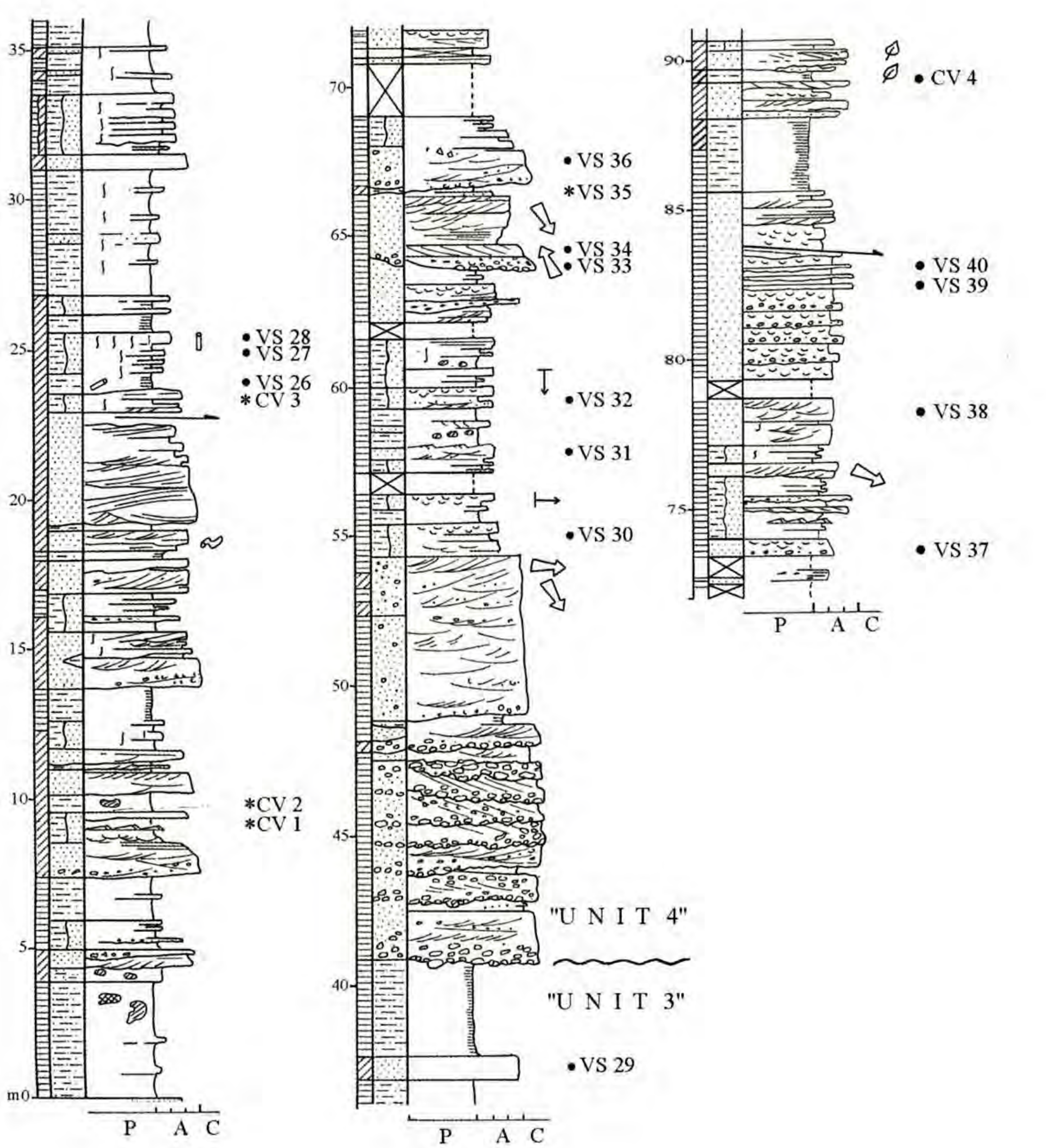

Fig. 6 – Stratigraphic section of Cala Viola (see Fig. 2 for legend); Units 3 and 4 *sensu* Gasperi & Gelmini, 1980.

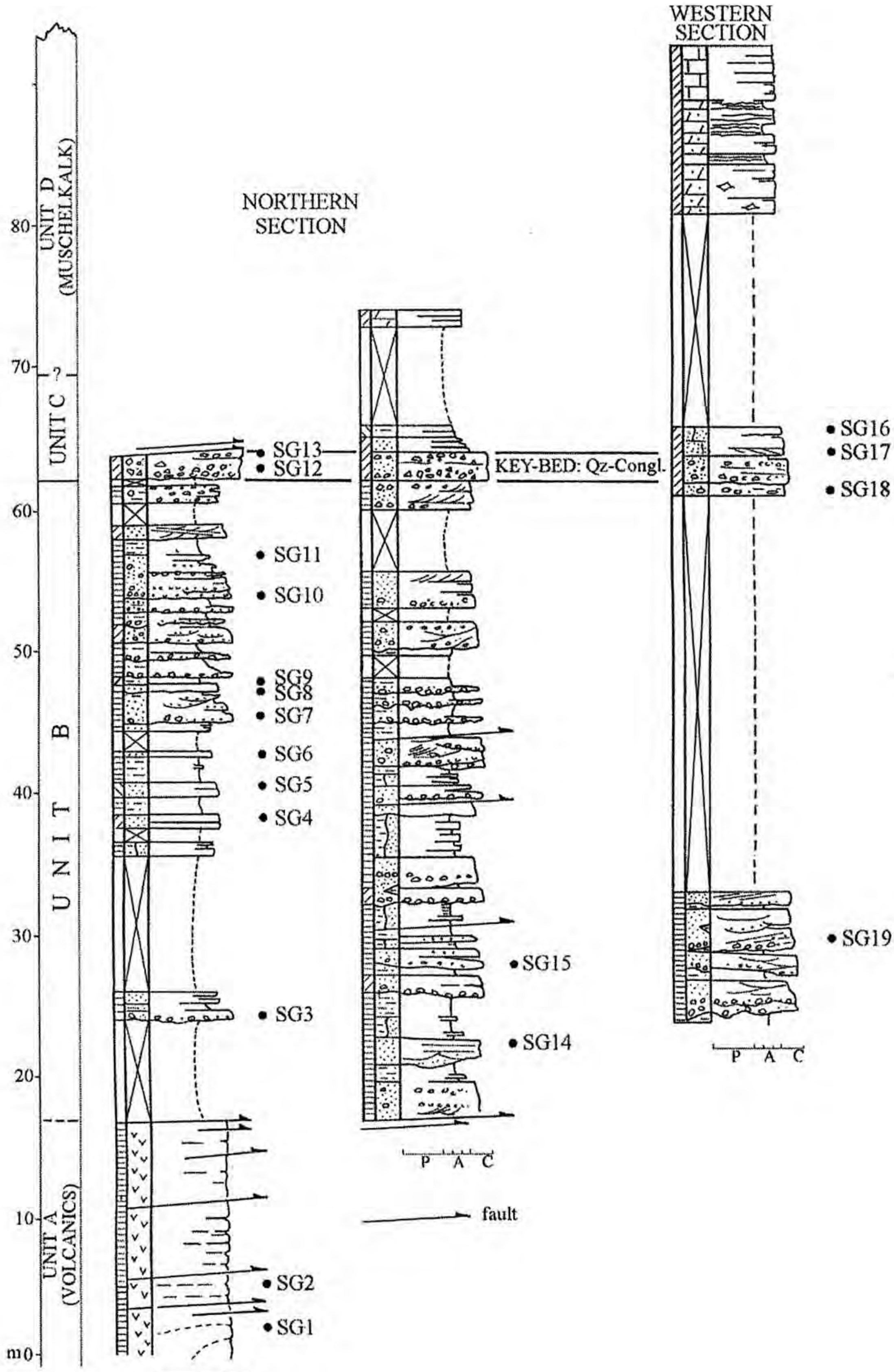

Fig. 7 – Monte Santa Giusta sections (see Fig. 2 for legend).

ically induced lowstand phase of the base level. The peculiar clast composition suggests a rejuvenation of the source area and a new phase of sedimentation after a long period of non-deposition. The presence of tidal sedimentary structures in the upper part of the unit might also suggest the progressive establishment of a transgressive setting in a coastal environment.

Monte Santa Giusta sections

Two sections (Fig. 7), representative of the sub-Muschelkalk succession, both poorly exposed, have been examined on the northern and western slopes of Monte Santa Giusta. They have been subdivided by Neri & Ronchi (1999) into the following units:

A) This unit is only recorded in the northern section and consists of volcanic products (ignimbrites), heavily weathered and tectonised. The rocks are made of reddish, roughly stratified "porphyries", with quartz and feldspar phenocrysts surrounded by an aphanitic matrix. Better-preserved volcanic rocks crop out just some tens of metres north of the section, but widespread cover masks their relationships with the sedimentary units (Lombardi *et al.*, 1974; Cassinis *et al.*, 1996).

B) Poorly exposed conglomerates with clasts deriving from both the Variscan basement and volcanics, sandstones and pelites, mainly reddish in colour. Although their fluviatile nature is evident, detailed facies analysis is prevented by the poor exposures. The thick, isolated, conglomerate/coarse sandstone body of the western section (Fig. 7), quite well-exposed, may have corresponded to a braided river setting.

C) The base of the unit consists of a light grey to whitish quartz conglomerate, a few metres thick, easily recognised and correlatable all around Monte Santa Giusta, and thus regarded as a key marker bed. It is overlain by a thin, poorly exposed silty-sandy succession.

D) This represents the base of the Muschelkalk and consists of dolostones with a high terrigenous content, silt to sand-sized, alternating with stromatolitic dolomites. The terrigenous content and the degree of dolomitisation decrease upwards, and fossiliferous fine-grained limestones become the main lithofacies.

Despite their poor exposure and relatively "condensed" nature, according to the authors, the sections of Monte Santa Giusta represent the key to understanding the Permian and Triassic succession of the Nurra region. If the volcanics occurring at the base of the northern section are correlatable with the volcanics intercalated in the lower part of unit 2 of the Porto Ferro-Lake Baratz area (Lombardi *et al.*, 1974), the conclusions may be:

– Unit B of Monte Santa Giusta correlates with units 2 (*p.p.*) and, possibly, 3 of Gasperi & Gelmini (1980).

– The quartz conglomerate of unit C may correlate, on the basis of lithofacies and stratigraphical setting, with the basal conglomerate of unit 4 *sensu* Gasperi & Gelmini (1980) in the Torre del Porticciolo and Cala Viola sections.

However, the observed multiple faults in the Monte Santa Giusta sections could be responsible of the above "condensed" and/or incomplete succession, and to explain the apparent lack of unit 3, *e.g.* possibly emphasised by the pulse of intra-Permian tectonic movements.

Furthermore, while some lithofacies affinities between the sedimentary units of the Monte Santa Giusta and Lu Caparoni-Cala Viola areas could be inferred, problems arise if we attempt to strictly compare the volcanic units in the differing successions.

LITHOSTRATIGRAPHICAL REVISION

The above-reported stratigraphical and sedimentological data allow a partial review of the lithostratigraphic classification of Gasperi & Gelmini (1980); however, due to the numerous uncertainties, we have not introduced new formation or other hierarchic names, as suggested by the stratigraphic codes. A formal lithostratigraphy will be provided when further investigations on the relatively continuous sections cropping out in the steep cliffs south of Punta Lu Caparoni, in great part only approachable from the sea, clarify these relationships and the nature of boundaries between units 1, 2 and 3 of Gasperi & Gelmini (1980).

At present, the post-Variscan and sub-Muschelkalk succession could be re-organised into the following units (Fig. 8):

I) "Punta Lu Caparoni Formation" (PLC), reviewed to include the clastic unit 1, with clasts derived both from the metamorphic basement and volcanic sources.

II) A volcanic "complex", traditionally ascribed to the lower part of unit 2. It crops out clearly in the Monte Santa Giusta section and is also documented in the Cugiareddu (Pomesano Cherchi, 1968) and Cuili Mola Casu wells (Lotti, 1931), where two "porphyrite" horizons occur, alternating with coarse clastic deposits with volcanic pebbles. According to the literature (Lombardi *et al.*, 1974; Cassinis *et al.*, 1996) and field observations of the authors, other volcanic outcrops also occur in the Nurra region (*e.g.*, Casa Satta, northeast of Lake Baratz), presumably at the base of unit 2, but we are not able to confirm their occurrence and stratigraphical setting. The stratigraphic relationships and the correlation between the two episodes of volcaniclastic activity (presumed calcalkaline at Lu Caparoni-Cala Viola area, and alkaline at Monte Santa Giusta, respectively) reported in the Nurra region by Lombardi *et al.* (1974), Cassinis *et al.* (1996) and Cortesogno & Gaggero (1999) are, at present, unclear.

However, the importance of volcanics is documented by the great amount of porphyric pebbles and cobbles in the conglomerates and conglomeratic sandstones of the unit 2 of Gasperi & Gelmini (1980).

III) A clastic unit of conglomerates, sandstones and minor pelites, rich in volcanic-derived clasts (units 2 *p.p.* and 3 of Gasperi & Gelmini), discontinuously cropping out along the coastline from Torre Negra to Porto Ferro. Although the contact with the overlying unit 3 has not been observed, the overall facies pattern suggests that units 2 and 3 may form a single, major fining-up sequence, evolving upwards from alluvial fan/braided river to coastal plain settings.

IV) This corresponds to unit 4 of Gasperi & Gelmini (1980) and lies disconformably on the underlying unit (but with no possibility of estimating the duration of the hiatus) and grades upwards into the carbonate deposits of the Muschelkalk.

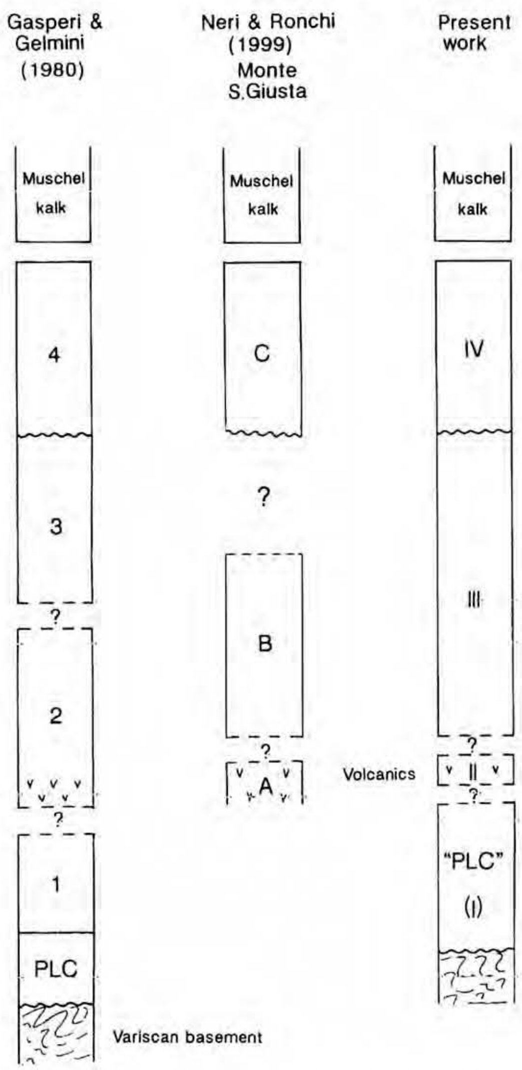

Fig. 8 – Comparison among the lithostratigraphic schemes of the Permian and Triassic succession of the Nurra region according to Gasperi & Gelmini (1980), Neri & Ronchi (1999) and that proposed in the present work. PLC: Punta Lu Caparoni Fm.

COMPOSITION OF ARENITES

The compositions of 50 selected samples of arenites and microconglomerates from the Torre del Porticciolo-Cala Viola stratigraphic sections were examined to determine the provenance of the detritus and to show the diagenetic overprinting.

The most common clast types in the examined samples are monocrystalline quartz grains (with slightly to highly undulose extinction and, in some cases, deformation bands and lamellae). Polycrystalline quartz, with a wide variety of subgrain sizes and fabrics, show frequently undulose extinction and crenulate crystal boundaries. Other types of terrigenous clasts include volcanic and low-grade metamorphic rocks. Volcanic rock fragments are acidic vitric and porphyric rocks; phenocrysts of quartz with resorbed embayments are common, but other phenocrysts are preserved only as ghosts of original grains. Micron-sized inclusions of opaque minerals cause the cloudiness of most of these fragments. The groundmass is commonly deformed between grains to form pseudomatrix. Silicified porphyries are also present, composed of quartz phenocrysts in a groundmass of microcrystalline and mesocrystalline quartz. Low-grade metamorphic rock fragments include: microcrystalline and mesocrystalline quartz rocks with fine crystals of mica, quartz-sericite schist, quartz-albite and quartz-albite-muscovite phyllites.

Detrital potassium feldspars (mainly perthitic) occur in very small amounts in samples of unit 3. Mica (muscovite) is abundant in fine-grained arenites and siltstones; mica flakes have an evident preferred alignment, except in bioturbated samples.

Various types of intraclasts, such as carbonate nodules composed of numerous small dolomite rhombs and mudstone clasts, are locally abundant in the upper succession (units III and IV).

The abundance of ductile rock fragments in some arenites, chiefly volcanic rocks which are highly susceptible to mechanical compaction, produces large amounts of pseudomatrix.

Modal analyses from this work and from previous studies (Cassinis *et al.*, 1996) are reported in Fig. 9. Samples are litharenites to quartzarenites and show a clear compositional trend due to the progressive enrichment of quartz from the basal to the upper units.

As for the provenance of the detrital material, rock types in arenites and conglomerates of the basal successions (vein quartz showing various degrees of strain, quartz-mica schist and phyllites, porphyries and silicified porphyries) indicate contributions from both the Hercynian basement and Permian volcanites. Sediments from the upper successions (units III and IV) show a higher textural and compositional maturity, with a noticeable increase in quartz content. In particular, arenites and conglomerates of unit IV are made up almost exclusively of quartzose grains, and most of these grains are fragments of vein quartz. Thus, the detritus of the upper unit is chemically stable and mechanically durable and could be the residue left after an intense chemical weathering and a long period of mechanical abrasion.

Post-depositional events recognised in the examined samples include: - kaolinite cementation; - formation of iron oxides; - mechanical compaction; - quartz cementation; - Fe-calcite cementation; - dolomite/ankerite cementation; - pressure solution, - recrystallisation of clay to sericite and illite, and - precipitation of calcite in fractures.

The ages of many of these events are not well constrained. Hematite forms a red pigment that stains fine-grained constituents and partly or completely engulfs and surrounds some coarse grains. Kaolinisation of acidic volcanic fragments is common, kaolinite also occurs as pore-filling cement.

The most common cement is quartz. Quartz overgrowths are present in most of the arenite samples and range from a trace to about 10%. Large quartz overgrowths are especially notable in the coarser arenites of unit IV. Margins of quartz grains have dust rings, generally of iron oxides. Pressure solution has destroyed or obscured the morphology of parts of some grains. The scarcity of quartz cement in fine-grained arenites and siltstones may reflect the almost complete loss of primary porosity by ductile grain deformation prior to the time of cementation by quartz. As for the source of silica for the quartz cement, certainly there has been sufficient pressure solution of quartzose grains in conglomerates and sandstones to furnish it.

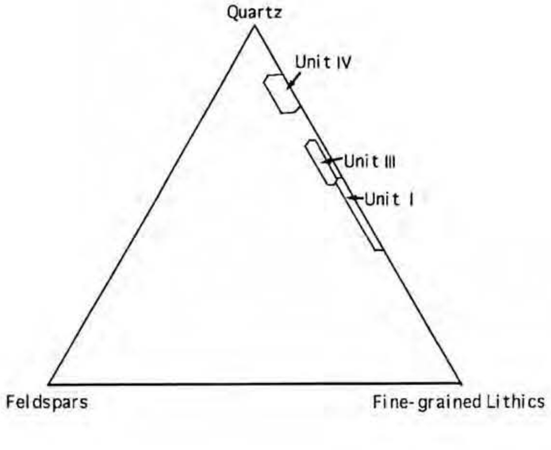

Fig. 9 – Compositional plot of the examined Permian and Triassic arenites from Nurra (from this work and from Cassinis *et al.*, 1996).

Diagenetic carbonates are Fe-calcite (mainly in lower samples) and dolomite and ankerite (in the upper samples); they occur as patches in some of the arenites and locally replace parts of detrital grains. Fractures developed after uplift were filled with sparry iron-free calcite .

CONCLUDING REMARKS

Based on detailed lithostratigraphical, sedimentological and petrographical data, supported by scattered and rare biostratigraphical evidence, a preliminary revision of the terrigenous Permian and Triassic successions of the Nurra region (northwestern Sardinia) has been attempted. The whole post-Variscan continental succession has been subdivided, from bottom to top, into the following horizons:

I) "Punta Lu Caparoni Formation". This formation, already instituted by Gasperi & Gelmini (1980) to indicate vertically and laterally discontinuous alluvial-to-lacustrine deposits, has been reviewed here to also include the overlying clastic succession (unit 1 of Gasperi & Gelmini, 1980).

II) Volcanic unit, documented in the Monte Santa Giusta section, the drillings of the Cugiareddu and Cuili Mola Casu Wells, and, according to the literature and field observations of the authors, in the Lake Baratz area. Its significant original extent is documented by the large amount of volcanic pebbles in the conglomerates and conglomeratic sandstones of the overlying unit.

III) A clastic unit of conglomerates, sandstones and minor pelites, weathered and reddish in colour, rich in volcanic-derived clasts. It crops out discontinuously along the shore from Torre Negra-Torre Bianca to Porto Ferro and in the Lake Baratz area. Based on our data, we interpret this unit (corresponding to units 2 and 3 of Gasperi & Gelmini, 1980) as a single major fining-upwards sequence, evolving from alluvial fan/braided river settings to meandering river and coastal plain deposits.

IV) The upper clastic unit consists of reddish sandstones and minor pelites, showing alternating fluvial and tidal influx. The basal contact is marked by an evident disconformity. The base of the unit, well exposed in the Torre del Porticciolo and Cala Viola area, consists of a quartz conglomerate, followed upwards by cross-bedded sandstones. This unit corresponds to unit 4 of Gasperi & Gelmini (1980). The upper contact with the Muschelkalk deposits is, even if partly covered, transitional in the Monte Santa Giusta section, while at Cala Viola it has been faulted out.

As for the provenance of the detrital material, rock types in arenites and conglomerates of the basal successions indicate contributions from the Hercynian basement and Permian volcanic rocks. Moving upwards in the successions, sediments show a higher textural and compositional maturity. In particular, arenites and conglomerates of unit IV are made almost exclusively of quartzose grains (vein quartz). This type of detritus is chemically stable and mechanically durable and could be the residue left after an intense mechanical abrasion and/or chemical weathering.

Although few biostratigraphical data are available, based on scattered macroflora and palynomorphs occurrence, they allow us to relate the base of sequence to the Early Permian ("Autunian") and the top to an Olenekian(?)-early Anisian to late Anisian time, but do not allow a zonation of the whole sequence or an evaluation of the duration of hiatuses at the main unconformities.

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A PERMIAN MARINE SEDIMENTARY RECORD IN THE FARMA VALLEY (MONTICIANO-ROCCASTRADA RIDGE, SOUTHERN TUSCANY)

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Key words – Continental rifting sedimentation; Brachiopods; Permian; Southern Tuscany.

Abstract – A new Permian, fossiliferous siliciclastic unit (Poggio alle Pigne Quartzite) has been recognised in the Farma Valley between the Carboniferous Carpineta Formation and the Civitella Formation of probable Early Triassic age. It represents part of the rare evidence for Permian marine sedimentation in Tuscany, documented by the occurrence in the upper part of the Poggio alle Pigne sequence, of brachiopods which can be related to Strophalosicea, probably *Tschernyschewia typica*.
The Poggio alle Pigne Quartzite represents a transgressive cycle, probably related to an early stage of continental-rifting activity before the beginning of the Alpine Cycle which led to the fragmentation of the Hercynian continent and the subsequent opening of the Mesozoic Tethys.

Parole chiave – Rifting Continentale; Brachiopodi; Permiano; Toscana Meridionale.

Riassunto – Nella valle del Farma, in Toscana meridionale, è stata riconosciuta una nuova successione silicoclastica fossilifera di età permiana, la Quarzite di Poggio alle Pigne, compresa tra la Formazione di Carpineta, di età carbonifera, e la Formazione di Civitella di probabile età triassica. Questa successione rappresenta uno dei rari esempi di sedimentazione permiana marina in Toscana. I fossili rinvenuti sono brachiopodi strophalosicei, probabilmente riferibili a *Tschernyschewia typica*, di ambiente marino neritico.
La successione di Poggio alle Pigne rappresenta un episodio trasgressivo probabilmente connesso con le fasi distensive precedenti l'inizio del ciclo alpino.

INTRODUCTION

In the Farma Valley, near Poggio alle Pigne, Permian brachiopods have been found within a siliciclastic succession previously assigned to the Triassic (Aldinucci *et al.*, 1999). This new unit, so far unrecognized, was mapped as the basal part of the unfossiliferous Civitella Formation, which in the Monticiano Roccastrada Ridge is the oldest member of the Verrucano Group (Costantini *et al.*, 1987). In the Northern Apennines the Paleozoic and Scythian-Carnian successions belong to the deepest units of the tectonic pile (Tuscan Metamorphic Units) which are affected by Alpine low-grade metamorphism in the green-schist facies (Franceschelli *et al.*, 1986; Conti *et al.*, 1991; Elter & Pandeli, 1993 *cum bibl.*).

In Tuscany, these epimetamorphic successions mainly crop out in the Middle Tuscan Ridge, extending from Alpi Apuane to Monte Leoni, on Elba Island, and in Monte Argentario (Fig. 1).

The Monticiano-Roccastrada Ridge, a segment of the Middle Tuscan Ridge south of the Arno River, ranging from Montagnola Senese to Monte Leoni, is deeply cut from west to east by the Farma River. The Farma Valley is a key area for the stratigraphy of the Permo-Carboniferous to Carnian in Tuscany. This time interval is particularly important in the tectono-sedimentary evolution of the Tuscan domain because it includes the transition from the late-/post-Hercynian extensional events to the Alpidic continental rifting (Rau & Tongiorgi, 1974; Bagnoli *et al.*, 1979; Gattiglio *et al.*, 1989; Rau, 1990; Pandeli *et al.*, 1994; Pandeli, 1998, 1999).

Many authors have discussed the stratigraphy, sedimentology, and textural and compositional features of the Triassic siliciclastic metasediments of the Tuscan Domain (Verrucano *Auct.*: Rau & Tongiorgi, 1974; Canuti & Sagri, 1974; Deschamps *et al.*, 1983; McBride *et al.*, 1987; Franceschelli *et al.*, 1987 *cum bibl.*; Costantini *et al.*; 1987), whereas data for the Permian successions are

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scarce (Pasini & Vai, 1997; Pandeli, 1998). In fact the known Permian successions are:
– metasiliciclastics, graphite-rich lacustrine San Lorenzo Formation *pp.* (Rau & Tongiorgi, 1974) and coastal-neritic Rio Marina Formation *pp.* (Bagnoli *et al.*, 1979) which represent the "Autunian" interval of the Late Carboniferous - Early Permian sedimentary cycle;
– hematite-rich, polymictic metarudites (Asciano breccia and conglomerate: Rau & Tongiorgi, 1974; Torri breccia and conglomerates: Costantini *et al.*, 1998), and volcanic-rich metasiliciclastics (Castelnuovo red-sandstone: Bagnoli *et al.*, 1979; Pandeli *et al.*, 1991; Borro del Fregione siltstone: Costantini *et al.*, 1998; Pandeli, 1998) which are related to a semi-arid continental environment of probable Middle/Late Permian age;
– graphite-rich marine metasiliciclastics and carbonates in the Monte Amiata subsurface ("Formation C": Elter & Pandeli, 1991) which contain early Late Permian (Kubergandian: Pandeli & Pasini, 1990) fossils.

The finding in the Farma Valley, north of Poggio alle Pigne, of a siliciclastic succession (Poggio alle Pigne Quartzite) bearing Late (?) Permian fossils adds new data to our knowledge of the Late-Paleozoic in Tuscany.

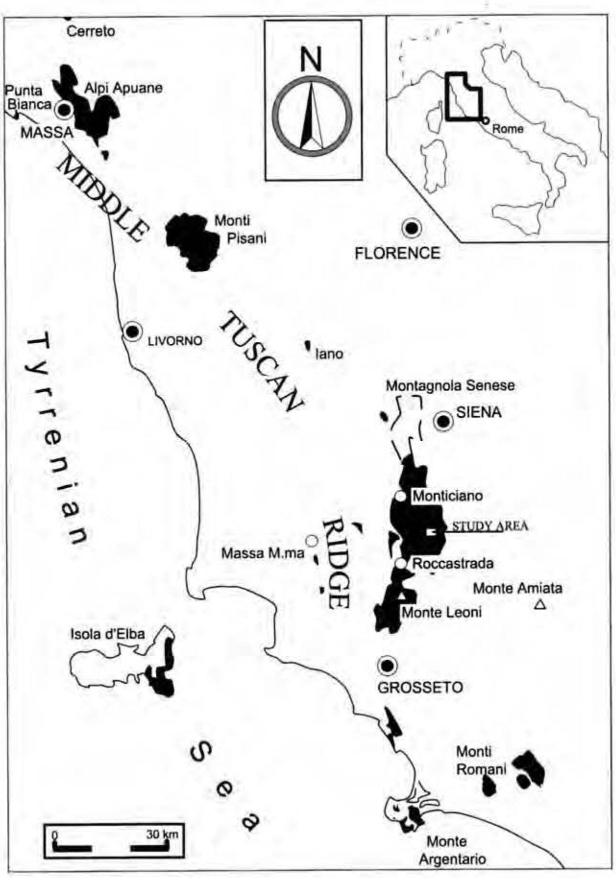


Fig. 1 – Regional distribution of the Paleozoic-Triassic siliciclastic metasediments (in black color) of the Northern Apennine.

GEOLOGICAL SETTING

In Southern Tuscany the Paleozoic to Carnian formations belong to the tectonic Monticiano-Roccastrada Unit which consists of two tectonic subunits: the Montepescali-M.Quoio-Iano (internal) sub-Unit and the M. Leoni-Montagnola Senese (external) sub-Unit (Costantini *et al.*, 1987). The Monticiano-Roccastrada Unit underwent two Alpine tectonometamorphic events in the lower greenschist facies, and later weak folding (Costantini *et al.*, 1987; Franceschelli *et al.*, 1986; Elter & Pandeli, 1993; Bertini *et al.*, 1991).

The Poggio alle Pigne Quartzite is part of the Monte Leoni-Montagnola Senese sub-Unit which in the study area consists of two lithostratigraphic units (Costantini *et al.*, 1987): the Paleozoic Group and the Verrucano Group (Fig. 2).

The Paleozoic Group includes from bottom to top:
a) the Carpineta Formation: graphite-rich metasiltite and phyllite with carbonate-siltitic/limonitic nodules and Upper Visean-Lower Namurian fossils (Redini, 1941; Cocozza, 1965; Pasini, 1978b, 1980a, 1980b; Pasini & Winkler Prins, 1981);
b) the Farma Formation: alternating turbiditic metasandstone and metasiltstone with dark-grey phyllite intercalations and locally with carbonate megabreccias (known as Lower Moscovian Sant'Antonio Limestone) and metacalcarenites of Upper Moscovian age (Cocozza, 1965; Pasini, 1978a, 1980a, 1980b).

The Carpineta Formation and the Farma Formation, which were deformed during a pre-Alpine event (probably related to the "Asturian phase" of the Hercynian Orogeny: Costantini *et al.*, 1987), are unconformably overlain by the Verrucano Group, currently regarded as representing the inception of sedimentation connected with the continental-rifting that led to the fragmentation of the Hercynian continent, at the beginning of the Alpine Cycle (Cassinis *et al.*, 1979).

The Verrucano Group consists of three formations which are, from bottom to top:
a) the Civitella Formation: i) greenish to light grey, massive to poorly graded quartz-metaconglomerates; ii) well-sorted, medium- to coarse-grained greenish to light-grey quartzite with local quartz-metaconglomerate lenses and green/purple metasiltstone intercalations; iii) green and purple metasiltstone and phyllite with interbedded quartzite. The inferred age of this unfossiliferous unit is Early Triassic, although a Late Permian age for its basal part cannot be dismissed (Costantini *et al.*, 1987).
b) the Monte Quoio Formation: purple metasiltstone, phyllite and quartzite with thick lens-shaped intercalations of purplish metaconglomerate ("anageniti" *Auct.*) characterised by the presence of pebbles of white and pink quartz, and subordinately of whitish and purple quartzite

and rare carbonate. Fossils of Scythian-Early Anisian (Cocozza *et al*, 1975) and uppermost Carboniferous-Early Permian age (Engelbrecht *et al.*, 1989) have been found within the carbonate clasts in the Farma Valley. As a consequence, the age of this unit has been ascribed to the Middle Triassic (Cocozza *et al.*, 1975);

c) the "Anageniti minute" Formation: alternating whitish to pink, finegrained quartz-metaconglomerate ("microanagenite" *Auct.*), whitish-pink to purple quartzite and purple metapelite, often arranged in fining-upwards sequences. Owing to the lack of fossils, a probable Ladinian age has been proposed for this formation on the ground of its stratigraphic position beneath the Carnian Tocchi Formation (Azzaro *et al.*, 1976; Costantini *et al.*, 1980).

POGGIO ALLE PIGNE QUARTZITE

The fossiliferous siliciclastic succession crops out in the valley of the Farma River, to the southeast of the villages

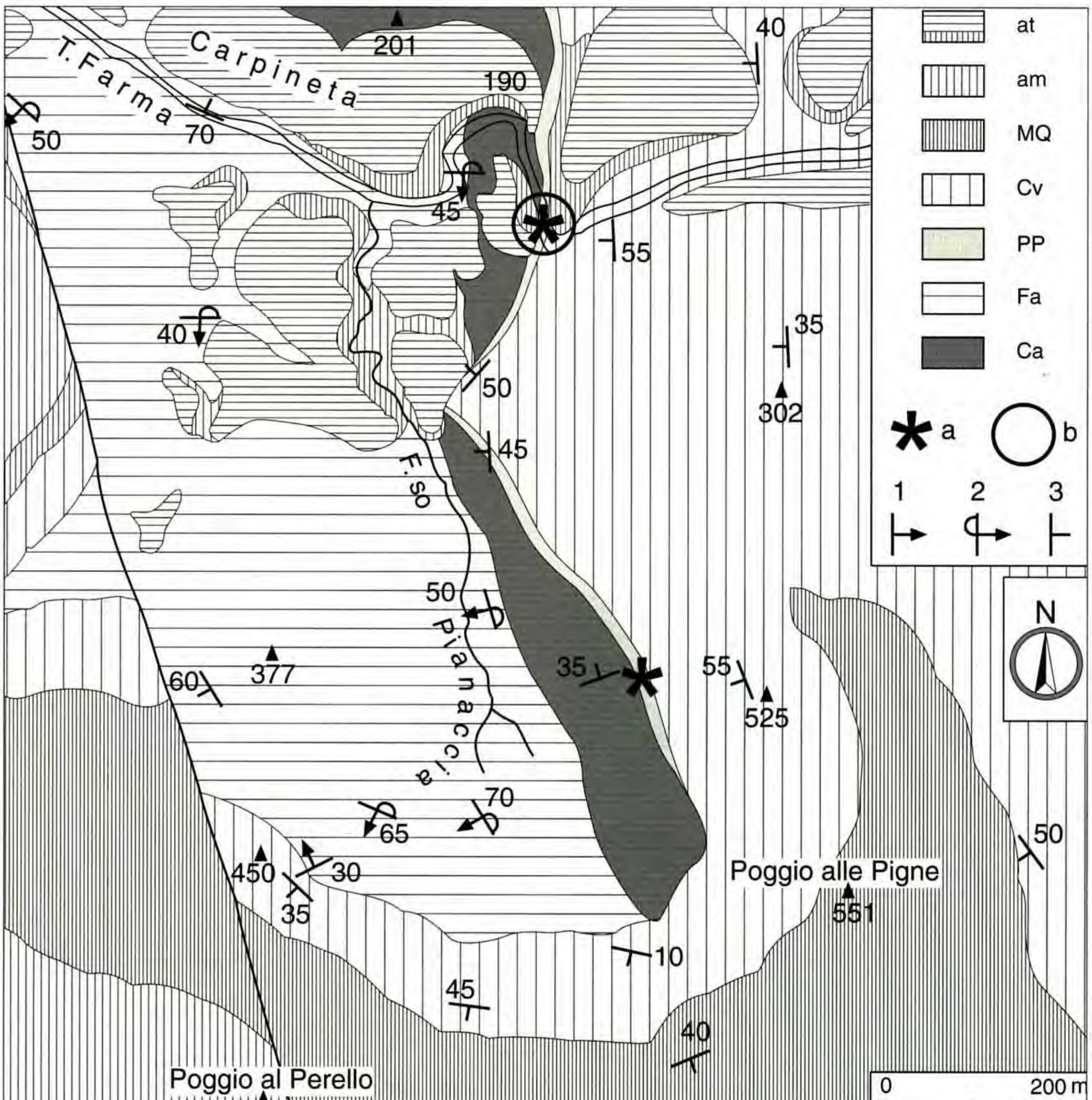


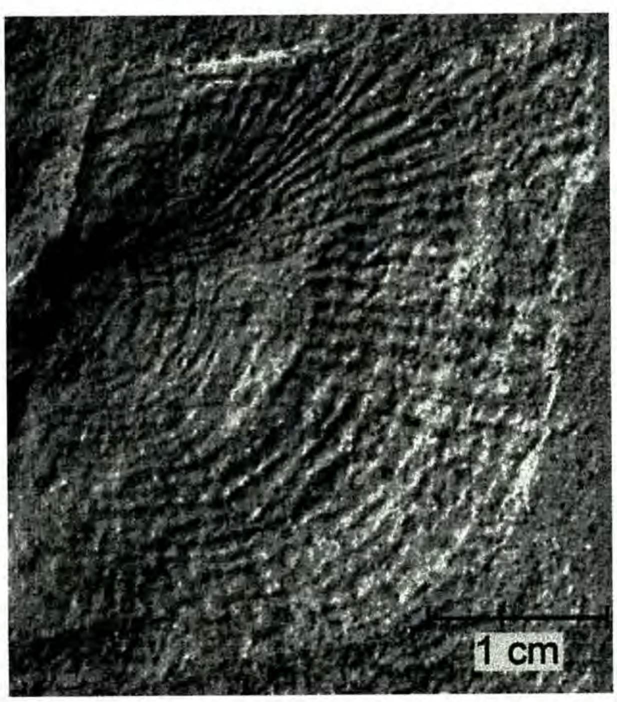
Fig. 2 – Geological map of the Poggio alle Pigne area, after Costantini *et al.*, 1987, modified: a – fossiliferous sites; b – type section; at – alluvial terraces; Am – Anageniti minute Fm.; MQ – Monte Quoio Fm.; Cv – Civitella Fm.; PP – Poggio alle Pigne Quartzite; Fa – Farma Fm.; Ca – Carpineta Fm.; 1 – facing of bedding; 2 – reversed bedding; 3 – strike and dip of bedding.

Iesa and Solaia. In particular the outcrops are located on the western slope of Poggio alle Pigne hill (Fig. 2).
The Poggio alle Pigne Quartzite is stratigraphically interposed between Civitella Formation (Verrucano Group), at the top, and Carpineta Formation (Paleozoic Group), at the bottom.

The lower contact is not well exposed, but at the cartographic scale, an unconformable and probably erosional surface separates the Poggio alle Pigne Quartzite from the underlying black silty phyllites of Carpineta Formation, which was previously deformed during the "Asturian event" of the Hercynian Orogeny. The contact with the overlying metaconglomerates of the Civitella Formation is clearly erosional, although no angular unconformity is recognisable.

The Poggio alle Pigne Quartzite contains, in the uppermost sandstone bed, a monotypical Strophalosiacean brachiopod assemblage represented by moulds of *Tschernyschewia* shells (Figs 3, 4), whose features are similar to the type species of the genus, *Tschernyschewia typica*, occurring in the Upper Permian of Armenia, also reported from Salt Range, Pakistan (B. Wardlaw: pers. comm.).

The investigated section of the Poggio alle Pigne Quartzite was measured in detail along the southern bank of the Farma River, and is composed of two fining-upwards sequences (Fig. 5) whith a total thickness of about 4 m:



Figs 3, 4 – Mould of valves of *Tschernyschewia* shell.

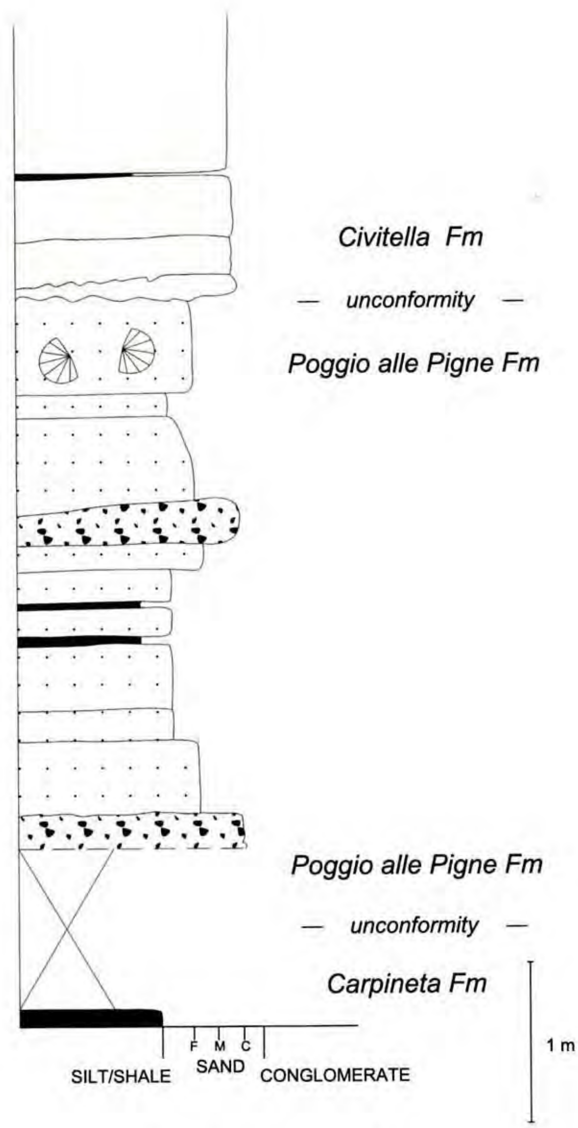


Fig. 5 – Poggio alle Pigne section.

– the lower sequence consists of a basal, polygenic, finegrained conglomerate followed by medium-to finegrained, well sorted quartzitic sandstone with grey to black siltstones and rare phyllitic interbeds;
– the upper sequence starts with a polygenic finegrained conglomerate followed by a medium- to finegrained, well sorted sandstone.

The lack of lateral continuity of the type section is due to the poor conditions of exposure. Nevertheless, at the outcrop scale, the bedding geometry appears to be mainly tabular and the contacts between the layers are sharp, but not erosional, except for the basal conglomerate of the upper sequence which is lenticular and rests with an evident erosional contact, on the lower quartzite.

The composition of the conglomerate beds consists of sub-rounded, white quartz clasts (up to 2 cm) and sub-angular lithic clasts (phyllite, siltstone/finegrained quartzite, radiolarite). The matrix, made up mostly of fine-sand- to silt-sized quartz and white mica, is generally poorly represented.

The sandstones correspond to finegrained quartzite, characterised by the total lack of feldspars and the occurrence of a minor amount of detritic white mica.

DISCUSSION

The finding of Late (?) Permian fossils in the Poggio alle Pigne Quartzite represents the first evidence of an Upper Paleozoic succession along the Monticiano-Roccastrada Ridge. In fact, the youngest Paleozoic unit in this area is represented by the Upper Westphalian-Cantabrian *Spirifer*-bearing shale which unconformably caps the Sant'Antonio Limestone, south of the Farma Valley.

The occurrence of well-preserved moulds of brachiopod shells scattered in the sandstone testifies to a neritic environment. Moreover, the mineralogical composition of the sandstones points to a cratonic provenance for the terrigenous material, whereas the frequency of the conglomerates of lithic clasts associated with the subrounded quartz pebbles suggests that at least part of the source area was not far from the coast and probably affected by block faulting.

The relationships of the Poggio alle Pigne Quarzite with the mixed siliciclastic/carbonate deposits of the Kubergandian "Formation C" recognised in the Monte Amiata deep drillings (Elter & Pandeli, 1991; Pandeli & Pasini, 1990), and with the possible Middle Permian continental siliciclastics (*e.g.* Castelnuovo red sandstone, Asciano breccia and conglomerate) and volcanics (*e.g.* Iano porphyric schists) still remain undefined.

If the Poggio alle Pigne Quartzite represents the stratigraphic cover of the Middle Permian red beds, then a Late Permian transgression in Tuscany would be documented for the first time. This transgression may be related to an early episode of continental rifting (possibly due to dextral megashears affecting the southern part of the Hercynian continent: Rau, 1990) which pre-dated the beginning of the Verrucano sedimentation.

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1.2. OTHER REGIONS

THE LATE PERMIAN ALKALINE MAGMATISM OF THE CINCO VILLAS MASSIF (WESTERN PYRENEES, SPAIN): A RECORD OF THE LATE-VARISCAN EVOLUTION OF THE PYRENEES

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Key words – basalts; dolerites; Late Permian; Western Pyrenees.

Abstract – A basic, primary magmatism, with subalkaline to alkaline affinity, is exposed in three Upper Permian outcrops in the Cinco Villas Massif (CVM; Western Pyrenees, Spain) in the form of basaltic sills, interbedded into clastic sediments.
These basalts are cogenetic (and probably coeval) with doleritic dykes emplaced into Upper Paleozoic rocks in the southern sector of the CVM.
This mantle-derived magmatism is poorly fractionated, as shown by its petrology and geochemistry, especially the rare earth elements.
The emplacement of this anorogenic magmatism is related to late Variscan distension, making it clearly different from the Late Carboniferous - Early Permian orogenic calcalkaline magmatism.
The studied Upper Permian magmatic rocks are the earliest expression of Alpine magmatic activity in the Pyrenees, prior to the Triassic tholeiitic magmatism.

Parole chiave – basalti; doleriti; Permiano superiore; Pirenei occidentali.

Riassunto – Il presente lavoro tratta del magmatismo basico, ad affinità da subalcalina ad alcalina, inerente a tre affioramenti tardopermiani del Massiccio di Cinco Villas (CVM; Pirenei occidentali, Spagna) sotto forma di *sill* basaltici interstratificati con sedimenti clastici. Questi basalti sono geneticamente connessi (e probabilmente coevi) con i dicchi doleritici posti, nel settore meridionale dello stesso Massiccio, entro rocce riferite al Paleozoico superiore. Il magmatismo di derivazione mantellica è scarsamente frazionato, come si desume dalle composizioni petrologica e geochimica essenzialmente delle terre rare. La comparsa e la messa in posto di questo magmatismo anorogenico fu legato alla distensione tardovarisica, che l'ha reso chiaramente differente dal magmatismo calcalcalino orogenico di età compresa tra il Carbonifero superiore e il Permiano inferiore. Il magmatismo tardo-permiano in esame rappresenta la primissima espressione dell'attività magmatica alpina nei Pirenei, prima del magmatismo tholeiitico triassico.

INTRODUCTION

The study of continental deposits with interbedded igneous rocks in selected outcrops of Upper Permian basins is one of the best tools for gaining knowledge of this important time interval, and allows comparison and correlation between different outcrops considered representative of Late Variscan geology in several European areas. Cassinis (1996) provided a synthesis of the information available on the Late Carboniferous and Permian basins of southwestern Europe, underlining the need for studies to complete a detailed paleogeographical and geological scheme for Permian times. In this context, the Cinco Villas Massif (CVM; Navarra, Western Pyrenees, Spain) provides new compositional and age information for three Upper Permian outcrops, affected by a post-Autunian to pre-Buntsandstein magmatism.

The first data about this magmatism in the CVM (Fig. 1) came from Le Fur-Balouet (1985) and Cabanis & Le Fur-Balouet (1989), who studied its petrography in the Larrun and Yanci-Aranaz areas, and Innocent *et al.* (1994), who studied the magmatism cropping out in the Larrun Basin, in the context of Late Variscan magmatic episodes of the Pyrenees. In this paper, we deal with the composition and emplacement age of this magmatism (not studied by previous authors), represented in three Permian outcrops (Larrun, Mendaur and Ibantelli) and also by cogenetic doleritic dykes in the Yanci-Aranaz area.

GEOLOGICAL SETTING

The studied outcrops are situated in the westernmost part of the Pyrenean Axial Zone (PAZ), in a Paleozoic geological unit known as the Cinco Villas Massif (CVM; Fig. 1). The CVM is isolated from the rest of the PAZ by a crustal-scale

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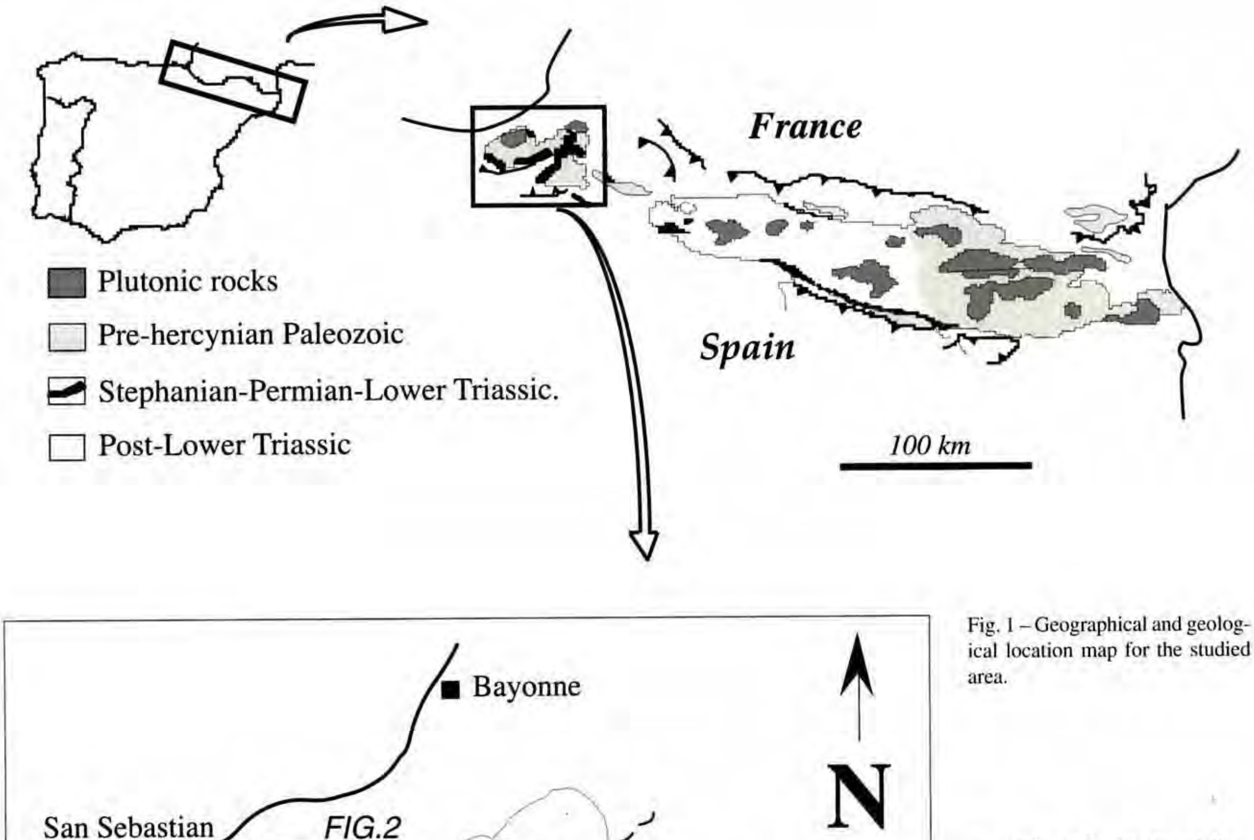

Fig. 1 – Geographical and geological location map for the studied area.

structure (ECORS Pyrenees Team, 1988), the Pamplona Fault, which also affected the North Pyrenean Fault (NPF). The Pamplona Fault divides the NPF into two sectors (eastern and western); these two sectors show different stratigraphic and structural features, as demonstrated when the CVM (located to the west) and the Alduides-Quinto Real Massif (in the eastern sector) are compared. The Leiza Fault, which has been considered to be a continuation of the NPF, is the southern limit of the CVM. The CVM is mainly composed of Devonian and Carboniferous rocks (Fig. 2), which were affected by the Variscan and Alpine orogenies (Martínez-Torres, 1997). The easternmost part of the CVM (also known as the "Allochtonous Unit") is affected by several Variscan thrust sheets; in contrast, the western part of the CVM ("Autochthonous Unit") was unaffected by Variscan thrusts. Late Permian magmatism in the CVM is represented by:

1) basaltic sills, emplaced into sedimentary materials in three areas of Permian rocks (Larrún, Ibantelli and Men-

daur); the base of the Permian sedimentary pile lies unconformably on the Variscan basement, while Buntsandstein facies materials overlie the Permian sedimentary rocks and basaltic sills.

2) doleritic dykes, injected into the Variscan rocks of the western and central sectors of the CVM.

The three Permian outcrops (Larrún, Ibantelli and Mendaur; Fig. 3A, B and C) share some features in common: a) their stratigraphical record, comprising pelites and microconglomerates, equivalent to those described by Gisbert (1981) for the "Unidad Roja Superior" (URS, Upper Red Unit, Upper Permian); b) the occurrence of basaltic sills (two to four events); and c) Buntsandstein facies Triassic rocks (Lucas *et al.*, 1980) resting unconformably on the Permian sequence. Stratigraphic sections for these outcrops are shown in Figs 4A (Larrun), 4B (Ibantelli) and 4C (Mendaur). The microconglomerates seem to be re-

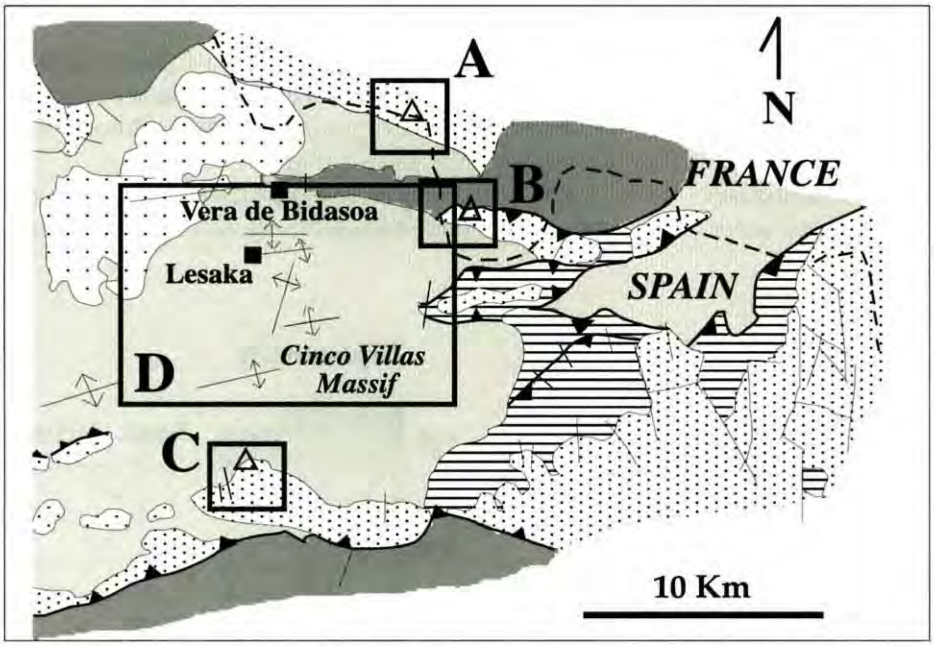

Fig. 2 – Geological map of the Cinco Villas Massif. Capital letters indicate the areas detailed in Fig. 3.

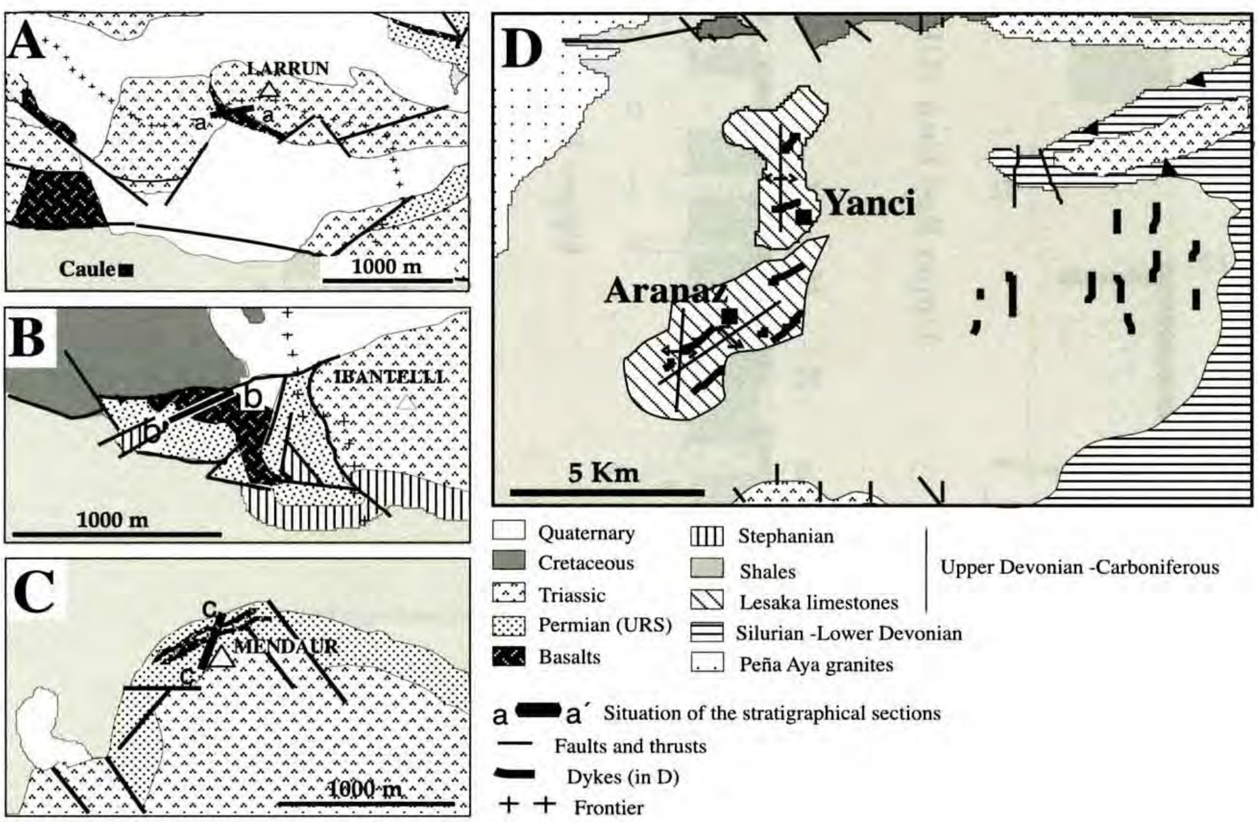

Fig. 3 – Detailed geological maps of the study areas, with locations of the stratigraphic sections. A: Larrun; B: Ibantelli; C: Mendaur and D: Yanci-Aranaz area, with location of doleritic dykes.

lated to basin-opening events, as is suggested by the occurrence of basaltic sills just above each microconglomerate layer (see Fig. 4C). Pelite fragments are included within, and partially disaggregated into, the basalt, suggesting emplacement under a thin layer of unconsolidated sediment of nearly constant thickness and over a large area, as deduced from the lack of contact metamorphism and lava structures. In some cases (lowermost sill in Larrún, of almost 120 m thickness), gravitational settling of crystals is observed (olivine concentrates at the base of the sill, while the top of the sill is richer in plagioclase).

Doleritic dykes crop out in Yanci-Aranaz sector (D in Fig. 2; Fig. 3D), emplaced in carbonate rocks (Lesaka limestones, Upper Famennian-Westphalian; Heddebaut, 1975).

AGE AND EMPLACEMENT CONDITIONS

The emplacement of the basaltic sills, took place at the same time as the deposit of the URS and before the sedimentation of Buntsandstein facies materials, as can be deduced from their stratigraphic position (Fig. 4).

Taking into account the lack of dykes injected into Triassic (or later) materials and the cogenetic character of these

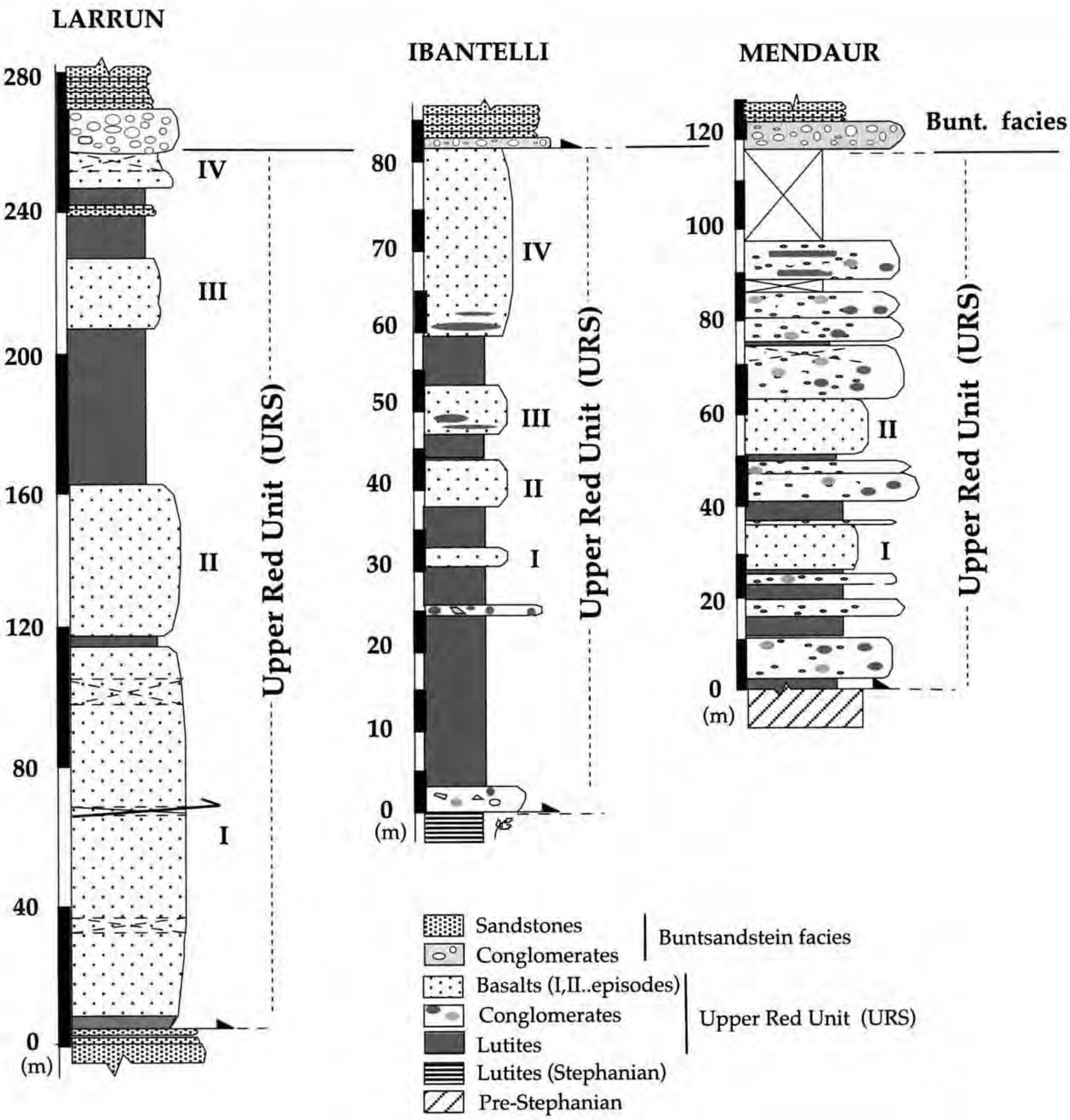

Fig. 4 – Stratigraphic sections for the three studied Permian basins.

dykes with respect to the basaltic sills deduced from their mineral and whole rock-compositions, the emplacement of both sills and dykes can be considered as coeval, favoured by reactivation of the Late Variscan strike-slip WNW-ESE trending fractures (Lucas & Gisbert, 1996; Arthaud & Matte, 1975; Carreras & Capellá, 1994).

The orientations of basins and hypabyssal dykes in the Pyrenees were controlled, during Permian and Triassic times by the sinistral transcurrent movement of the NPF (or the equivalent strike-slip faults at that time) in the western and central Pyrenees, and by dextral shear movement of NE-SW faults in the eastern Pyrenees, as indicated by Lucas & Gisbert (1996; Fig. 5A). These structures were still active in the early Triassic, controlling the evolution of the intracontinental Triassic basins.

The strike of the dykes cropping out in the Yanci-Aranaz area shows a bimodal distribution, with N-S and NE-SW directions (Fig. 5B). The angle between the two sets of dykes is coincident with that indicated in Fig. 5A for basins and dykes in the western Pyrenees, although slightly rotated. This change in orientation of the studied dykes, with respect to that proposed by Lucas & Gisbert (1996), can be related to the counter-clockwise rotation of Iberia during Alpine times.

If the Late Variscan transtensional regime is considered, the formation of subsiding half-graben basins, filled with thick sedimentary piles and basaltic injections (favoured by the extensional regime; Fig. 5C), is a reasonable model for the development of Permian basins.

PETROLOGY AND GEOCHEMISTRY OF SILLS AND DYKES

The basalts (sills) display a variety of textures, but the most

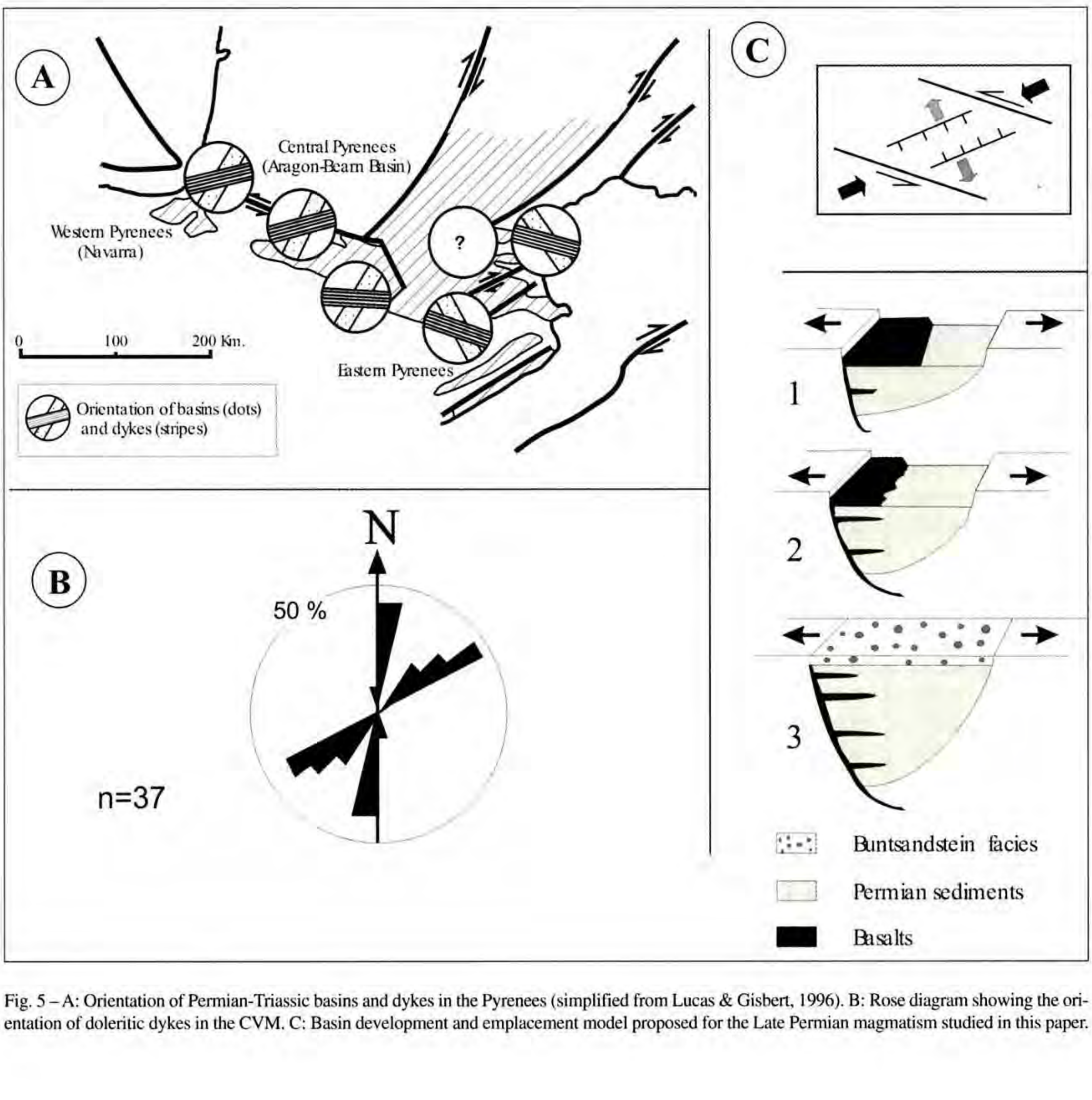

Fig. 5 – A: Orientation of Permian-Triassic basins and dykes in the Pyrenees (simplified from Lucas & Gisbert, 1996). B: Rose diagram showing the orientation of doleritic dykes in the CVM. C: Basin development and emplacement model proposed for the Late Permian magmatism studied in this paper.

common types are hypocrystalline-porphyritic, and rich in vesicles (filled by chlorite and/or carbonates). The main primary minerals are olivine (always altered), Ti-augite and plagioclase feldspar. The modal proportions of these mineral can vary, and two rock types can be defined (olivine-pyroxene basalt, rich in plagioclase and olivine-plagioclase basalt, without pyroxene) in Larrun, but only one (olivine-plagioclase basalt) in Ibantelli and Mendaur.

The dolerites (dykes) have doleritic and/or ophitic textures and two rock-types can be defined: Ti-augite-rich dolerites and plagioclase-rich (An $_{63\text{-}57}$) dolerites, with minor Ti-augite (less than 10%).

The study of mineral compositions in sills and dykes is frequently hindered by the alteration of olivine and plagioclase; accordingly, we have focused our study on clinopyroxene compositions. A selection of clinopyroxene analyses is shown in Table I. These compositions correspond to Ti-rich augite (Fig. 6A), near the diopside-augite boundary, and their evolution is defined by positive correlations of Na *vs*. $*Fe^{3+}$ (Fig. 6B; $*Fe^{3+}$ stands for the estimated value of Fe^{3+}, following the algorithm of Droop, 1987). and Al *vs* Ti (Fig. 6C) The progressive increase in Ti and Na (reaching their highest concentrations at the end of the crystallisation) is consistent with the evolution of an alkaline magma, as

	Basalts					Dykes				
SiO_2	48,763	49,380	48,299	47,881	50,029	47,879	47,818	47,959	48,008	48,304
TiO_2	2,437	1,733	1,948	2,023	2,460	2,626	2,422	2,476	2,574	1,836
Al_2O_3	3,340	2,321	2,723	2,969	3,427	5,255	4,919	4,035	4,341	3,418
V_2O_3	0,203	0,176	0,128	0,135	0,202	0,201	0,229	0,254	0,222	0,113
Cr_2O_3	0,014	0,000	0,089	0,106	0,000	0,242	0,129	0,047	0,082	0,070
Fe_2O_3	2,813	4,073	4,758	5,066	1,158	3,055	4,756	2,753	3,646	5,312
MgO	12,568	13,065	13,453	13,553	13,120	0,031	0,002	0,000	0,000	0,056
CaO	21,080	21,597	21,576	21,583	21,593	12,117	13,236	12,549	12,962	11,406
MnO	0,212	0,250	0,296	0,242	0,143	22,468	22,236	21,841	20,820	22,023
FeO	8,074	6,754	5,484	4,665	8,564	0,191	0,293	0,283	0,224	0,330
CoO	0,000	0,014	0,000	0,000	0,000	4,888	4,560	6,831	7,688	7,530
NiO	0,000	0,000	0,000	0,080	0,000	0,034	0,000	0,000	0,000	0,002
ZnO	0,000	0,098	0,000	0,000	0,063	0,030	0,020	0,020	0,045	0,035
SrO	0,000	0,000	0,000	0,000	0,020	0,000	0,000	0,000	0,000	0,000
Na_2O	0,595	0,544	0,453	0,492	0,470	0,495	0,518	0,450	0,423	0,640
K_2O	0,016	0,008	0,000	0,000	0,010	0,004	0,000	0,000	0,008	0,000
TOTAL	**100,115**	**100,011**	**99,207**	**98,795**	**101,259**	**99,517**	**101,139**	**99,498**	**101,044**	**101,077**
Si	1,836	1,859	1,829	1,818	1,854	1,780	1,772	1,813	1,792	1,817
Ti	0,069	0,049	0,056	0,058	0,069	0,073	0,068	0,070	0,072	0,052
Al	0,148	0,103	0,122	0,133	0,150	0,230	0,215	0,180	0,191	0,152
V	0,006	0,005	0,004	0,004	0,006	0,006	0,007	0,008	0,007	0,003
Cr	0,000	0,000	0,003	0,003	0,000	0,007	0,004	0,001	0,002	0,002
Fe^{3+}	0,080	0,115	0,136	0,145	0,032	0,085	0,133	0,078	0,102	0,150
Mg	0,705	0,733	0,760	0,767	0,725	0,001	0,000	0,000	0,000	0,001
Ca	0,850	0,871	0,876	0,878	0,858	0,727	0,731	0,707	0,721	0,640
Mn	0,007	0,008	0,010	0,008	0,004	0,895	0,883	0,884	0,833	0,888
Fe^{2+}	0,254	0,213	0,174	0,148	0,266	0,006	0,009	0,009	0,007	0,011
Co	0,000	0,000	0,000	0,000	0,000	0,152	0,141	0,216	0,240	0,237
Ni	0,000	0,000	0,000	0,002	0,000	0,001	0,000	0,000	0,000	0,000
Zn	0,000	0,003	0,000	0,000	0,002	0,001	0,001	0,001	0,001	0,001
Sr	0,000	0,000	0,000	0,000	0,000	0,000	0,000	0,000	0,000	0,000
Na	0,043	0,040	0,033	0,036	0,034	0,036	0,037	0,033	0,031	0,047
K	0,001	0,000	0,000	0,000	0,001	0,000	0,000	0,000	0,000	0,000
TOTAL	**4,000**	**4,000**	**4,000**	**4,000**	**4,000**	**4,000**	**4,000**	**4,000**	**4,000**	**4,000**
WO	46,810	47,740	48,150	48,750	45,501	47,981	46,539	46,680	43,748	46,112
EN	38,830	40,170	41,770	42,590	38,461	38,968	38,538	37,311	37,890	33,223
FS	14,370	12,090	10,080	8,660	16,038	13,050	14,923	16,009	18,362	20,666

Table I – Selected clinopyroxene compositions for basalts and dolerites (dykes). $*Fe3+$ stands for the calculated value for $Fe3+$, following the algorithm of Droop, 1987.

can be deduced from a plot of compositions on a Ti *vs* (Ca+Na) diagram (Leterrier *et al.*, 1982; Fig. 6D).

The whole-rock chemical analyses (on 17 basalts and 9 dolerites; selected data are shown in Table II) were carried out at the Universities of Oviedo (major elements and some of the trace elements by XRF) and Granada (rest of the trace elements, including REE, by ICP-MS). Chemical compositions of all the analysed samples show their subalkaline affinity (Fig. 7). On the other hand, Ti enrichment with respect to V (consistent with the Ti enrichment in clinopyroxene), high contents of alkalis, P_2O_5 and Ta, together with relatively low values for Y and Sr (Table II and Fig. 9), argue for an alkaline affinity. The cogenetic character of the sills and dykes can be inferred from the good positive, almost linear, correlations between incompatible elements, as shown by Ta *vs* La (Fig. 8A), Th/Ta *vs* Th/Hf (Fig. 8B), Sm/Nd *vs* Th (Fig. 8C) and Lu/La *vs* Ce (Fig. 8D) diagrams.

Trace element contents (averages of Larrún, Ibantelli, Mendaur and dykes; Table II) have been normalised to N-

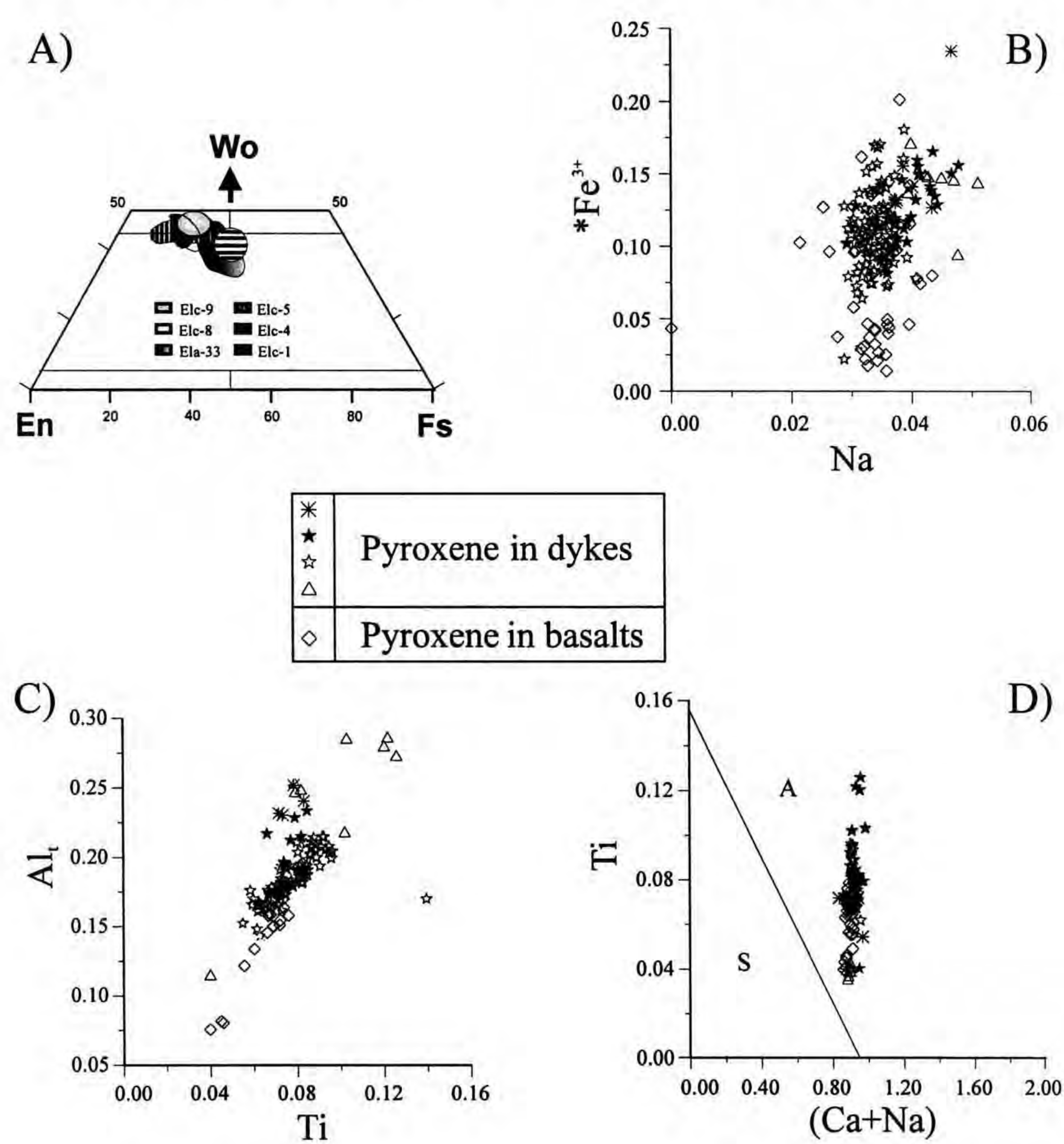

Fig. 6 – Clinopyroxene compositions of basalt and dolerite. A: Classification diagram (Morimoto *et al.*, 1988). B: *Fe3+ *vs* Na diagram. C: Al *vs* Ti diagram and D: Ti *vs* (Ca+Na), according to Leterrier *et al.*, 1982, where A: alkaline domain and S: subalkaline domain.

MORB values (Saunders & Tarney, 1984). The resulting patterns (Fig. 9) indicate the existence of an evolutionary trend (increasing differentiation) from the Ibantelli basalts to the Mendaur basalts; this sequence is in good agreement with that obtained by petrographical criteria.

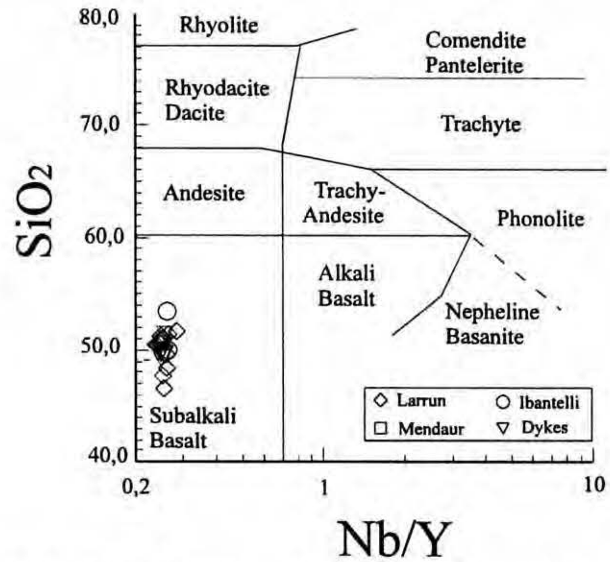

Fig. 7 – SiO_2 *vs* Nb/Y classification diagram (Winchester & Floyd, 1976). All the analysed samples fall in the subalkaline basalt field.

On the other hand, the unexpectedly high values for Th, La and Ce can be explained by crustal contamination processes, which can also be inferred from the Ba, Sr and Rb contents, although these elements may also be affected by alteration processes. Finally, a mantle origin is proposed for the magma, taking into account its unevolved composition and the isotopic ratios $^{87}Sr/^{86}Sr$ (0.706-0.710) and $^{143}Nd/^{144}Nd$ (0.5123-0.5125) obtained by Innocent *et al.* (1994) on Larrun basalts. These authors also suggest that the CVM magmatism can be considered as coeval with that of the Anayet area (Central Pyrenees).

CONCLUSIONS

Three Permian outcrops, located in the CVM (Western Pyrenees), show a sedimentary record supplemented by interbedded basaltic sills, which are cogenetic with doleritic dykes emplaced in other areas of the CVM. Basalts and dolerites evolved from a basic mantle-derived magma, with subalkaline to alkaline affinities, as can be deduced from the

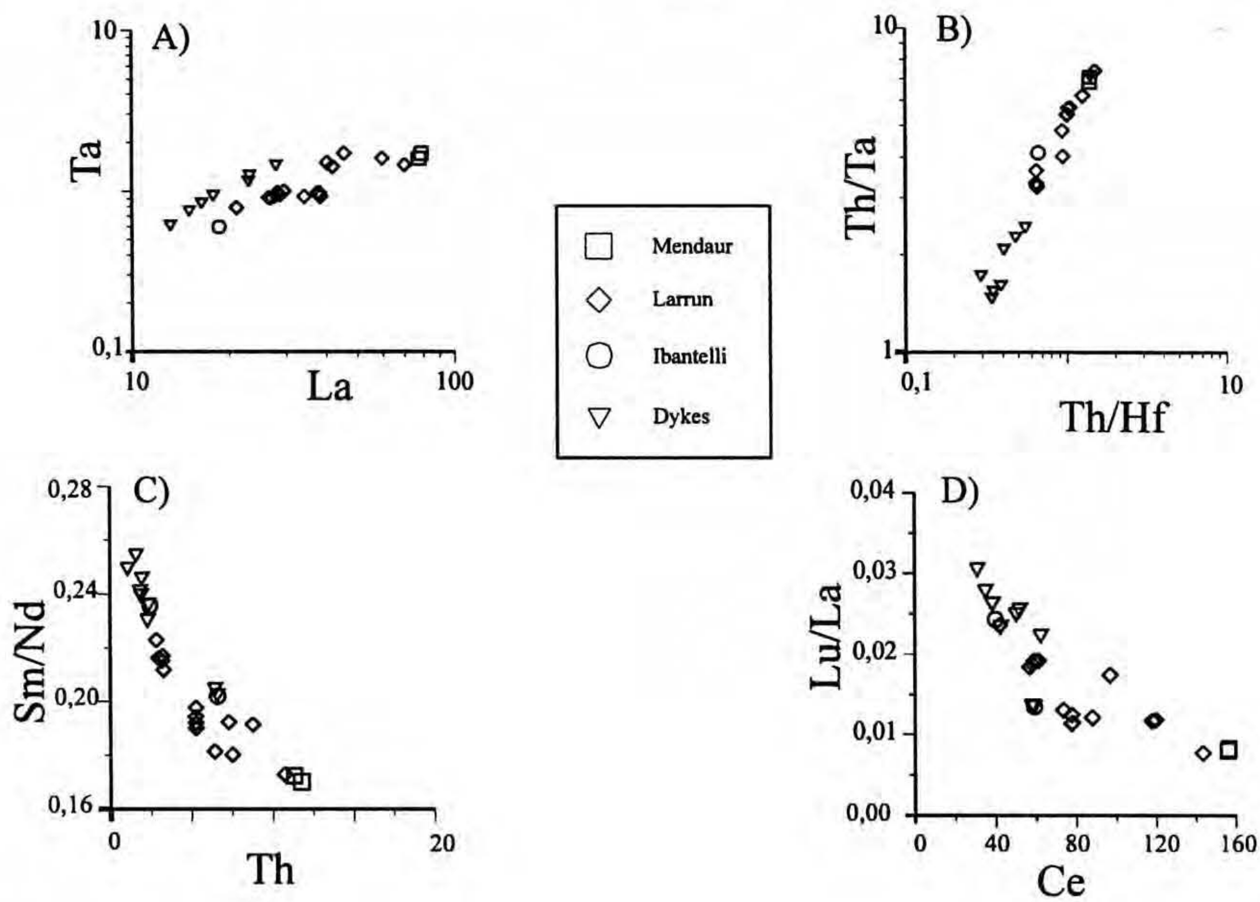

Fig. 8 – Compositions of basalts and dolerites, plotted on variation diagrams for incompatible elements and their ratios. The fairly good correlation of data points suggest the cogenetism of both basalt and dolerite.

	Larrun n=14	Ibantelli n=1	Mendaur n=2	Dykes n=9
SiO_2	46,78	46,54	49,07	46,85
TiO_2	2,02	1,48	2,34	1,95
Al_2O_3	17,16	17,69	16,54	16,82
Fe_2O_3	10,41	8,98	10,75	9,97
MnO	0,10	0,09	0,11	0,15
MgO	7,24	8,24	6,40	6,40
CaO	3,91	6,30	2,48	6,77
Na_2O	4,07	3,75	3,78	3,88
K_2O	1,52	0,41	0,16	1,03
P_2O_5	0,45	0,25	0,63	0,35
LOI	5,76	6,16	7,37	5,22
TOTAL	99,42	99,89	99,60	99,39
Li	109,22	110,40	177,98	47,00
Rb	16,08	9,32	3,12	33,64
Cs	2,64	2,76	8,19	1,74
Be	1,88	0,98	2,09	2,19
Sr	233,67	313,20	65,19	523,36
Ba	314,73	225,36	77,38	170,00
Sc	28,25	29,83	23,86	29,24
V	179,37	181,53	149,80	179,63
Cr	144,87	261,00	56,92	157,31
Co	34,94	32,89	36,22	36,51
Ni	40,02	30,65	24,28	73,35
Cu	33,77	16,50	14,55	42,13
Zn	226,58	134,37	227,09	83,54
Ga	19,64	17,47	22,31	18,96
Y	39,01	26,88	42,89	36,35
Nb	17,38	8,47	22,19	15,05
Ta	1,15	0,60	1,66	0,99
Zr	247,29	146,98	332,15	211,52
Hf	5,81	3,70	8,30	4,67
Mo	1,21	0,70	1,25	1,26
Sn	4,00	2,37	2,85	4,51
Tl	0,06	0,02	0,03	0,19
Pb	15,98	23,59	12,80	7,30
U	0,99	0,62	1,90	0,68
Th	6,17	2,46	11,54	2,91
La	38,59	18,63	78,37	21,27
Ce	82,18	39,98	156,19	47,65
Pr	9,62	5,09	18,58	6,21
Nd	37,53	20,44	69,28	25,95
Sm	7,44	4,82	11,86	6,01
Eu	1,81	1,64	2,37	1,65
Gd	7,07	4,98	9,56	6,10
Tb	1,11	0,82	1,44	0,99
Dy	6,87	5,08	8,20	6,20
Ho	1,45	1,09	1,74	1,31
Er	3,86	2,98	4,60	3,50
Tm	0,58	0,46	0,68	0,53
Yb	3,55	2,92	4,22	3,16
Lu	0,53	0,45	0,64	0,47

Table II – Average whole-rock chemical compositions for basalts (Larrun, Ibantelli, Mendaur) and dolerites (dykes).

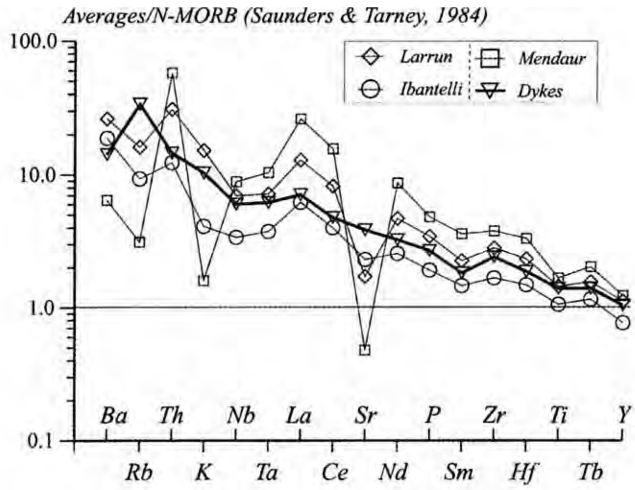

Fig. 9 – N-MORB-normalised spider diagram of average compositions for basalts (Larrun, Mendaur, Ibantelli) and dolerite. See text for interpretation and further explanation.

Ti enrichment in the clinopyroxene and the high contents of P, Ta and Ti in whole-rock compositions. The cogenetic character of the sills and dykes can be inferred from their incompatible trace element ratios and REE patterns, which also show the existence of an evolutionary trend from Ibantelli basalts (the less evolved) to Larrun, and finally to Mendaur basalts (with minor modal olivine).

This magmatism, related to the late-Variscan extensional strike-slip regime, can be considered coeval with that in the Anayet area (Central Pyrenees; Innocent *et al.*, 1994). These Late Permian alkaline magmatic episodes, associated with basin-opening events, represent the first expression of the extension-related Mesozoic magmatism, clearly different from the Variscan magmatism under a transpressional regime (Late Carboniferous-Early Permian), with calcalkaline affinities and a dominant crustal signature. Post-Autunian magmatic episodes are only expressed in central (Anayet; Innocent *et al.*, 1994) and western (CVM, this paper) sectors of the Pyrenees. The available data for the CVM Upper Permian magmatic units are not sufficient to ascertain whether they share the same protolith (the same kind of mantle), or similar melting rates, or both. On the other hand, the emplacement conditions observed for the CVM magmatism are singular and, in this context, the CVM could be considered as a key reference area for the Late Permian magmatism in the Pyrenees.

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THE PERMIAN CALC-ALKALINE MAGMATISM OF THE IBERIAN BELT (SPAIN): AN UPDATED SYNTHESIS

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Key words – magmatism; calcalkaline; hypabyssal; Iberian Belt; Permian.

Abstract – The Upper Stephanian-Permian calcalkaline magmatism of the Iberian Belt is represented by hypabyssal intrusions (sills and dykes) and, less commonly, by pyroclastic deposits, emplaced into Permian sedimentary basins, which are conformably overlain by Buntsandstein facies materials Andesitic compositions dominate, with minor basalts and rhyolites. Stratigraphical, paleontological and radiometric age data, together with tectonic criteria, indicate that, in relation to the movement of NNW-SSE and NW-SE trending faults, this magmatism was emplaced in several stages during the Late Stephanian and Early Permian. The least evolved types (basalt) were derived from the upper mantle, but the original magma composition was modified by interaction with crustal materials, as deduced from the inclusion of granitoid and high-grade metapelitic enclaves, partially or completely assimilated into andesite. These enclaves also provide evidence of deep fractures and, furthermore, they suggest hybridisation between the original magma and crustal restite-rich partial melts.

Parole chiave – magmatismo; calcalcalino; ipo-abissale; Catena Iberica; Permiano.

Riassunto – Il magmatismo calcalcalino tardo-stefaniano e permiano della Catena Iberica è contraddistinto da intrusioni ipo-abissali (in filoni-strato e dicchi) e, meno comunemente, da prodotti piroclastici, collocatisi in bacini sedimentari permiani, che sono ricoperti in concordanza da depositi a facies di Buntsandstein. Si hanno in prevalenza rocce a composizione andesitica, e subordinatamente basalti e rioliti. Datazioni stratigrafiche, paleontologiche e radiometriche, nonché criteri tettonici, indicano che questo magmatismo, in relazione al movimento di faglie orientate NNW-SSE e NW-SE, si mise in posto a più riprese durante il tardo-Stefaniano e il Permiano inferiore. I tipi meno evoluti (basalti) provennero dal mantello superiore, ma la compozione primaria del magma fu modificata a seguito di un'interazione avvenuta con materiale crostale, come si può dedurre dalla presenza di inclusi granitoidi e metapeliti d'alto grado, parzialmente o completamente assimilati dai magmi andesitici. Questi "enclaves" suggeriscono anche l'esistenza di profonde fratture, così come un'ibridizzazione tra il magma primario e parziali fusi crostali ricchi in restite.

INTRODUCTION

The Permian calcalkaline magmatism of the Iberian Belt is remarkably interesting due to: 1) the great number - nearly 1000 mapped outcrops- of subvolcanic intrusive bodies (sills and dykes); 2) all the members of a typical calcalkaline trend being exposed, andesite being the dominant rock type; 3) its epizonal emplacement into upper crustal sedimentary rocks ranging in age from Cambrian to uppermost Permian ("Thuringian"), 4) emplacement through basement fractures which favoured fairly rapid ascent of the magma; 5) the significant volume of crustal enclaves (granitoid rocks and metapelites); 6) the interaction between a mantle component and crustal materials, indicated by assimilation processes that affected metapelitic enclaves and finally, 7) in several Permian basins of the Iberian Belt, pyroclastic deposits – coeval with the hypabyssal rocks – include fossil plants, which facilitates paleontological dating of these materials.

GEOLOGICAL FRAMEWORK

The Iberian Belt is an intraplate orogen, located in the NE of the Iberian Peninsula (Fig. 1) and divided into two sectors: the western sector is known as "Rama Castellana" and the eastern as "Rama Aragonesa". Both sectors are affected by Alpine structures which are compatible with reactivation of the main Variscan faults of the Paleozoic basement. Most of these faults acted as NW-SE and NNW-SSE trending strike-slip faults, several kilometres in length, in response to the Variscan structuration of the belt (Capote, 1983).

Calcalkaline magmatism occurred in the Iberian Belt

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during the late Stephanian-Permian interval, and was related to the opening of sedimentary basins, controlled by steeply dipping normal faults that affected the Paleozoic basement. Pyroclastic deposits are interbedded in the Stephanian-Permian succession, which unconformably overlies the pre-orogenic materials and, in the studied outcrops, is overlain by Upper Permian and Lower Triassic terrigenous sediments (Lago *et al.*, 1992).

These pyroclastic deposits crop out at Sauquillo de Alcázar (Figs 1A and 2; Lago & Pocoví, 1991 a), Fombuena (Figs 1B and 2; Conte *et al.*, 1987), locally over the top of the Ojos Negros sill (Figs 1C and 2; Lago *et al.*, 1994), Orea (Figs 1D and 2; Lago *et al.*, 1995), Atienza (Fig. 1L; Hernando *et al.*,1980) and Rillo de Gallo (Figs 1I and 2; Ramos *et al.*, 1976). In all these sectors, they are interbedded within sedimentary layers containing fossil plants which indicate an "Autunian" age (Lago *et al.*, 1991, 1995; Lago & Pocoví, 1991 a, b; Conte *et al.*, 1987); this age agrees with those obtained by K/Ar radiometric dating for andesites in Fombuena (292 ± 2.5 Ma; Conte *et al.*, 1987) and Atienza (287 ± 12 Ma; Hernando *et al.*, 1980). Thus, the magmatic episode took place during Late Stephanian and Early Permian times and, in any case, prior to the Buntsandstein facies sedimentation.

Hypabyssal intrusions (sills and dykes) are widely exposed in the Iberian Belt, especially at the SE sector of the Rama Aragonesa (Torres, 1989; Torres *et al.*, 1993); the largest volumes crop out at the Sierra de Albarracín (Rama Castellana; Lago *et al.*, 1993, 1996). In all cases, these intrusions have three features in common: a) andesite - both basaltic and amphibole-bearing - is the main rock type; basalt (*e.g.* Ojos Negros, Fig. 1C) and rhyolite (*e.g.* outcrops at Ateca – Fig. 1J – and NW Montalban Anticline – Fig. 1K) being less frequent; b) intrusion took place in several events, as deduced from the interference of andesite dykes crosscutting more differentiated intrusions; and c) they are all cogenetic. An absolute K/Ar age determination was made on biotite from the Loscos gabbro (Fig. 1H; Lago *et al.*, 1991) giving a 293 Ma (late Stephanian-early Permian) emplacement age. This finegrained gabbro is cut by andesite dykes

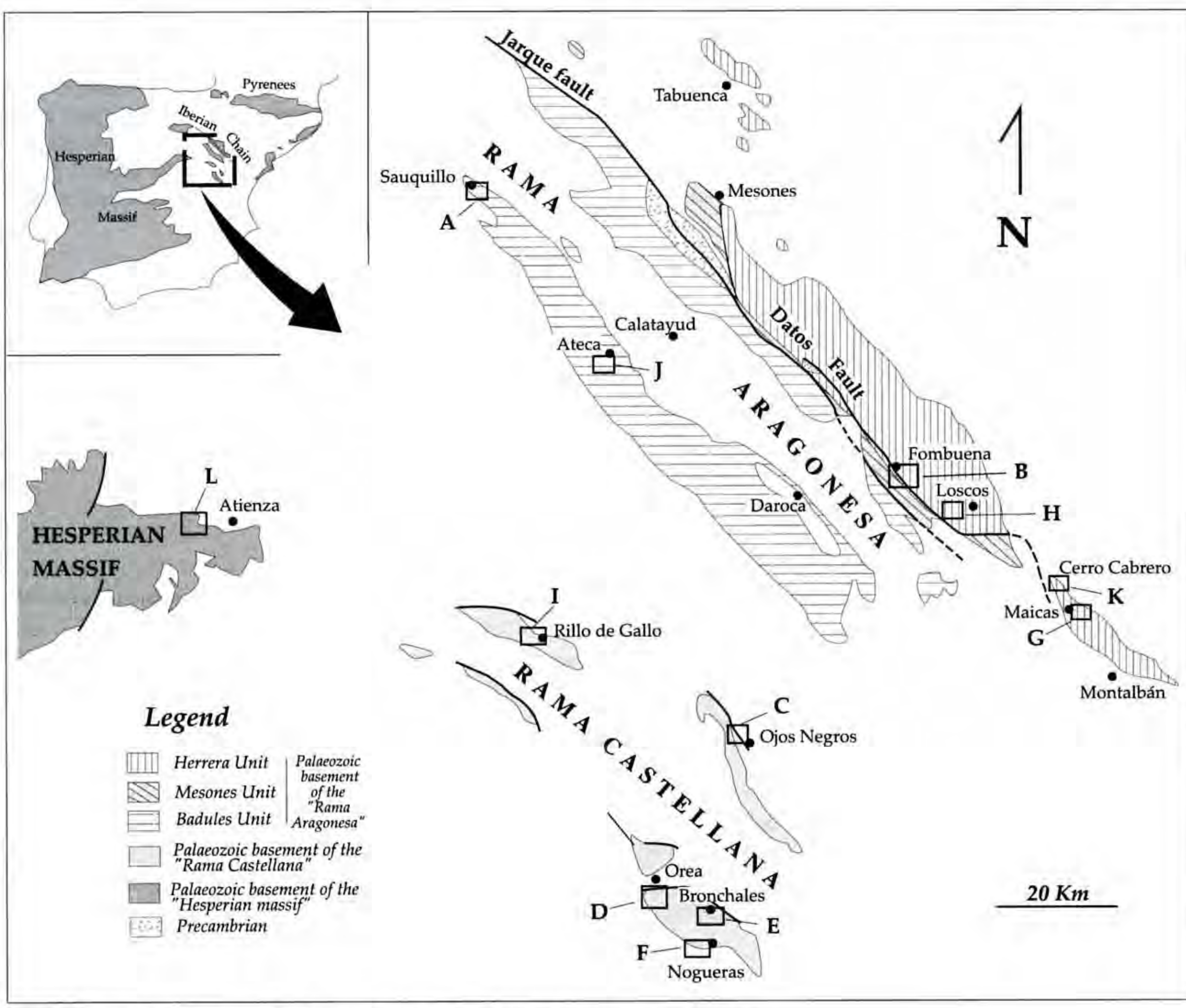

Fig.1 – Geological sketch map of the Iberian Belt, with location of the outcrops cited in the text.

which are cogenetic with all the other studied andesitic intrusions and pyroclastic materials.

MINERAL COMPOSITION AND GEOCHEMISTRY

Most of the igneous bodies (both hypabyssal and pyroclastic) are pervasively affected by secondary alteration processes, and only a few intrusions preserve their original mineral associations almost unaltered. These are the Ojos Negros basalt (Lago *et al.*, 1994), the Loscos gabbro (Lago & Conte, 1989) and some andesites from the Montalban Anticline (Torres, 1989; Torres *et al.*, 1993). The study of their mineral composition (by electron microprobe) and petrology, provides the basis for the definition of the main rock types (Fig. 3). A detailed chemical study

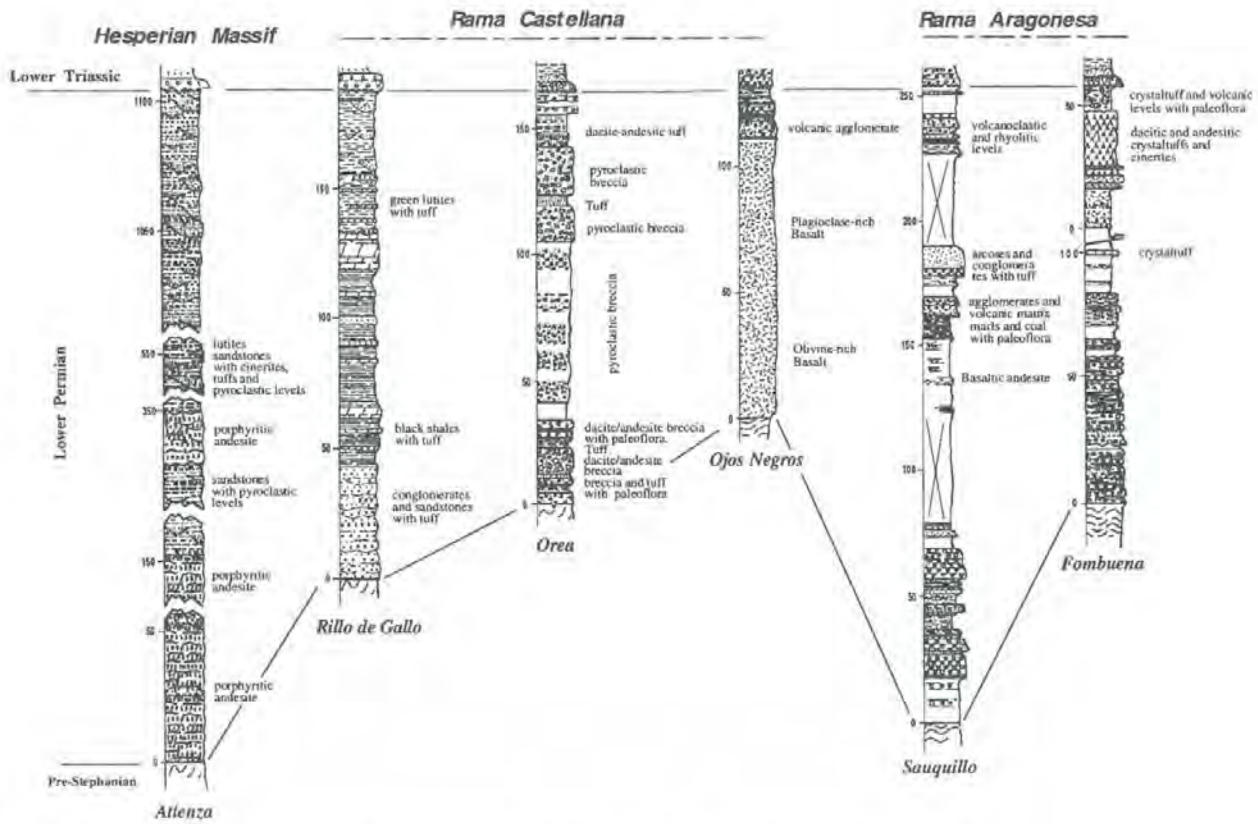

Fig. 2 – Stratigraphical sections for selected outcrops (see Fig. 1 for location) of Permian basins with pyroclastic deposits.

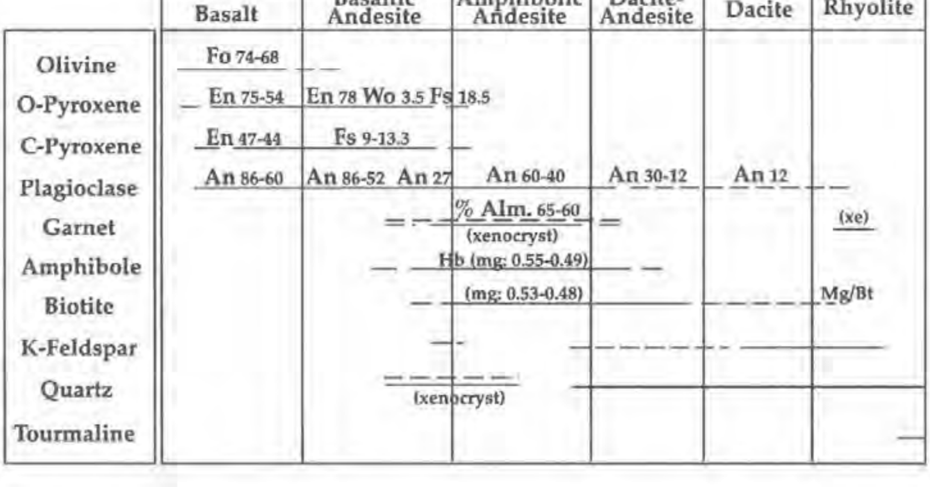

Fig. 3 – Crystallisation sequence for hypabyssal rock-types.

of the ortho- and clinopyroxene and amphibole has also been carried out and the results of this study, together with that of the whole rock geochemical compositions obtained by XRF and ICP-MS- (Table 1) indicate a calcalkaline affinity (Lago *et al.*, 1989 a).

The basalts show a hypocrystalline porphyritic texture, characterised by olivine (Fo_{73-68}), clinopyroxene ($Fs_{8.8}$ to $Fs_{11.8}$) and plagioclase (An_{83-72}) zoned phenocrysts. Basaltic (=pyroxene-bearing) andesite is hypocrystalline and porphyritic, with scarce isolated orthopyroxene crystals ($En_{78}Wo_3Fs_{19}$ on average), clinopyroxene ($En_{50}Wo_{40}Fs_{10}$ on average) and plagioclase (An_{81-27}) crystals as main mineral phases. Amphibole (hornblende) is a minor phase. Quartz xenocrysts are common in these rocks.

samples	M-und. N = 5	M-dif. N = 3	Ojos Negros N = 5
SiO2	52.540	53.700	51.000
TiO2	0.796	0.900	0.803
Al2O3	15.820	16.267	16.120
Fe2O3	7.178	6.937	6.698
MnO	0.128	0.113	0.122
MgO	5.622	5.010	5.176
CaO	7.890	6.980	10.532
Na2O	1.726	2.277	2.102
K2O	0.778	1.177	1.452
P2O5	0.120	0.143	0.128
L.O.I.	6.540	5.417	5.558
*mg**	0.641	0.622	0.638
Total	99.134	98.917	99.691
Li	76.600	60.667	38.800
B	36.400	35.667	45.600
Sc	24.020	20.900	27.960
V	137.400	126.667	207.600
Cr	214.000	180.000	406.000
Co	18.600	18.333	21.800
Ni	18.200	27.000	135.800
Cu	16.460	17.833	35.860
Zn	63.180	67.967	109.760
Rb	27.200	43.333	39.200
Sr	238.600	212.000	268.800
Y	17.600	17.333	23.000
Zr	128.400	148.333	112.400
Nb	9.600	9.667	3.600
Ba	124.200	201.667	273.200
La	18.140	21.233	18.180
Ce	36.820	43.533	36.980
Pr	4.560	5.300	4.220
Nd	18.060	20.400	17.900
Sm	4.220	4.800	3.520
Eu	1.114	1.277	1.152
Gd	4.080	4.700	3.360
Tb	0.540	0.633	0.560
Dy	3.720	4.000	3.800
Ho	0.700	0.730	0.662
Er	2.040	2.133	1.940
Yb	1.960	2.033	2.020
Lu	0.276	0.293	0.216
Hf	2.960	3.633	2.760
Tl	0.300	0.433	0.380
Th	5.120	5.867	4.520
U	1.680	1.600	1.400

Table 1 – Average compositions of selected non-evolved rocks, representative of the studied magmatism. M-und: Pyroxene andesite from Montalban anticline; M-dif.: Amphibole andesite from Montalban Anticline. Ojos Negros: Ojos Negros basalt.

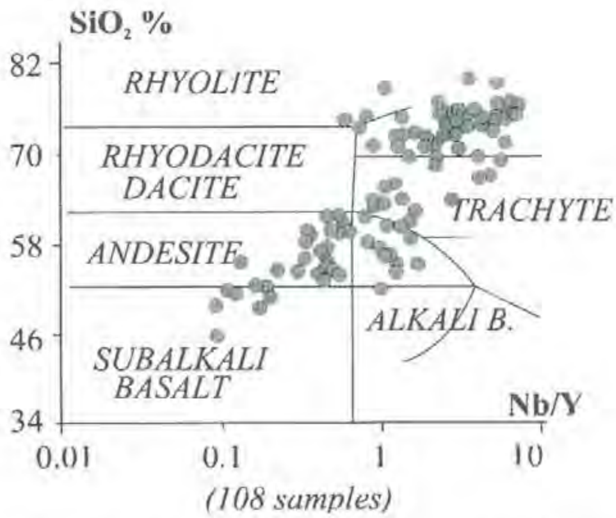

Fig.4 – %SiO_2 vs Nb/Y classification diagram (Winchester & Floyd, 1976).

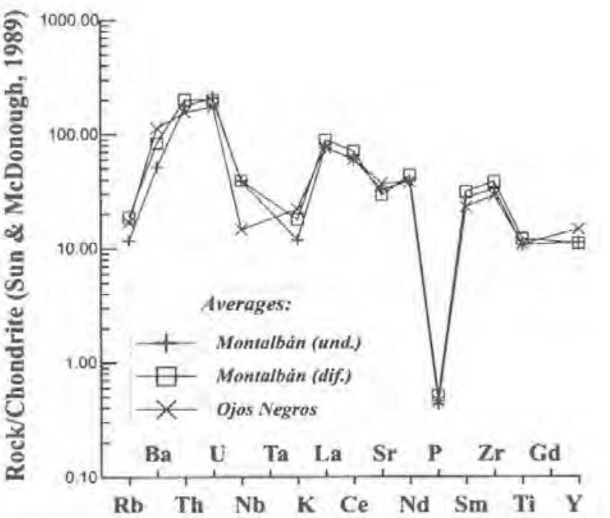

Fig. 5 – Multi-element, chondrite-normalised plot for the average compositions in Table 1. Depletions in P, Nb and K are clearly shown. See text for further details.

The amphibole andesite is hypocrystalline and porphyritic, with minor clinopyroxene, plagioclase (An_{70-20}) and amphibole (hornblende, with Mg^* values ranging from 0.52 to 0.48) as the main mineral phases. Garnet xenocrysts (Alm_{70-60}) are frequent. They occasionally show zoning and reaction rims, showing their instability within the andesitic magma and some degree of re-equilibration of the original metamorphic composition.

The dacite is composed of plagioclase, alkali feldspar, biotite and quartz, with minor amphibole. Finally, the mineral association in the rhyolite (both massive and fluidal varieties) is composed of alkali feldspar, albite, quartz and very scarce Fe-rich (schorl) tourmaline crystals.

Averaged whole-rock chemical compositions for selected samples – representative of the least evolved members of the calcalkaline trend – are shown in Table 1. A SiO_2 *vs* Nb/Y plot of all the available data (Fig. 4) supports the calcalkaline affinity and displays an almost continuous trend from basalt to rhyolite. A chondrite-normalised multi-element plot (chondritic values after Sun & McDonough, 1989) for the compositions in Table 1 shows negative anomalies for Nb, K and P, together with enrichments for Ba, Th, U, LREEs (La, Ce, Sm and Nd) and Zr (Fig. 5).

When normalising REE contents to chondrite (Boynton, 1984), enrichments in LREEs with respect to HREEs, as well as fractionation from basalt to andesite (Fig. 6), is clearly shown; these patterns are in good agreement with other trace element data (Ni: 136 ppm in basalt and <30 in andesite; Cr: 406 ppm in basalt and <200 ppm in andesite, for a Mg^* (= Mg/Mg+Fe) average value of 0.64). The observed REE patterns are typical of a calcalkaline trend, with anomalous enrichment in LREEs (La, Ce and Nd) and Th. These enrichments in incompatible elements can be related to the important contribution of crustal materials to the original basic magmas. In spite of the lack of other evidence (*i.e.* radiogenic isotope ratios), the origin of these basic magmas can be placed in the upper mantle since: 1) there is no evidence of Variscan basic crustal melts for the Iberian Belt and 2) all the rocks fit on to a calcalkaline trend.

CRUSTAL ENCLAVES AND THEIR SIGNIFICANCE

Two kinds of deep-basement enclaves occur in this magmatic unit: a) granitoids and b) metapelites.

Granitoid enclaves (granite, syenite and aplite, all rich in almandine garnet -Alm_{70-60}-) are common in the hypabyssal intrusions of the Rama Aragonesa (*e.g.* Maicas, Monforte de Moyuela and Vistabella; Lago *et al.*, 1991). The particularly interesting outcrop of an andesite dyke near Maicas (Fig. 1G; Lago *et al.*, 1987) contains an uncommon accumulation (up to 30% of the intrusion volume) of granitoid, metamorphic and sedimentary enclaves, which have several features in common: a) their long axis is clearly oriented, being almost parallel to the base of the andesite dyke; b) contacts between enclaves (granite-granite, granite-aplite, etc) are frequent and c) reaction rims can be observed between some of the enclaves and their host, giving rise to a hybrid rim surrounding the whole enclave.

Metapelite enclaves are also frequent; they show variable degrees of assimilation, from a thin reaction rim to complete assimilation, producing garnet (almandine) xenocrysts as relicts of the original enclaves. Two outcrops are specially rich in this kind of enclave: a) Maicas (Fig. 1G), in which biotite – sillimanite – spinel – plagioclase ± corundum ± garnet metapelitic enclaves are very common (Serra *et al.*, 1997 a), and b) the dacite outcrop at Noguera (Sierra de Albarracín; Fig. 1F). In this case, metamorphic enclaves display reaction rims, granoblastic and decussate textures, and in most cases they preserve relicts of an original foliation, defined by biotite and amphibole, and two cleavages. The high-grade mineral association is composed of: biotite - sillimanite - spinel (± garnet [Alm_{72-63}] ± corundum), and also plagioclase (An_{60-33}), which is a late poikiloblastic phase, usually including spinel and/or biotite crystals (Serra *et al.*, 1997 b).

Garnet xenocrysts (Alm_{70-61}) are common in both hypabyssal intrusions (sills and dykes) and pyroclastic deposits (*e.g.* Fombuena; Conte *et al.*, 1987). They are relicts of otherwise completely assimilated metapelitic enclaves; this interpretation is also supported by the clear correlation, for each outcrop, between the quantity of garnet

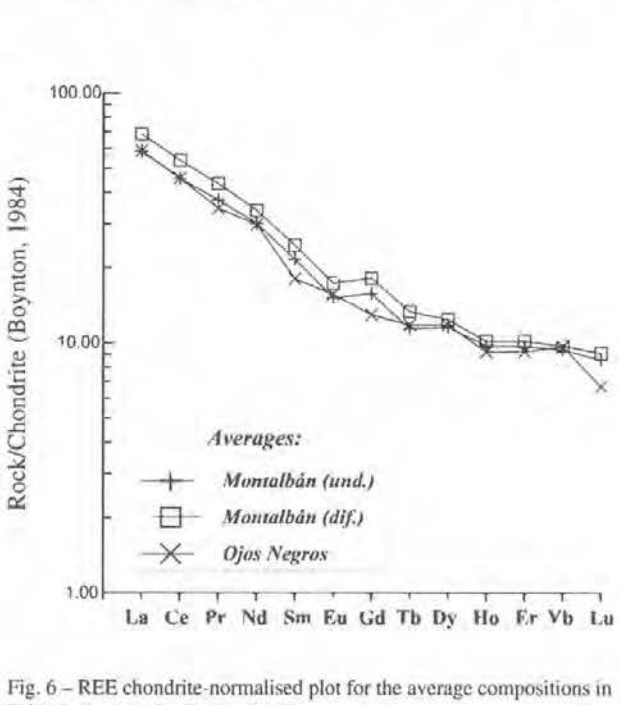

Fig. 6 – REE chondrite-normalised plot for the average compositions in Table 1. See text for further details.

xenocrysts and that of partially assimilated metapelitic enclaves (Lago *et al.*, 1989 b).

The occurrence of high-grade metapelitic enclaves and granitoid fragments within this magmatic unit provides evidence for a) the great depths of the magmatic conduits and b) the interaction between a magma originating in the upper mantle, and lower-middle crustal materials, which were partially or completely assimilated, modifying the original composition of the calcalkaline magma. At this stage, we cannot rule out the possible formation of restite-rich crustal melts that partially hybridised the basaltic magma, giving rise to more evolved andesitic and dacitic compositions.

CONCLUSIONS

The late Stephanian-Permian calcalkaline magmatism of the Iberian Belt, represented by hypabyssal intrusions (with dominance of andesite over basalt and rhyolite) and pyroclastic deposits, was controlled by the Late Variscan extensional regime which was responsible for the formation of deep, crustal-scale faults. This geodynamic framework made possible the ascent of a calcalkaline magma which was eventually emplaced in the form of hypabyssal intrusions (favoured by structural or stratigraphic discontinuities) and pyroclastic deposits which are interbedded into the sedimentary infill of Permian sedimentary basins, limited by normal faults. The primitive composition of the basic magma, derived from the upper mantle, was modified by interaction with lower-crustal materials (now expressed as granitoid and metapelitic enclaves, frequently assimilated) which were carried up by the magma during its ascent and emplacement.

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THE PERMIAN OF SOUTHERN FRANCE: AN OVERVIEW

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Key words – Continental deposits; intramontane basins; macro- and microflora; tetrapod footprints; ostracod; calcalkaline episode; alkaline episode; magmatic suite; "Autunian"; "Saxonian"; "Thuringian".

Abstract – This paper deals with the recent advances in stratigraphy, magmatism and paleogeography of the Permian areas of southern France (French Pyrenees, Massif Central, Provence, Corsica, French Alps), especially since the publication of the Synthèse des Bassins Permiens Français (Châteauneuf & Farjanel, 1989). The lithological successions of the basins have been dated using essentially macroflora, microflora, footprints, crustaceans, insects, and some radiometric ages. The comparison between the different basins reveals significant variations in the time of deposition of the red beds during the Permian and the earliest Triassic. In contrast, the Early Permian seems to be well-defined by the "Autunian" biostratigraphic content.

The geodynamic setting of the studied area shows two contrasting magmatic episodes. The first episode is mainly high-K calcalkaline, its geochemical properties providing evidence for a significant crustal component. At the beginning of the second episode, the magmatic suites became abruptly alkaline. This second episode derived from an OIB source replacing the older lithospheric source with minor crustal contribution, and it is mainly represented by alkaline rocks showing widespread distribution (Morocco, Catalonia, Pyrenees, Corsica-Sardinia, Provence, the Alps).

In many areas (Corsica, Pyrenees, Massif Central, etc.), supplementary evidence for a mid-Permian tectono-magmatic event, reflected by the magmatic episode at 270 Ma, is marked by an angular unconformity at the base of the "Saxonian" and sometimes by slight tectonic movements.

Parole chiave – Depositi continentali; bacini in tramontani; macro- e microflora; impronte di tetrapodi; ostracodi; episodio calcalcalino; episodio alcalino; serie magmatica; "Autuniano"; "Sassoniano"; "Turingiano".

Riassunto – Questo lavoro espone i recenti avanzamenti compiuti in stratigrafia, magmatismo e paleogeografia nelle aree permiane della Francia meridionale (Pirenei francesi, Massiccio Centrale, Provenza, Corsica, Alpi Francesi), soprattutto a partire dalla pubblicazione relativa alla "Sintesi dei Bacini Permiani Francesi" (Châteauneuf & Farjanel, 1989). Le successioni litologiche dei bacini sono state datate ricorrendo essenzialmente allo studio della macroflora, microflora, impronte di vertebrati, crostacei e insetti, ed ai risultati di alcune età radiometriche. Il confronto tra i diversi bacini rivela significative variazioni di età dei *red beds* durante il Permiano e l'inizio del Trias. Al contrario, il Permiano inferiore sembra essere ben definito dal contenuto biostratigrafico dell'"Autuniano". L'assetto geodinamico dell'area studiata mostra due contrastanti episodi magmatici. Il primo episodio è essenzialmente calcalcalino, con alto contenuto in K, e le sue proprietà geochimiche forniscono evidenze di una significativa componente crostale. All'inizio del secondo episodio, la serie magmatica divenne improvvisamente alcalina. Questo secondo episodio provenne da una sorgente OIB che sostituì, con un apporto crostale minore, la più vecchia sorgente litosferica, ed è soprattutto rappresentato da rocce alcaline che assumono un'ampia distribuzione geografica (Marocco, Catalonia, Pirenei, Corsica-Sardegna, Provenza, Alpi). In molte aree (Corsica, Pirenei, Massiccio Centrale, ecc.), l'evidenza supplementare di un evento tettono-magmatico medio-permiano, che è riflesso dall'episodio magmatico di 270 Ma, è segnato da una discordanza angolare alla base del "Sassoniano" e talora da deboli movimenti tettonici.

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INTRODUCTION

Southern France and Corsica constitute intracontinental terranes, part of the large Variscan orogen that resulted from the collision between the African and the North European plates and extended from the Appalachians to the Urals (Fig. 1). These terranes were subjected during the entire Permian period to post-collisional and pre-oceanic extensional episodes, resulting in contrasting magmatic, sedimentary and tectonic regimes (Bonin *et al.*, 1993, 1998). The Late Paleozoic deposits of southern France (Châteauneuf & Farjanel, 1989) are mainly continental sandstones, pelites, and volcanosedimentary rocks, as in the other intramontane basins of Spain (Martinez-García, 1983; Sopeña *et al.*, 1988, 1993; Pieren *et al.*, 1995; Gand *et al.*, 1997c) and Italy (Cassinis *et al.*, 1992, 1995; Broutin *et al.*, 1994; Cassinis, 1996). In southern France, the corresponding Permian outcrops are located in the Pyrenees, at the edges of the French Massif Central, in Provence, in Corsica, and in the Alps. In addition, many boreholes have found generally assumed Permian deposits below the Mesozoic cover in Aquitaine and in the South East Basin: Languedoc, Provence and Dauphiné.

At the end of the 19th century, Munier-Chalmas and de Lapparent defined two continental stages in the Lower Permian deposits: the Autunian, from the Autun Basin, Burgundy, and the Saxonian by analogy with the red facies known in Saxony. These stages labeled, on the one hand, the dominantly greyish and highly fossiliferous strata overlying the Stephanian deposits and, on the other hand, the dominantly reddish and poorly fossiliferous strata overlying the former. The use of these names has been abandoned in the framework of the international time scale which rejects any non-marine stratotype. As the terms were also used in the lithostratigraphic sense, for example by many mining geologists, one of us suggested the use of "*Autunian*" and "*Saxonian*" only as groups, namely lithostratigraphic units next in rank above formation (Gand *et al.*, 1997 c). Very recently, Broutin *et al.* (1999) showed that the paleontological content of the "*Autunian*" of the Autun Basin dated from the latest Ghzelian to early Sakmarian. Therefore, they proposed to keep up the "*Autunian*" chronostratigraphic sense for any continental Permian deposits which could be characterised by the Autunian flora.

The third Permian continental stage defined by Renevier also at the end of the 19th century, namely the Thuringian, has always been considered as equivalent to the German Zechstein (Upper Permian).

DESCRIPTION OF THE PERMIAN OUTCROPS

French Pyrenees

The Paleogene orogeny of the Pyrenees exhumed the oldest part of the belt in the central domain (*Zone primaire axial*). On both French and Spanish sides, the Upper Paleozoic and Mesozoic series have been strongly folded and faulted. The Permian deposits of the Pyrenees (Lucas, 1985) consist of thick red beds comprising conglomerates, sandstones, and pelites, the volcanic rocks associated with the sediments being very common. They are located along the external part of the "Alpine" belt. From west to east, we distinguish (Table 1a): the Basque deposits of La Rhune and Bidarray areas, the Haut Béarn deposits of the Ossau Massif (linked to the Anayet area in Aragon, Spain) and to the east of the belt a number of small outcrops in Hautes-Pyrénées department (Aure valley), Ariège department (Labastide de Sérou, Niort, Sainte-Colombe), and Pyrénées-Orientales department (Ségure, Durban, Baixas) namely Roussillon deposits in Table1a.

The Permian in the Pyrenees was characterised by five volcanic episodes described by Bixel (1984, 1988). Unfortunately, no cross-section shows the superimposition of the five events. The descriptions of the different stages are mainly based on the geochemistry of the magmatic products intercalated in the lithostratigraphic succession. A first episode dated from Late Carboniferous-Early Permian is well-exposed in the Ossau area. Aluminous rhyolites dated from 278 Ma (Innocent *et al.*, 1994) and calc-alkaline dacites dated from 272 Ma are intercalated within the Ossau Fm. A third episode is only revealed in the easternmost part of the belt, notably in the Baixas cross-section. The fourth and fifth alkaline episodes are well-exposed in

Fig. 1 – Location of the Permian deposits of Southern France and surrounding areas.
a. Variscan basement; b. Exposed Permian deposits; c. Main Late-Variscan faults; d. Main Alpine thrusts.
Numbers refer to the Permian outcrops:
Main Permian outcrops of Southern France. 1 to 5 French Pyrenees: 1. La Rhune, 2. Bidarray, 3. Haut-Béarn (Ossau), 4. Aure Valley, 5. Roussillon (Baixas); 6 to 11 Southern Massif Central: 6. Saint Sauves d'Auvergne, 7. Brive, 8. Rodez, 9. Saint-Affrique, 10. Lodève, 11. Prades (Largentière); 12 and 13 Provence: 12. Toulon-Cuers, 13. Bas-Argens-Estérel; 14. Corsica (Monte Cintu); 15 to 19 French Alps: 15. Barrot, 16. Argentera, 17. Belledonne, 18. Galibier, 19. Vanoise.
Other Permian deposits of Northern France and surrounding areas. 20. Exeter (England); 21. Carentan (Normandy, France); 22. Mosel (Germany); 23. Saar-Nahe (Germany); 24. Landau (Germany); 25. Seuil de la Haardt (Alsace, France); 26. Schwarzwald (Germany); 27 to 31 Vosges (France): 27. La Plaine-Nideck, 28. Champenay, 29. Saint-Dié, 30. Villé, 31. Ronchamps-Giromagny; 32 to 37 Northern Massif Central (France): 32. Autun, 33. Montreuillon, 34. Decize, 35. Le Creusot, 36. Bourbon-l'Archambault, 37. Bert; 38 to 41 Italian Alps: 38. Southern Alps, 39. Sesia, 40. Dora Maira, 41. Liguria; 42. Sierra de Cadi (Spain).

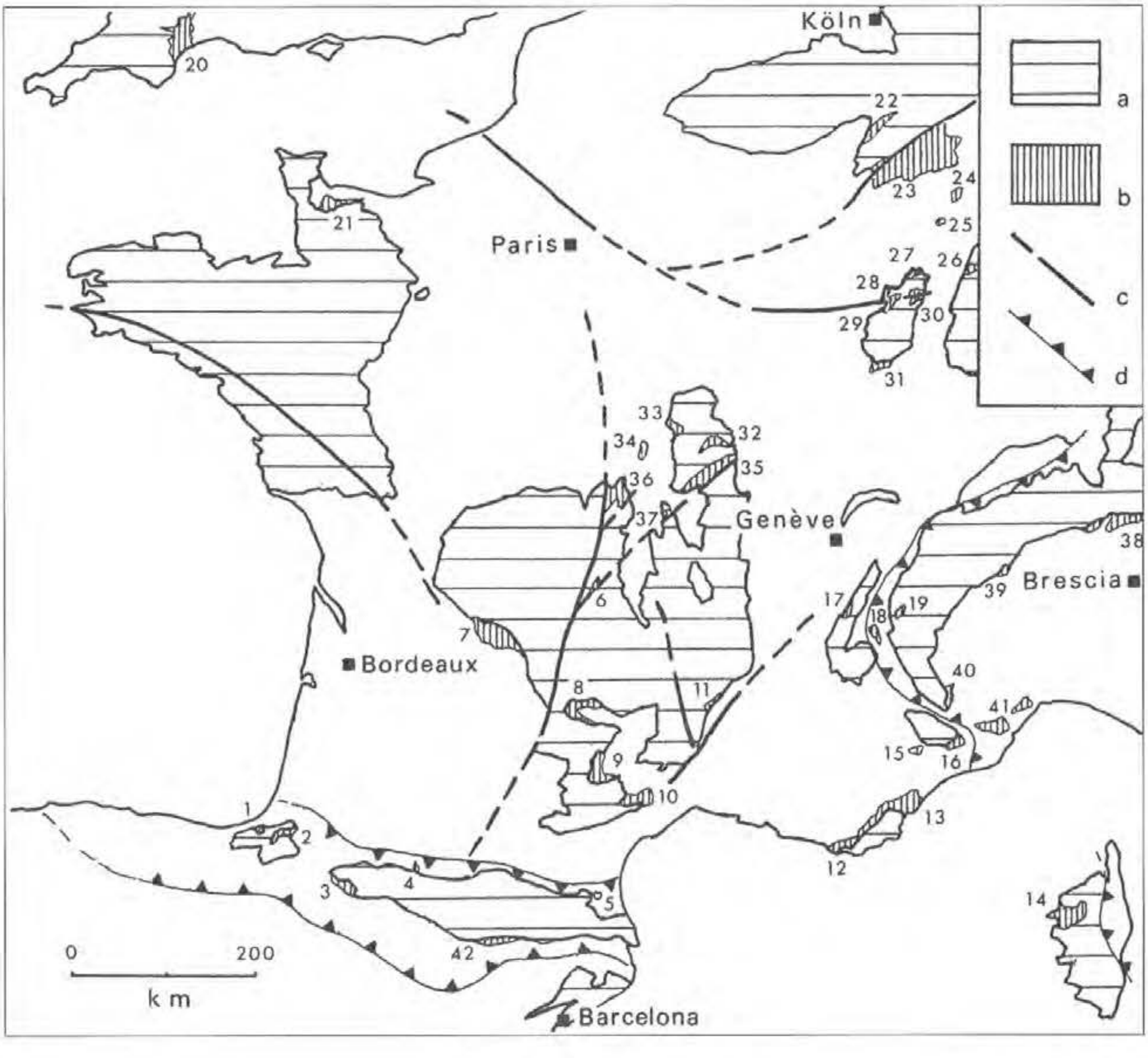
the western part of the Pyrenees, in the Ossau Massif (fourth and fifth) and in the Basque Country (fifth).

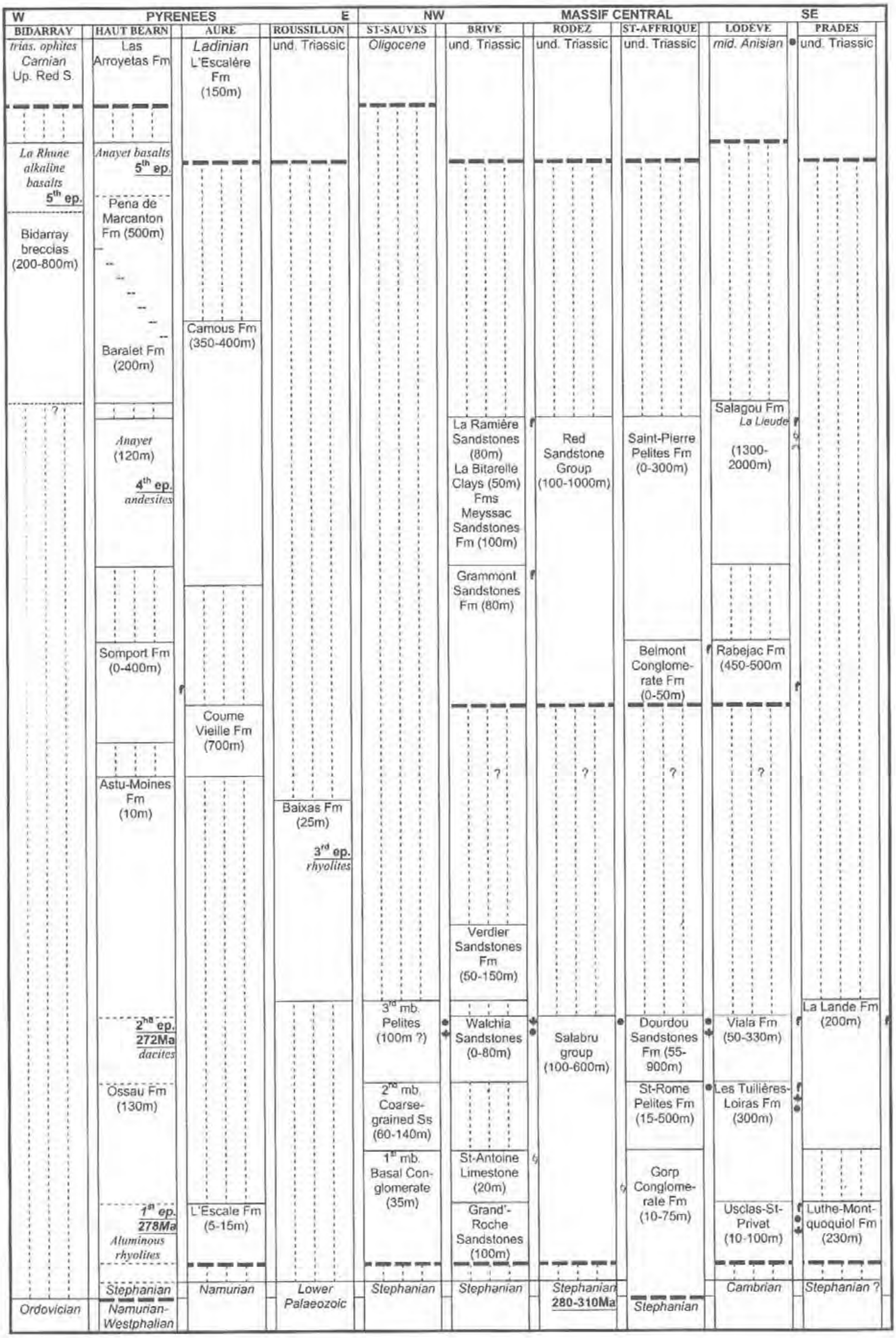
The Pyrenean Permian formations are poor in paleontological evidence. Some microfloras indicate a Stephanian (at the bottom) to Autunian age for the Ossau Fm. Rare footprints have been found by one of us in the Somport Fm. (Gand, 1988). In spite of the absence of fossils, other methods, such as paleomagnetism, indicate a Permian age for some non-fossiliferous series, for example in the Camous Fm. (Schott, 1985).

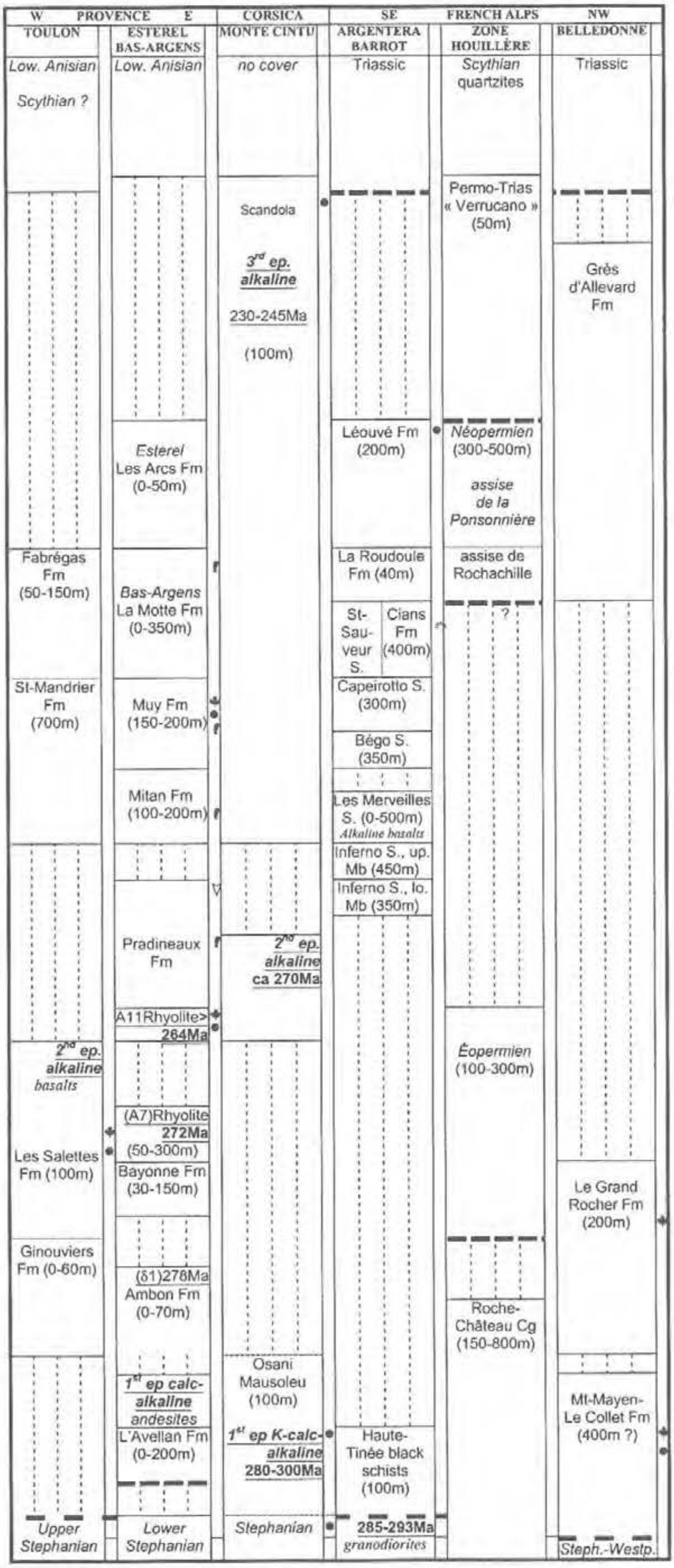
From a structural point of view, the Permian outcrops are mainly – except for the Basque Country – situated in the north of the *Zone primaire axiale* (ZPA), in the so-called *Zone Nord-Pyrénéenne*. In the Central Pyrenees, different settings could be defined. In the Aure Trough, at the northern front of the ZPA, the detrital series is very thick (more than 1,000 m). In contrast, on the Paleozoic North-Pyrenean massifs the Permian deposits are usually thin, but generally present below the base of the Mesozoic formations. Some Permian rocks could also be involved in allochthonous Alpine units, such as the Gavarnie Nappe. All the paleomagnetic data show a considerable rotation for the outcrops of Central and Eastern Pyrenees (Baixas for example), such as the Spanish Permian and Triassic outcrops of Catalonia (between 90 and 100°). Thus, except for the Permian layers of the Basque Country, consistent with a fixed Europe, the Permian deposits were mainly deposited on the mobile block of Iberia.

Edges of the French Massif Central

The Permian deposits of the southern Massif Central are essentially located along its borders: the basins of Brive, Rodez, La Grésigne, Saint-Affrique, Lodève and Prades (Table 1a). One small occurrence appears surrounded by




W
PYRENEES
E
NW
MASSIF CENTRAL
SE
BIDARRAY
HAUT BEARN
AURE
ROUSSILLON
ST-SAUVES
BRIVE
RODEZ
ST-AFFRIQUE
LODEVE
PRADES
trias. ophites
Carnian
Up. Red S.
La Rhune alkaline basalts
5th ep.
Bidarray breccias (200-800m)
Ordovician
Las Arroyetas Fm
Anayet basalts
5th ep.
Pena de Marcanton Fm (500m)
Baralet Fm (200m)
Anayet (120m)
4th ep.
andesites
Somport Fm (0-400m)
Astu-Moines Fm (10m)
2nd ep.
272Ma
dacites
Ossau Fm (130m)
1st ep.
278Ma
Aluminous rhyolites
Stephanian
Namurian-Westphalian
Ladinian
L'Escalère Fm (150m)
Camous Fm (350-400m)
Coume Vieille Fm (700m)
L'Escale Fm (5-15m)
Namurian
und. Triassic
Baixas Fm (25m)
3rd ep.
rhyolites
Lower Palaeozoic
Oligocene
3rd mb. Pelites (100m ?)
2nd mb. Coarse-grained Ss (60-140m)
1st mb. Basal Conglomerate (35m)
Stephanian
und. Triassic
La Ramière Sandstones (80m)
La Bitarelle Clays (50m) Fms
Meyssac Sandstones Fm (100m)
Grammont Sandstones Fm (80m)
Verdier Sandstones Fm (50-150m)
Walchia Sandstones (0-80m)
St-Antoine Limestone (20m)
Grand'-Roche Sandstones (100m)
Stephanian
und. Triassic
Red Sandstone Group (100-1000m)
Salabru group (100-600m)
Stephanian
280-310Ma
und. Triassic
Saint-Pierre Pelites Fm (0-300m)
Belmont Conglomerate Fm (0-50m)
Dourdou Sandstones Fm (55-900m)
St-Rome Pelites Fm (15-500m)
Gorp Conglomerate Fm (10-75m)
Stephanian
mid. Anisian
Salagou Fm
La Lieude
(1300-2000m)
Rabejac Fm (450-500m)
Viala Fm (50-330m)
Les Tuilières-Loiras Fm (300m)
Usclas-St-Privat (10-100m)
Cambrian
und. Triassic
La Lande Fm (200m)
Luthe-Montquoquiol Fm (230m)
Stephanian ?
?



Tables 1a and b

Synthesis of the Permian of Southern France. a) from the Pyrenees to the French Massif Central. b) from Provence-Corsica to the Alps.

A compilation of unpublished results of Châteauneuf & Farjanel (1989), and of:

- for the Pyrenees: Bixel (1984), Bixel & Lucas (1983), Cabanis (1996), Cabanis & Le Fur-Balouet (1989), Innocent *et al.* (1994), Lucas (1985), Lucas & Gisbert-Aguilar (1996) and Schott (1985).
- for the Massif Central: Brive Basin by Feys (1976) and Gand (1991); Rodez Basin by Bourges (1987); Saint-Affrique Basin by Gand (1993 b), Gand *et al.* (1996), Rolando (1988) and Rolando *et al.* (1988); Lodève Basin by Broutin *et al.* (1992), Gand *et al.* (1997a, b, d), Odin (1986) and Schneider *et al.* (1999); Prades Basin by Gand (1993a).
- for Provence: Demathieu *et al.* (1992), Gand *et al.* (1995), Lethiers *et al.* (1993), Toutin (1980), Toutin-Morin (1987, 1992), Toutin-Morin & Bonijoly (1992), Vozenin-Serra *et al.* (1991) and Zheng *et al.* (1991-1992).
- for Corsica: Bonin *et al.* (1987), Cabanis *et al.* (1990), Gondolo (1989), Vellutini (1977).
- for the French Alps: Deroin *et al.* (1996), Vinchon (1984).

Vertical dashed lines represent the stratigraphic gaps. Bold horizontal dashes represent the main discontinuities.

Main biostratigraphic features of each formation

Flora:
(macroflora)
(microflora)

Crustaceans:
(brachiopods: conchostracans, triopsids...)
(ostracods)

Insects

Footprints

the basement, near the Sillon Houiller, at Saint-Sauves-d'Auvergne. The basins are mainly characterised by well-developed red beds cropping out widely above a relatively thin, rhythmically deposited Autunian unit, representing the repeated alternation of torrential with lacustrine deposits. The red beds range in age from Autunian *sensu* Broutin *et al.* (1999) to Upper Permian (*"Thuringian"*). Volcanic rocks are poorly represented in the Brive and Rodez areas but form thick pyroclastic deposits in the Lodève Basin, possibly related to the volcanism of Sardinia, Provence, and Corsica.

Paleontological discoveries have recently been made in the so-called "*Saxonian*" group *sensu* Gand *et al.* (1997d) of the Lodève Basin which was long thought to be non-fossiliferous. Triopsids (*Lepidurus*, *Triops*), conchostracans (associated with triopsids but with a larger vertical distribution), insects (ten orders represented), arthropod tracks and burrows have been identified for the first time (Gand *et al.*, 1997 a, d). Among the triopsids, *Triops* corresponds to a species very close to present-day *Triops cancriformis*. This indicates that these arthropods have evolved little since Permian times. The range of insects suggests Leonardian to early Kazanian ages for part of the Salagou Fm. The Conchostracans are represented by species similar to those described in the "Rotliegende II" *sensu* Schneider *et al.* (1995). They suggest a Kungurian (or Sakmarian) to Tatarian age for the "*Saxonian*" group of the Lodève Basin. In the La Lieude strata the occurrence of conchostracans with ornamentation suggests a Tatarian age for the uppermost part of the Salagou Fm. Other investigations in the Saint-Affrique Permian basin have discovered some triopsids, conchostracans, and insects showing the same fauna as in Lodève (Gand *et al.*, 1997 b).

The intramontane basin of Saint-Sauves-d'Auvergne extends for about 10 km^2. It has been considered as "*Saxonian*" by Brousse (1981) in spite of (or due to ?) the lack of any paleontological data. Unpublished analyses performed by one of us (JJC) provide the first elements for dating this small basin. Spores and palynomorphs from one of the greyish layers of the "Pelites Fm." have shown the spores: *Thymospora* sp., *Spinosporites spinosus*, *S. exiguus*, *Laevigatosporites perminutus*, *Lycospora pusilla*, *Verrucosisporites* sp., *Lophotriletes* sp., *Granulatisporites*; and the pollen grains: undifferentiated smooth Disaccates and Monosaccates, some *Florinites* and *Cordaitina*. At first sight, this microflora could be dated to the Stephanian or the Early Permian, but not the "*Saxonian*". The solution has been found by using weathering effects (epigenesis of organic matter by pyrite) involving the degradation of the biggest species (Mono- and Disaccates). The high quantity of Disaccate and Monosaccate remains suggests a Permian age, which is in better accordance with the occurrence of the red sandstones than a Stephanian age well-known for coal-deposits along the Sillon Houiller (Singles, Messeix).

In the Permian basin of Largentière, on the southeastern border of the Massif Central where no other Permian strata crop out, the reinterpretation of the post-Carboniferous and pre-Triassic deposits by Gand (1993 a) has shown the presence of two formations: Luthe-Montquoquiol and La Lande Fms. A number of fossils and fossil tracks have been identified including vertebrate footprints, macroflora mainly composed of conifers and callipterids, and many *Scoyenia* beds. Using methods such as the comparison with the same ichnofacies from the Thuringian Forest, a correlation with the Asselian - upper Sakmarian of the marine time-scale has been proposed for the La Lande Fm.

Provence

The Permian basins of Provence (Toutin-Morin in Châteauneuf & Farjanel, 1989) follow one another along the border of the Variscan Maures massif, from the east to the west: Estérel and Bas-Argens, bounded to the north by the Tanneron massif, then Le Luc and Cuers-Toulon, which are partly concealed beneath the Mesozoic cover (Table 1b). Whereas the Upper Carboniferous sediments are confined within a few narrow submeridian troughs, affected by a post-Stephanian compressive phase, each of these Permian grabens trends more or less E-W. The thick continental siliciclastic fills are organised in fining-upwards sequences evolving from fanglomerates (Delfaud & Toutin-Morin, 1993) to meandering stream deposits (Durand, 1993) or playa mudstones where paludal to lacustrine carbonates occur sporadically (Toutin-Morin, 1992). They are interrupted at various levels by felsic or mafic volcanic units, which are mostly developed in the Estérel and Bas-Argens where their cumulative thickness reaches a few hundred metres. It is for the eastern basins that the most stratigraphical data are available (Toutin-Morin *et al.*, 1994 a, b). The oldest deposits (L'Avellan Fm., Estérel) are tilted in a very small ESE-WNW graben and capped by the only volcanics in the area pertaining to the calcalkaline episode; thus an Autunian age *sensu* Broutin *et al.* (1999) is assumed for this formation alone. Above the main marker is the A7-Rhyolite, in the middle part of the series, whose ^{40}Ar-^{39}Ar date is very close to 272 Ma (Zheng *et al.*, 1991-92) fitting well with the age of the Artinskian-Kungurian boundary proposed by Jin Yu-gan *et al.* (1997). Below, the Ambon and Bayonne Fms, devoid of any fossils, are separated by a basalt flow of probable (early?) Artinskian age (278 Ma). Above, the Pradineaux Fm. may constitute a condensed section: the lower part is older than 264 Ma (age of a crossing dyke) but the upper part can be correlated with the lowermost Tatarian of the Russian platform owing to ostracod studies (Lethiers *et* *al.*, 1993). This last result does not seem in conflict with paleobotanical and paleo-ichnological data obtained from the same formation (Gand *et al.*, 1995). Finally, the macro- and microflora yielded by the Muy Fm. (Bas-Argens), very similar to those from the German Zechstein, are the youngest known within the entire French Permian. Westwards correlations become more and more difficult. Nevertheless the ichnofauna found in the uppermost red-siltstone unit of each basin (Pélitique Fm. in Le Luc Basin; Gonfaron Fm. in Cuers Basin, and Fabregas Fm. in Toulon Basin) is nearly identical to the La Lieude association in the Lodève Basin (Demathieu *et al.*, 1992). Around the middle part of the Toulon series, lacustrine laminated limestones and siltstones, deposited in a narrow submeridian half-graben, yield a rich macroflora and a palynological assemblage closely similar to that from the Tregiovo Fm (Southern Alps, Italy) and for which a post-Kungurian ante-Tatarian age can be suggested (Broutin & Durand, 1995). In other respects, the possibility of correlation of the acidic tuffs described in the Saint-Mandrier Fm. (Durand, 1993) and the more distal ones discovered in the Lodève Basin (Nmila *et al.*, 1992) remains to be borne out.

From a structural point of view the Permian basins of Provence are clearly related to N-S distensional conditions (Toutin-Morin & Bonijoly, 1992), but a transcurrent element is shown by the diachronic activity of normal faults expressed by many intra-Permian angular unconformities (Baudemont, 1988). An overall depocentre migration can be traced in an E-W direction. Movement ceased before the end of the Permian and was followed by a long period of pediplanation. The oldest Triassic terrigenous deposits, located in the Toulon area, display paleocurrents in an opposite direction to those from the Permian (Durand *et al.*, 1989; Durand, 1993).

Corsica and neighbouring areas

After the Variscan event, in the continental block now defined by Corsica and Sardinia, a new orogenic event akin to the Alleghanian was accompanied by Upper Carboniferous to Lower Permian magmatic episodes. Several calcalkaline magmatic pulses made up the huge 500 km-long and 50 km-wide Corsican-Sardinian Batholith (Orsini, 1980). The older tonalitic units are interpreted as the result of renewed subduction processes (Finger & Steyrer, 1990). The younger post-collisional plutonic units were coeval with volcanic formations. The compositionally expanded high-K calcalkaline suite comprises cumulates, gabbro, diorite, porphyritic granodiorite and monzogranite, leucomonzogranite (Orsini, 1980), and an andesite-dacite-rhyolite volcanic association (Vellutini, 1977). Trace element and isotopic compositions indicate magmas derived from a depleted mantle source with a subduction component and mixed with crustal products (Cabanis *et al.*, 1990; Cocherie *et al.*, 1994).

The Permo-Triassic alkaline province was built during two episodes. The mid-Permian (270 ± 10 Ma) was a critical period marked by emplacement of numerous complexes. No time interval is substantiated between the latest high-K calcalkaline and the earliest alkaline intrusions at about 280 Ma. After a quiescent period, renewed magmatic activity took place at the Permian-Triassic boundary (245 ± 10 Ma). The magmatic rocks are exposed within caldera volcanoes, dyke swarms and ring complexes. The compositionally expanded suite (Bonin, 1986) encompasses gabbro, monzogabbro, monzonite, syenite, alkali feldspar granite and syenogranite at the subvolcanic level (Platevoet, 1990), scarce basic lava flows and voluminous trachyte-rhyolite ignimbritic flows at the volcanic level (Vellutini, 1977).

Trace-element compositions of dolerite dykes indicate derivation from an OIB source (Cabanis *et al.*, 1990). Metaluminous, slightly peraluminous or highly peralkaline granites exhibit high, yet variable, trace-element contents related to mineral fractionation and volatile transfer through F-rich complexing fluids (Egeberg *et al.*, 1993). Initial Sr isotope ratios range from 0.703 up to 0.737 (Bonin *et al.*, 1987) and εNd from +0.5 down to -5.8 (Poitrasson *et al.*, 1995). Chemical evidence precludes simple mixing processes by crustal contamination of mantle-derived magmas and indicates primary magmas similar to the Ivrea mafic complex (Sinigoi *et al.*, 1994), with an insignificant to minor contribution from old upper crustal units and a varying contribution of lower crust.

The mid-Permian episode can be considered as post-orogenic, namely post-Alleghanian. It is characterised by two magmatic alignments (Bonin *et al.*, 1987), one from Morocco to Catalonia to Corsica-Sardinia to Estérel and Briançonnais to the Southern Alps, and the second from Corsica to Vosges-Schwarzwald to the Oslo Rift, defining a Y-shaped fault system associated with large sinistral shear zones. The Late Permian to Triassic episode was markedly early anorogenic (Bonin, 1990) and was related to an incipient rifting regime in Corsica, Estérel and Catalonia. It heralded the Mesozoic evolution of the western Mediterranean area, and can also be considered eo-Alpine.

Thus, the Permian of Corsica is essentially magmatic in nature (Gondolo, 1989). Only two small outcrops at the western edge of the Monte Cintu caldera have provided paleontological data (Table 1b). In Mausoleu, near the Osani Stephanian basin mined for coal, flora revealed an "*Autunian*" age for sediments intercalated within calcalkaline andesites of the first magmatic episode. In Scandola (Capu Puppiaghia) some microfloras containing *Platysaccus* and *Lueckisporites* provide a Thuringian age for sandstones and lacustrine limestones intercalated within alkaline volcanic rocks of the second episode.

French Alps

The French Alps are located in the westernmost part of the orogen (Table 1b). The main structural Alpine units concerned are: the external zone (*Dauphiné*) and the internal zones (*Briançonnais* and *Valais*). In the Alpine segment of the Variscan belt, the Upper Paleozoic features are often difficult to decipher because of the superimposition of the Cenozoic orogenic episodes. In the external zones, some occurrences of Permian deposits are generally less deformed than their Mesozoic cover above the Triassic gypsum layers, for example in the Belledonne Massif (Deroin *et al.*, 1996). The Permian deposits are better exposed in the southern French Alps, in the Barrot and Mercantour-Argentera areas (Vinchon, 1984). Near Léouvé (Barrot dome) the discovery of *Arctotypus verneti* (Odonata) has provided a "Thuringian" age for the top of the Cians Fm. (Laurentiaux-Vieira & Laurentiaux, 1963). The same age has been proposed for the Léouvé Fm. using palynomorphs.

In the northern Alps, Savoie and Dauphiné, the Late Paleozoic deposits including Upper Carboniferous and Permian strata appear within very narrow troughs, mainly along the Belledonne granitic massif. Near Allevard, the pre-Mesozoic series is composed of a succession of: (i) black shales and coarse-grained sandstones dating from the Westphalian to the Stephanian, using macroflora; (ii) a 200 m thick sequence of sandstones (Grand-Rocher Fm.) including conglomerate and dolomite layers possibly Early Permian in age (some rare macroflora); (iii) sandstones probably Triassic as well as Upper Permian in age corresponding to the Flumet- and Allevard Sandstone Fm.). In the internal zones, the Permian deposits are widely exposed in the Briançonnais (*Zone houillère*), notably in the Vanoise and Galibier areas. Unfortunately, the amount of strain has destroyed any paleontological evidence. Thus, the Permian facies are mainly defined on structural considerations, the typical cross-section showing the "*Éopermien*" or Early Permian overlain by the "*Néopermien*", namely the Late Permian. This megasequence begins with conglomeratic layers (Roche-Château Conglomerates), probably Upper Carboniferous as well as Early Permian in age, and is overlain by the Verrucano facies, reputedly Permo-Triassic. All the "*Permian*" megasequence rests unconformably on Carboniferous or older deposits.

MAGMATISM

Carboniferous to Triassic plutonic activity

Progressive amalgamation of almost all continental terranes into the last Pangaea during the Carboniferous resulted from the complex collisional interactions of two megacontinents, Laurasia to the north and Gondwana to the south. The scattered occurrence of Cadomian ages in south Europe and eastern North America indicate that Gondwana-derived terranes, such as Avalonia, Armorica, Iberia, and so on, collided early with the Laurasia megacontinent (Ziegler, 1986). Collision of the Mid-European Moldanubian terrane with the Intra-Alpine terrane (Stampfli, 1996) resulted in contrasting plutonic episodes (Bonin *et al.*, 1998, and references therein).

In the Moldanubian terrane, the post-collisional magmatic activity consisted of short-lived syntectonic episodes of a few million years, controlled by transcurrent crustal-scale faults. (1) Mg-K-rich magmatic suites, recorded at 345-330 Ma from the External Alps to the Bohemian Massif and at 310 Ma in the Aar Massif, probably derived from a subduction-enriched lithospheric mantle source with crustal contamination. (2) Peraluminous magmas were emplaced along dextral strike-slip faults at 307 Ma in the Mont-Blanc and Aiguilles-Rouges and had dominantly crustal sources, but a mantle contribution is documented by coeval gabbro and mafic microgranular enclaves. (3) The last magmatic episode was post-orogenic and is represented in the Mont-Blanc, Central Aar and Gotthard Massifs by alkali-calcic plutonic suites at 303-292 Ma, and by abundant intermediate to acidic volcaniclastic deposits at 303-299 Ma. Isotopic evidence points to an ultimately mantle source, with an increasing contribution from the upwelling asthenospheric mantle relative to the subcontinental lithospheric mantle source, and to crustal contamination of unknown magnitude. In this intra-continental zone, magmatic activity was poorly developed during the Permian.

By contrast, in the marginal Intra-Alpine terrane, magmatic episodes were largely represented during the Permo-Carboniferous. (1) 340 ± 5 Ma Mg-K-rich magmatic suites in the Pyrenees, Corsica and the western Alps resemble those within the Moldanubian terrane. (2) During the Late Carboniferous and the Early Permian in Morocco, Catalonia, Pyrenees, Corsica-Sardinia, Provence, the Internal Alps (Internal Crystalline Massifs, Briançonnais zone, Austro-Alpine realm and Southern Alps) and south Italy, normal to high-K calcalkaline batholiths and volcanic formations derived from a subduction-enriched lithospheric mantle source with crustal contamination, but partial melting involved smaller amounts of K-rich minerals. Peraluminous anatectic massifs, well-developed in the Velay Dome of the French Massif Central, are also represented. (3) The magmatic suites became abruptly alkaline 280 Ma ago, with no smooth transition from the previous suites, and produced two episodes, the major one at 270 ± 10 Ma in Morocco, Catalonia, the Pyrenees, Corsica-Sardinia, Estérel, and the Internal Alps, and the second minor one at 245 ± 10 Ma in Catalonia, Corsica and Estérel. They derived from an OIB source replacing the older lithospheric source, with minor, yet varying, crustal contribution.

The Triassic (250-200 Ma) was marked by volcanic-plutonic complexes emplaced along fault zones, such as the 245 Ma Matterhorn-Mont Collon-Dents de Bertol transitional suite, the Austro-Alpine realm, the 237-232 Ma Monzoni-Predazzo alkaline complex, the Southern Alps, and the tholeiitic ophites of the Pyrenees. Lastly, a 200 Ma thermal anomaly reset mineral and some whole-rock isotopic clocks.

Permian volcanic activity

Significant volcanism occurred during the Permian and the earlier Triassic in the western Mediterrannean province. Two main episodes, high-K calcalkaline and alkaline, are classically recognised. They correspond to the transition from post-collisional crustal-derived or contaminated magmatism to anorogenic asthenospheric magmatism.

The first episode

The first volcanic episode including andesites, dacites and rhyolites of Early Permian age (290-275 Ma) is high-K calcalkaline with a significant crustal component. It is well-represented in the Pyrenees (Cabanis & Le Fur Balouet, 1989; Innocent *et al.*, 1994; Cabanis, 1996), the Corsica-Sardinia block (Cabanis *et al.*, 1990; Cortesogno *et al.*, 1998) and the Iberian Cordillera. Similar volcanism occurred in the Ligurian and Southern Alps (Barth *et al.*, 1993; Cortesogno *et al.*, 1998; Rottura *et al.*, 1998).

In the Pyrenees and Corsica, geochemical studies show evidence for a crustal component (Cabanis, 1996; Cabanis *et al.*, 1990). Trace-element data from andesites show a strong LILE enrichment, particularly in K, U and Th. The Th/Ta ratio (3-6) is low and differs from the typical orogenic andesite related classically to an actively subducted slab (Cabanis & Thiéblemont, 1988). The strong decrease of compatible elements (Cr, Ni, Sc, Co) with increasing Th (Th considered as a differentiation index) shows that fractional crystallisation involving spinel, olivine and clinopyroxene was the main petrogenetic process. Low La, Ti and P contents in the most basic andesites could be interpreted in terms of an important crustal contribution (Cabanis & Le Fur-Balouet, 1989). These crustal characteristics are well-illustrated in La-Y(Tb)-Nb(Ta) (Cabanis & Lécolle, 1989) in which most of the rock compositions plot within a field intermediate between the typical orogenic andesite field and the non-orogenic basalt field. Similar characteristics have been found in the Ligurian Alps and the Southern Alps (Cortesogno *et al.* 1998; Rottura *et al.*, 1998).

In the Pyrenees, isotopic studies (Gilbert *et al.*, 1994; Innocent *et al.*, 1994) confirm the importance of a crustal component. In the Sm-Nd diagram, the basic andesites of the Ossau caldera (France) and Sierra del Cadi (Spain) plot in the field of Archean to Cretaceous sediments and have the same isotopic composition as the Querigut monzogranite and granodiorite (Ben Othman *et al.*, 1984); their Nd signature yields a low initial eNd of –3 to –9 with very constant Sr isotopic ratios, showing the crustal influence. Very similar results have been found in the Southern Alps (Barth *et al.*, 1993).

The second episode

The second Permian episode (265-235 Ma) corresponds to the emplacement of bimodal volcanic products (basalts-ignimbritic rhyolite) into caldera structures during an extensional regime. It is localised in different areas in the Pyrenees: La Rhune, Ossau-Anayet (Cabanis & Le Fur-Balouet, 1989; Innocent *et al.*, 1994; Cabanis, 1996); Corsica (Cabanis *et al.*, 1990); Estérel (Gondolo, 1989; Poitrasson & Pin, 1998; Lapierre *et al.*, 1999) and Toulon (Leroy & Cabanis, 1993).

In a MORB- normalised spidergram (Pearce, 1982) basalts display a regular decrease from Th to Tb similar to alkaline or transitional basalts. Variable negative anomalies in Ta and Ti suggest a mantle-source still affected by an orogenic component. The variable Th/Ta ratios – close to 3 in Corsica, nearer 1 in the Toulon Basin and Anayet, and 3-8 in La Rhune and Estérel – are consistent with this hypothesis. Constant values of Cr, Ni, Co and Sc with a large increase in Th suggest variable degrees of partial melting in contrast with the fractional crystallisation processes of the first episode.

Isotopic data (Rb-Sr, Sm-Nd) on Pyrenean volcanism (Innocent *et al.*, 1994) agree with the trace-element study. Anayet basalts clearly originate from the melting of the asthenosphere, and La Rhune basalts have Sr and Nd isotopic characteristics similar to the CFB continental tholeiites, and could have heralded the Triassic ophite volcanism.

Position of Permian magmatism in the geodynamic setting

Although considered initially as a continuous process, magmatism actually occurred in discrete episodes accompanying or heralding major geological discontinuities. At the Carboniferous-Permian boundary (ca. 300 Ma), extensive volcanic and associated plutonic activity occurred throughout the area. It consisted of high-K calcalkaline andesite-dacite-rhyolite suites, mostly erupted from fissural vents in intramontane basins, filled with molasse deposits produced by local erosion of high relief. The mid-Permian episode (280-260 Ma) was characterised by largely complete peneplanation and widespread magmatism, yielding the famous red landscapes of the western Mediterranean coasts (Lodévois, Provence basins, etc). Volcanic-plutonic complexes and fissural formations indicate a rapid shift from the older high-K calcalkaline suites, related to the waning post-collisional stages of the Variscan orogeny, to the younger alkali-calcic to alkaline suites, implying new mantle-enriched source(s) and the onset of Pangaea break-up. Lastly, at the Permian-Triassic boundary (ca. 245 Ma) and within the Early Triassic (up to 220 Ma outside the study area), alkaline to peralkaline caldera volcanoes and ring complexes were emplaced within the future passive margins of the Tethyan ocean basin. Evidence for continental dislocation is provided by the large Triassic (ca. 220-200 Ma) thermal anomaly recorded in mineral ages of the older basement rocks.

BIOSTRATIGRAPHY

The stratigraphy of Permian terranes in southern France, Corsica, and the surrounding regions, is essentially based on continental biofossils. As a consequence the correlations with the marine time-scales, international or Tethyan, are not reliable, except in the places where continental and marine deposits are intercalated. Moreover, the continental basins frequently dried up during the Permian, leading to disconformities, gaps and erosion of the deposits, as well as the destruction of biota by oxidation. Imprints or permineralised fragments of macroflora, spores and pollen grains, ostracods and tetrapod footprints, and now conchostracans and insects, are the most important tools to have been used for stratigraphy.

Catalonian Pyrenees

The Permian is crops out near Palenca de Noves, Baro, and along the Pallerols river. The basal conglomeratic or silty sediments have been dated from the Autunian *sensu* Broutin *et al.* (1999) by macroflora (Broutin & Gisbert, 1983). In the same area, above the red Permian facies, the lower Buntsandstein facies which contains a *Lueckisporites* association has been attributed to the Thuringian (Broutin *et al.*, 1988).

Southern edge of the Massif Central

In the Lodève, Saint-Affrique, Rodez, and Largentière Permian basins, plant remains and footprints are abundant. Above the coal-bearing and generally tilted Stephanian (Rodez and Saint Affrique) or the Cambrian basement (Lodève), the Permian sediments have been divided into five formations, on the basis of sedimentological criteria. The biostratigraphic content includes flora and fauna. Macroflora and microflora (spores and pollen grains) essentially are abundant in the grey formations. Systematic studies on the flora are as old as the 18th century. The "*Walchia*" assemblages include a large range of Permian species belonging to the following genera: *Pecopteris, Callipteridium, Rhachiphyllum, Lodevia, Odontopteris, Autunia, Culmitzschia, Ginkgophyllum, Baiera, Arnhartia, Odontopteris, Cordaites* and *Supaia* (Broutin *et al.*, 1992). The age interpretation of such a flora fits very well with the paleo-ichnological results, except for the Usclas-Saint-Privas to Les Tuilières-Loiras Fms which are ascribed to the upper "Autunian". The microfloral content is likewise abundant from the base of the Usclas-Saint-Privas Fm to the middle part of the Viala Fm. It has been divided into three biozones (Lo1 to Lo3) by Doubinger *et al.* (1987), which are successively assigned to the upper "Autunian", the Artinskian and the Thuringian (Ufimian to Kazanian). These microfloral interpretations are in conflict with tetrapod and macroflora data.

The recent paleontological discoveries in the Lodève Basin improved the stratigraphical scheme by means of insects and conchostracans (Gand *et al.*, 1997 d; Nel *et al.*, 1999 a, b; Schneider *et al.*, 1999). These fossils are located in the Salagou Fm. dating from Leonardian to Tatarian (Gand *et al.*, 1997 d). From Schneider *et al.* (1999), the whole Usclas-Saint-Privas-Les-Tuilières-Loiras and Viala Fms could be Asselian in age, and the Rabejac and Salagou Fms could have a Sakmarian-Dzulfian age.

Provence

With the exception of those recently discovered in the Toulon Basin (Broutin & Durand, 1995), the study of which is still in progress, the fossiliferous horizons of Provence are restricted to the eastern basins in the Pradineaux, Mitan, Muy and La Motte Fms. They have yielded vertebrate footprints, macroflora, spores, pollen grains and invertebrate remains (mainly ostracods).

The vertebrate assemblages include the following species: *Hyloidichnus major, Dromopus didactylus, Antichnium salamandroides, Limnopus zeilleri, Varanopus rigidus* and *curvidactylus, Dimetropus latus* and *Laoporus* sp. The macroflora comprises: *Walchia* spp., *Pseudovoltzia* spp., *Sphenopteris kukukiana, Ullmannia Bronni* and *U. frumentaria*, and *Ginkgophytoxylon permiense*. Amongst the palynological content, the following genera are well-represented: *Nuskoisporites, Lueckisporites, Jugasporites, Klausipollenites, Limitisporites, Labiisporites, Taeniaesporites* and *Vittatina*. This last set is assigned to the "Thuringian" (Kazanian to Tatarian). This fits very well with the recent data from ostracods discovered in the Pradineaux Fm., giving an early Tatarian age.

Corsica-Sardinia block

In the Osani area, northwestern Corsica, some fluvial and lacustrine layers are intercalated within the volcanic and pyroclastic deposits of the first calcalkaline episode (Gondolo, 1989). Plant debris of *Calamites, Cordaites* and *Walchia* are associated with the siltstones attributed to the Autunian *sensu* Broutin *et al.* (1999). In the same area, at the edge of Scandola caldera, sandstones and lacustrine mudstones linked to the second volcanic episode contain a

Platysaccus and *Lueckisporites* association, most likely "Thuringian" in age. In Sardinia, new paleontological data (Ronchi *et al.*, 1998) have recently enhanced the dating of "Autunian" sediments in the Guardia Pisano area (Barca *et al.*, 1991). Some additional sections in the north (Lu Caparoni) and the southeast (Perdasdefogu, Escalaplano) have yielded plant remains, spores, pollen grains, algal colonies and ostracods, which have enabled the correlation of such fossiliferous sections not only with others within Sardinia, but also with the eponymous Autun Basin and, further to the west, with the Gerri de la Sal section in Spain.

CONCLUDING REMARKS

1. The Permian of southern France is famous for its volcanic formations, like many other Mediterranean areas. The first magmatic episode is mainly high-K calcalkaline in character, and provides evidence for a significant crustal component. The second episode is mainly alkaline and derived from an OIB source replacing the older lithospheric source with minor crustal contribution. It has a widespread distribution (Morocco, Catalmonia, Pyrenees, Corsice-Sardinia, Provence, Alps). In most areas (Corsica, Pyrenees, Massif Central, etc.) the Permian tectonomagmatic event was reflected by the magmatic episode at 270 Ma and tectonism (Bonin *et al.*, 1993; Broutin *et al.*, 1994; Deroin *et al.*, 1990). The Permian-Triassic boundary is marked by scarce (per)alkaline magmatic complexes (Catalonia, Corsica, Estérel).

2. The Permian sedimentary facies cannot be used as a stratigraphic marker. An example of evident diachronism is provided by the widespread transition from grey to red facies. In other respects carbonate deposits, including stromatolites (Freytet *et al.*, 1999), appear as peculiar facies scattered in space and time (Toutin-Morin, 1992). Complementary studies call into question some correlations between volcanic units; in Le Luc Basin, for instance, a rhyolite believed to be coeval with the A7-rhyolite from the Estérel showed a normal magnetic polarity component (Merabet & Daly, 1986) which was not detected in the typical A7. Even biostratigraphical data give rise to debate. Conclusions based on macroflora, palynomorphs and vertebrate footprints may be in conflict (*e.g.* Broutin *et al.*, 1992), possibly owing to varying ecological and taphonomical conditions (Broutin *et al.*, 1990). That is why there remain many discrepancies in correlations between the different basins of southern France.

3. From a paleogeographical point of view, it is important to emphasise the very close facies relationships displayed on the one hand by the Permo-Triassic siliciclastic sequences in the Toulon Basin (Durand, 1993; Durand *et al.*, 1989), and on the other hand by the "*Verrucano Sardo*" in the Nurra region, NW Sardinia (Cassinis *et al.*, 1996). Furthermore the Autunian paleoflora recently discovered in the black-grey shaly deposits from Perdasdefogu, SE Sardinia, is very similar to the one from the Gerri de la Sal succession, Catalonian Spanish Pyrenees (Ronchi *et al.*, 1998). These data are in full agreement with the paleoposition of the Corsica-Sardinia "block" stretching E-W, in a palinspastic reconstruction such as that proposed by Broutin *et al.* (1994) in a paleogeographical reconstruction based on structural and magmatic data. In such a model, the Permian volcanics of northeastern Corsica are facing those of the Estérel massif, and the southern part of Sardinia is linked with the eastern Pyrenean domain, the Late Paleozoic setting of which is harder to decipher.

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SYNSEDIMENTARY VOLCANISM IN THE LATE CARBONIFEROUS SALVAN-DORÉNAZ BASIN (WESTERN ALPS)

NICOLA CAPUZZO [1] and FRANÇOIS BUSSY [2]

Key words – Western Alps; Aiguilles-Rouges; Late Carboniferous sediments; Salvan-Dorénaz basin; volcanism; fall-out deposits.

Abstract – The Salvan-Dorénaz basin is a NNE-SSW elongated intracontinental trough located in the Aiguilles-Rouges crystalline massif (Western Alps), which formed during the Late Carboniferous post-orogenic evolution of the Variscan fold-belt. Only part of the original basin is preserved with a maximum thickness of 1.5 – 1.7 km of continental sediments, represented by alluvial fan, alluvial plain, lacustrine and river deposits. Four different lithological units formed in response to intrabasinal differential subsidence and tectonic movements: Unit I, at the base, corresponds to the emplacement of the basin and is composed of alluvial fan and braided river deposits; Unit II consists of palustrine and anastomosed river deposits; Unit III comprises meandering river deposits; and Unit IV consists of alluvial fan deposits interfingering with Units II and III and spreading into the fluvial basin from the northwestern margin.
Syndepositional volcanism is documented in the Salvan-Dorénaz basin by different volcanic and volcano-sedimentary layers, for which facies analysis and compositional studies revealed primary as well as reworked characters for these products. In particular, basal subaerial flows and autobreccia deposits are possibly related to the emplacement of a rhyodacitic lava dome along the northwestern margin of the basin.
On the other hand, several tuffs and resedimented volcaniclastic layers found within sediments of Units II and III testify periods of high-explosive volcanism from distant volcanic vents. Facies, petrographical and zircon typological analyses suggest coexistence of different magma sources during synsedimentary volcanism.

Parole chiave – Alpi Occidentali; Aiguilles-Rouges; sedimenti tardo-carboniferi; bacino di Salvan-Dorénaz; vulcanismo; depositi da "fall-out".

Riassunto – Il bacino continentale di Salvan-Dorénaz si imposta nel Carbonifero Superiore sul basamento cristallino del Massiccio dell'Aiguilles-Rouges (Alpi Occidentali) durante la fase post-collisionale dell'orogenesi Varisica. Allungato in direzione NNE-SSW presenta uno spessore massimo di 1.5-1.7 km di depositi continentali costituiti da associazioni di facies di conoide alluvionale, piana alluvionale, lacustri e fluviali. Quattro unità lito-stratigrafiche si formarono in seguito a intensa attività tettonica sindeposizionale e a differenti tassi di subsidenza intrabacinale. L'Unità I, alla base, corrisponde alle fasi iniziali di formazione del bacino ed è composta da facies di conoide alluvionale e di corsi d'acqua intersecantisi; l'Unità II consiste in sedimenti palustri e depositi di fiumi anastomizzati; l'Unità III è caratterizzata da sedimenti di fiumi meandriformi, e l'Unità IV consiste in conoidi alluvionali sviluppatesi dal lato nord-occidentale del bacino, all'interno delle Unità II e III. L'attività vulcanica sinsedimentaria è documentata da vari livelli di vulcaniti e vulcanoclastiti, per le quali l'analisi di facies e lo studio composizionale hanno rilevato caratteri di prodotti sia primari sia rimaneggiati. In particolare, colate subaeree e depositi di autobreccia che affiorano alla base della successione sedimentaria si formarono possibilmente in relazione alla messa in posto, lungo il margine nord-occidentale del bacino, di un duomo vulcanico di composizione riodacitica. D'altra parte, numerosi livelli di cineriti e di vulcanoclastiti presenti nei sedimenti delle Unità II e III testimoniano un vulcanismo altamente esplosivo, proveniente da edifici vulcanici distanti. L'analisi di facies, lo studio petrografico e la tipologia degli zirconi dei diversi prodotti vulcanici suggeriscono coesistenza di varie sorgenti di provenienza del magma.

INTRODUCTION

The Variscan orogeny in Europe is the result of the complex and oblique continental collision between Gondwana to the south and Laurasia to the north. Thrust tectonics and continental subduction led to a thickened continental crust and a Barrowian-type metamorphism culminating during Mid-Carboniferous times in western and southern Europe (Ziegler, 1986; Stampfli, 1996; von Raumer, 1998). Current geodynamic reconstruction for the Carboniferous infers a consolidated continent-continent fold-belt in the western part of the Variscan orogen, whereas its eastern

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part would still be characterized as an active margin with oblique subduction of Palaeotethys (Ziegler & Stampfli, 1999 in this volume).

The anti-clockwise rotation of Africa relative to Europe during the Late Carboniferous induced the development of crustal-scale dextral strike-slip zones in the consolidated part of the Variscan belt (Arthaud & Matte, 1977; Ziegler, 1990) (Fig. 1), where post-orogenic readjustment process-

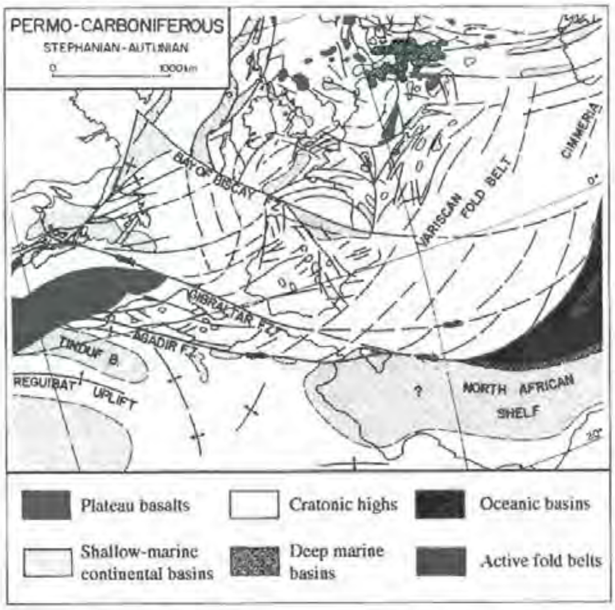

Fig. 1 – Late Carboniferous framework illustrating the western part of the Variscan fold belt, after Ziegler (1990). Abbreviations: B - basin, EUB - El Biot Uplift, FZ - fracture zone.

es of the thickened crust took place (*e.g.* Burg *et al.*, 1994, and reference therein). As a result, numerous strike-slip and pull-apart basins formed in central and southern Europe (Ziegler, 1990; Cassinis *et al.*, 1992; Krainer, 1993). The main characteristics of this post-orogenic development were: (a) regional uplift with erosional processes that reached mid-crustal levels; (b) localized areas of intense tectonic subsidence filled with thick clastic continental series; and (c) widespread intrusive and extrusive magmatism of mantle and crustal derivation (Benek *et al.*, 1996; Bonin *et al.*, 1993; Cortesogno *et al.*, 1998; Bussy *et al.*, in press).

This volcano-sedimentary cycle is recognized in both the southern and western regions of the European Variscan orogen, for instance the Collio and Tregiovio basins in the Southern Alps (Cassinis & Neri, 1990), Tödi basin in the Aar Massif (Franks, 1966; Schaltegger & Corfu, 1995), Salvan-Dorénaz (Sublet, 1962; Niklaus & Wetzel, 1996) and Pormenaz (Lox & Bellière, 1993; Dobmeier & von Raumer, 1995) basins in the Aiguilles-Rouges massif, Saar-Nahe basin in central Germany (Schäfer & Korsch, 1998). These intramountain and perimountain basins rest on top of the orogenically deformed and erosionally truncated Variscan basement and are fault bounded. They subsided rapidly and were isolated from each other by metamorphic or igneous structural heights. As a general characteristic, they are strongly controlled by the relation between tectonic, volcanic and sedimentary processes.

This first Late Carboniferous-Early Permian tectonosedimentary cycle is separated from a second Late Permian-Early Triassic post-tectonic cycle by a marked region-

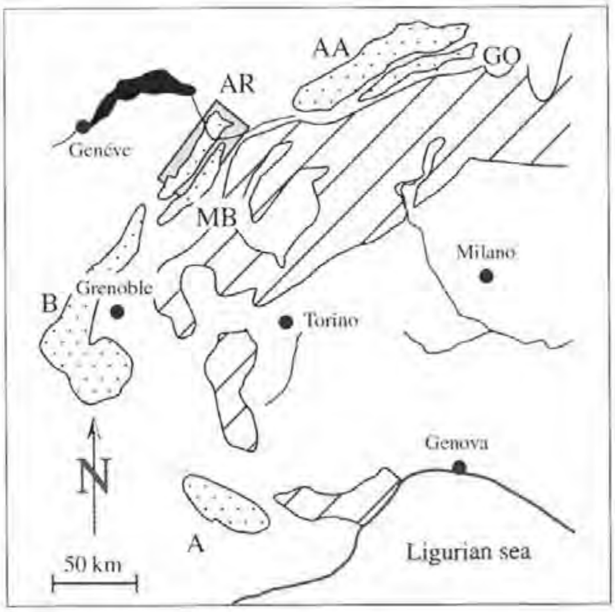

Fig. 2 – Pre-Mesozoic basement in the Alpine External Massifs (stippled). AA: Aar; Go: Gotthard; AR: Aiguilles-Rouges; MB: Mont-Blanc; B: Belledonne; A: Argentera. Study area is shaded.

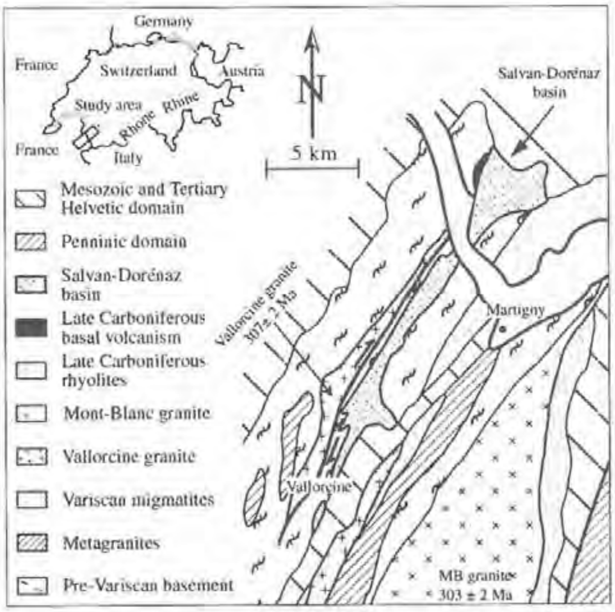

Fig. 3 – Schematic geological map of the northern part of the Aiguilles-Rouges and Mont-Blanc massifs, modified after Brändlein *et al.* (1994); insert show location of the study area.

al unconformity associated with a time-gap of uncertain duration (Italian IGCP Group (ed.), 1986; Cassinis *et al.*, 1988; Massari, 1988).

The aim of this paper is to describe and characterize new evidences of syndepositional volcanism found at different levels of a well-exposed continental sedimentary basin outcropping within the Aiguilles-Rouges massif, which developed during the Late Carboniferous – Early Permian tectono-sedimentary cycle.

REGIONAL GEOLOGICAL SETTING

The Salvan-Dorénaz basin formed during the Late Carboniferous (Late Westphalian – Stephanian; Jongmans, 1960) as an intramountain trough within the crystalline and metamorphic basement of the Aiguilles-Rouges massif. The latter is one of the so-called External Crystalline Massifs of the Alps (Fig. 2), which were part of the internal zone of the Variscan belt (von Raumer, 1998).

The Salvan-Dorénaz basin is a 20 km long by 4 km wide, NNE-SSW oriented trough, bounded by crustal-scale faults (Fig. 3). It hosts 1.5 to 1.7 km in thickness of clastic sediments (Fig. 4) deposited in continental environments. They represent only part of the original series, which were partly obliterated during an intense erosion and peneplanation stage linked to a Late Permian tectonic inversion with possible tilting of the basin (Fig. 5). During Alpine orogeny, the basin was deformed into a complex syncline structure (Pilloud, 1991; Badertscher & Burkhard, 1998), and Alpine metamorphic overprint in the Upper Carboniferous sediments only reached anchizone grades, as documented by illite crystallinity measurements on mudstones by Pilloud (1991).

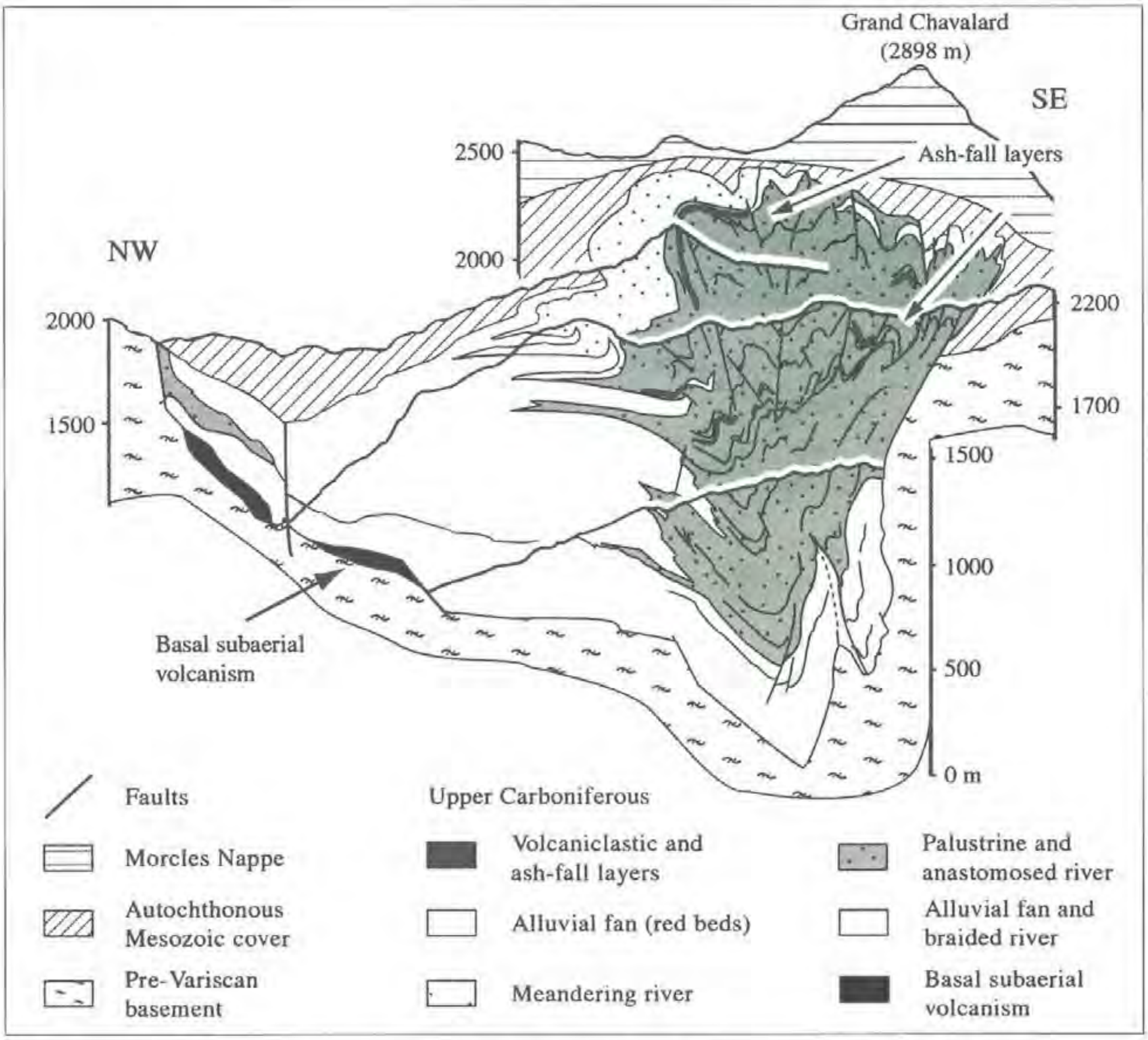

Fig. 4 – Multiple cross sections of the northeastern part of the Salvan-Dorénaz syncline, not exaggerated, modified after Pilloud (1991).

This sedimentary trough displays an asymmetric infill with a shifting depocenter located next to the northwestern side, a thick sedimentary series in comparison to basin area, rapid lateral and vertical facies variations, and high-average subsidence rates. All of these characteristics are typical for basins formed in strike-slip tectonic settings (Nilsen & Sylvester, 1995).

The formation of the Salvan-Dorénaz basin started at the very end of the Westphalian, as indicated by paleo-flora assemblages (Jongmans, 1960) and new radiometric age data on volcanic zircons (Capuzzo & Bussy, 2000). Large amounts of coarse-grained clastic deposits were transported from the northwestern margin via wet alluvial fans with dominant sediment-gravity flow deposits located proximal to sources of clastic material. These mass flow deposits progressively change downstream into high-energy shallow-stream flows characteristic of braided river facies associations.

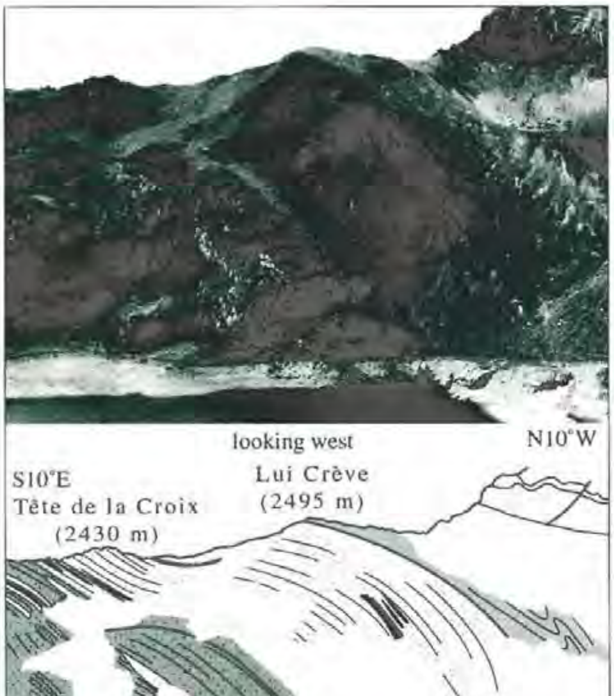

Fig. 5 – Outcrop profile and geological interpretation of the upper part of the Salvan-Dorénaz basin, Montagne de Fully area, illustrating lithologic Unit II and Unit III, and interfingering of the distal part of Unit IV with Unit II and III. The low angular unconformity between Unit III and the transgressive Triassic deposits crops out just below Lui Crève. LC-Late Carboniferous sediments. Description of the different Units in text.

The average grain-size of the sediments drastically decreases up-section. Facies assemblage and architectural elements of Unit II display characteristics of low sinuosity streams with consolidated, fixed banks, which flowed in a mud-dominated and vegetated fluvial plain. In Unit III, on top of the anastomosed river and alluvial plain deposits, the average grain-size of the sediments increases. Typically, channalized architectural elements which are characteristic of meandering fluvial channels occur. They show an increased length/width ratio and lateral accretion surfaces which testify to the migration of the channels along the fluvial valley. Sediments of Unit III are capped unconformably and at low angle by Triassic shallow-marine transgressive facies (Fig. 5).

Unit IV corresponds to dry alluvial fan deposits prograding into and retreating from the upper part of Unit II and Unit III, entering the fluvial system from the northwestern margin of the basin. Coarse-grained conglomerates of mass flow deposits represent proximal areas to the basin margin, while sheet-flood sediments dominate distal parts of the fan. The alluvial fan of Unit IV differs from that of Unit I in its reddish matrix color, possibly due to drained environments during early diagenesis, with localized oxidizing conditions.

SYNDEPOSITIONAL VOLCANISM

Basal volcanism

The initial stage of development of the sedimentary basin is marked by a period of subaerial volcanism ranging in composition from dacites to rhyodacites. It is characterized by variable proportions of coherent and autoclastic scoriaceous volcanic facies, interlayered toward the top with the Upper Carboniferous sediments (Pilloud, 1991; and own observations). The coherent facies consists of solidified lava flows having a porphyritic texture, and typically enclosed by a carapace and floor of autobrecciated products. The latter implies non-explosive fragmentation of the flowing lava commonly affecting the outer surfaces (top, base, sides) of the lava flows (McPhie *et al.*, 1993). The autoclastic facies consists of monomict, clast-supported, matrix poor, and poorly sorted breccias which are composed only of volcanic clasts (Fig. 6A), and produced by the brittle response to stress of the cooler and more viscous outer part of the flow. Typical bread-crust structures are visible in the coarser blocks and in the outer parts of the flows (Fig. 6B) and denote rapid cooling and quenching in the surface environment. These deposits were only found along the northwestern margin of the basin and they have a maximum thickness of 50-70 m, measured in non-continuous sections. They could represent the emplacement of a subaerial lava dome along the faulted margin of the basin. This was a common scenario during the evolution of the Late Palaeozoic first sedimentary cycle (*e.g.* Seui basin, Sardinia, Cortesogno *et al.*, 1998; Collio basin, Southern Alps, Breitkreuz *et al.*, 1999).

In thin section the coherent lava flows display porphyritic textures with large euhedral and subhedral phenocrysts of quartz and plagioclase (Fig. 7A), with subsidiary altered biotite flakes and few k-feldspar dispersed in a microcrystalline groundmass. Accessory minerals such as garnet, apatite, zircon and opaque minerals also occur. Quartz frequently shows perlitic fractures and typical resorption features (Fig. 7A) due to increased SiO_2 solubility during rise and eruption of the magma (McPhie *et al.*, 1993). Coarse-grained garnets are interpreted as xenocrysts pointing to crustal contamination from desegregating wall rocks during intermediate stages of the ascending magma. Fine-grained quartz reaction rims around garnets (Fig. 7B) evidence chemical disequilibrium with the melt. Both pheno- and xenocrysts are frequently cracked as a result of shear during flow and/or pressure release during rise and eruption. Zircons separated from coherent lava flows yielded an U/Pb isotopic age of 308 Ma (Capuzzo & Bussy, 2000), interpreted as the oldest possible age for the emplacement of this subaerial volcanism. Zircon morphology (Pupin, 1980) is typical for crustal derived S-type magmas (Fig. 14).

Numerous rounded volcanic pebbles, texturally and petrographically similar to the basal volcanic deposits, were found in the lower conglomeratic levels of Unit I. In thin section, these rhyolitic pebbles show eutaxitic texture with different welded pumice fragments and devitrified glass shards (Fig. 8); quartz and plagioclase phenocrysts are frequently aligned along flow direction. Coarse-grained garnet xenocrysts also occur. These volcanic rocks were probably erupted as ignimbritic flows in the source areas of detrital material, and subsequently eroded and transported through alluvial fan systems into the Salvan-Dorénaz basin. Their textural and compositional characteristics and their tectonically undeformed fabric suggest

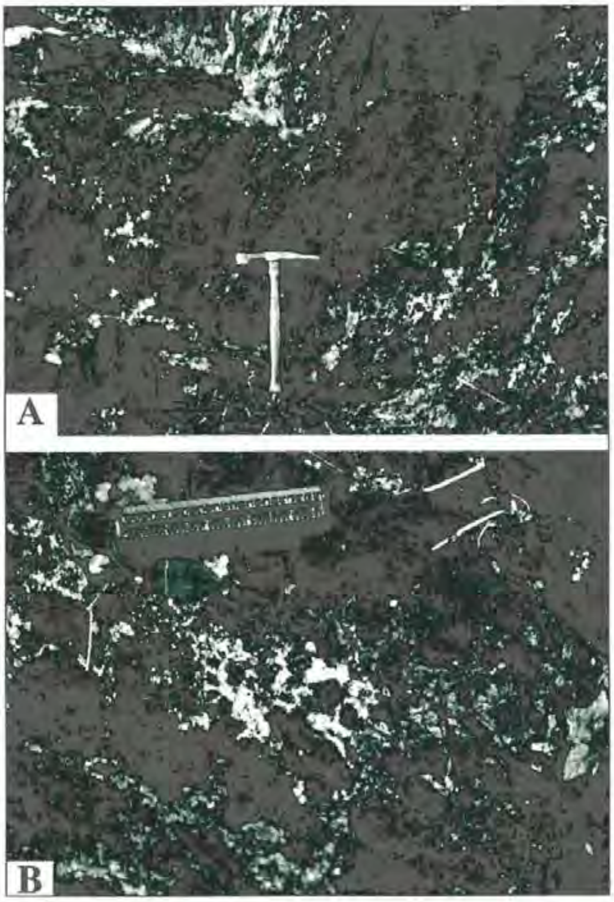

Fig. 6 – A) Autoclastic breccia from the basal rhyodacitic volcanism; B) Bread-crust structure denoting rapid cooling and quenching, from outer part of a volcanic flow. Both photographs were taken in the area of Plex (coord. 113800/570100).

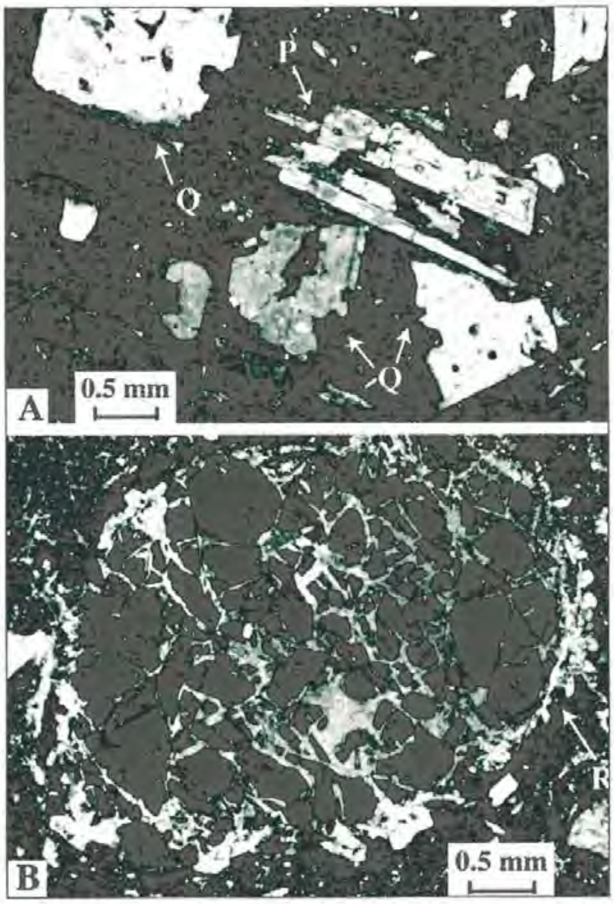

Fig. 7 – Microphotographs of the basal rhyodacitic volcanics. A) Porphyritic texture with resorbed quartz (Q) and broken plagioclase (P) phenocrysts; B) Garnet xenocryst surrounded by a microcrystalline quartz reaction rim (R). Both photographs were taken with crossed polars.

a direct link with the local magmatic pulse represented by the Salvan-Dorénaz basal volcanic deposits and by the shallow-seated Vallorcine granite (Brändlein *et al.*, 1994).

Primary to re-worked volcaniclastic deposits

Several tuff and tuffaceous sandstone layers were recently identified in the alluvial plain and overbank deposits belonging to the Salvan-Dorénaz Units II and III. Their facies and composition revealed both primary and reworked characters.

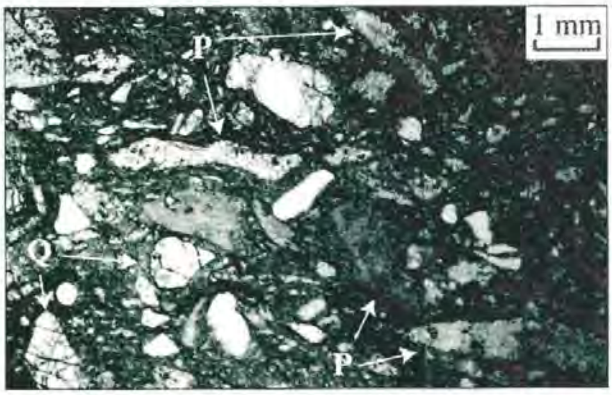

Fig. 8 – Microphotograph of a rhyolitic pebble having eutaxitic texture with broken quartz (Q) and plagioclase phenocrysts and silicified welded pumice (P) fragments. Pebble from Unit I conglomeratic alluvial fan deposits, Trient area (coord. 101175/564435). Plane-polarized light.

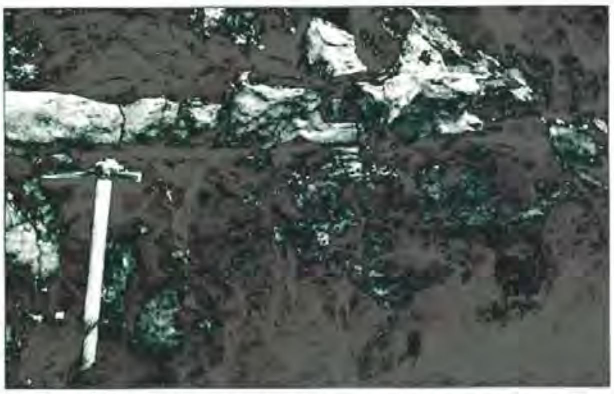

Fig. 9 – Tuff layer deposited from volcanic-ash fallout and intercalated into black alluvial plain mudstone, Montagne de Fully area (coord. 112805/572423). Hammer is 37 cm long.

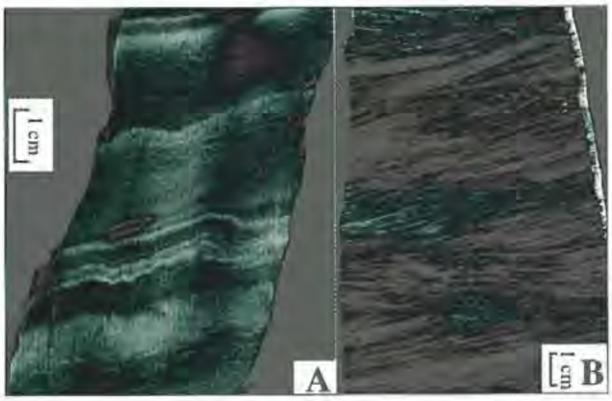

Fig. 10 – A) Delicate planar lamination partly disturbed from water escape structures. Polished surface of a silicified tuff level from the Lac Inférieur de Fully area (coord. 113252/573895); B) Cross-stratification from climbing ripples in a fine-grained volcaniclastic sandstone, Tête du Portail area (coord. 112040/572935).

Tuffs from volcanic fallout

Fine to very fine-grained tuffs range from 5 to 70 cm in thickness and show lateral continuity, mantle bedding, delicate planar laminations and fining-upward tendency (Figs 9 and 10A). They were deposited from volcanic-ash fallout and therefore are considered as primary pyroclastic deposits. They occur within black mudstone and micaceous siltstone characteristic of low energy alluvial plain and shallow lacustrine depositional environments, which allowed for their preservation (Fig. 11A). Therefore they were more frequently found within the alluvial plain and anastomosed river deposits of Unit II. Microscopical analyses revealed that they are mainly composed of a fine-grained matrix with dispersed chloritized glass shards, few fragmented quartz and plagioclase crystals, biotite flakes, and opaque and heavy minerals (Fig. 12A). The cryptocrystalline matrix is intensely silicified and might contain some carbonate patches. The main mineral phases determined with powder X-ray diffraction analyses are quartz, albite, illite and clinochlore, while their Ni vs. Zr/Ti trace element ratio (Winchester *et al.*, 1980) is consistent with a magmatic origin (Fig. 13). Textures and composition all suggest derivation from silica-rich, highly explosive volcanic eruptions from distant volcanic vents, and zircon U/Pb geochronological analyses pro-

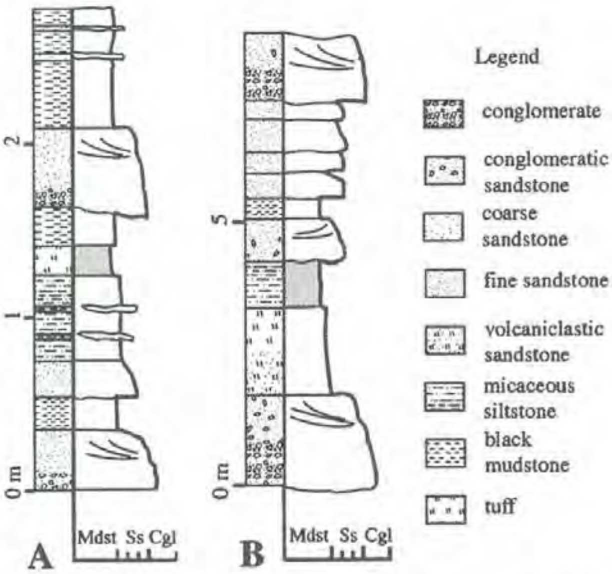

Fig. 11 – A) Stratigraphic section of a tuff layer intercalated into black mudstone. Alluvial plain deposits interbedded with crevasse splays and overbank fines (Unit II). Ash-fall layer shown in Fig. 9. B) Stratigraphic section of a reworked volcaniclastic sandstone in crevasse splay deposits from the upper Unit III, Montagne de Fully area (coord. 114100/572585).

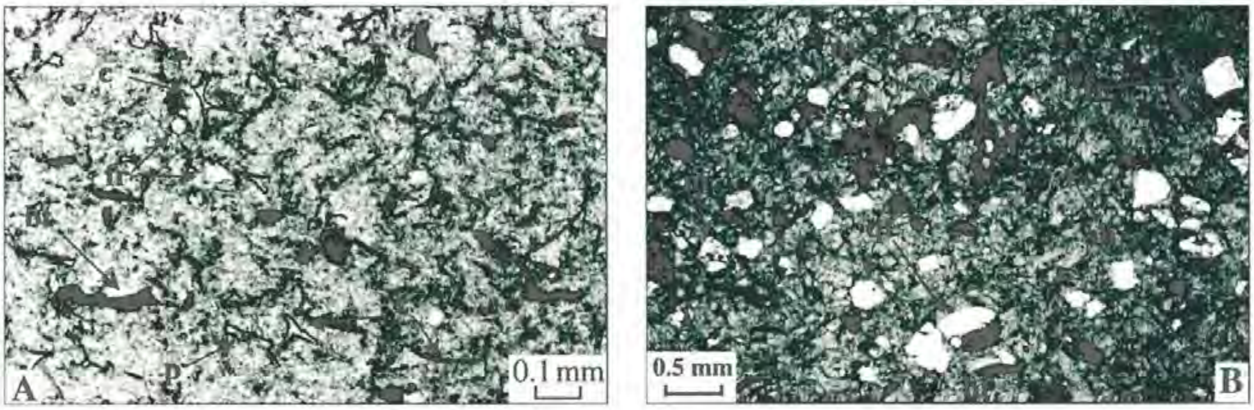

Fig. 12 – A) Microphotograph of a fine-grained tuff layer, mainly composed of light gray fine ash with dispersed cuspate (c) and platy (p) glass shards, few biotite (bt) flakes and plagioclase and quartz crystal fragments (fr); B) Microphotograph of a matrix-supported volcaniclastic layer, composed of glass shards (sh), crystal fragments (fr), biotite flakes (bt), polycrystalline quartz (qt) and metamorphic (m) fragments, denoting their mixed volcanic and detrital nature. Both photographs in plane polarized light.

duced a consistent age for this synsedimentary volcanism at 295 +4/-3 Ma (Capuzzo & Bussy, 2000). Bimodal zircon morphology distribution (Pupin, 1980) suggests derivation from alkaline magma series with crustal contamination (Fig.14).

Volcaniclastic sandstone

Strong volcanic influence during sedimentation is also recorded from several sandstone layers, which consist of mixed volcaniclastic and detrital grains and sometimes provide evidence of reworking by traction currents (Fig. 10B). They range in thickness from 50 to 160 cm and are interbedded with siltstone and sandstone layers interpreted as overbank and crevasse splay deposits (Fig. 11B). They mainly occur within Unit III. Their fabric is matrix- to grain-supported with clastic components mainly derived from volcanic sources. Broken devitrified glass shards, monocrystalline quartz and plagioclase fragments, and biotite flakes are abundant (Fig. 12B), while micaschist and sillimanite-bearing metamorphic clasts, polycrystalline quartz, intensely altered K-feldspar and white mica flakes indicate provenance from detrital sources. Their Ni vs. Zr/Ti trace element ratio (Winchester *et al.*, 1980) is also consistent with a prevalent magmatic origin (Fig. 13).

CONCLUSIONS

The Salvan-Dorénaz basin is a tectono-sedimentary trough, which formed towards the end of the Palaeozoic (Westphalian) within the Aiguilles-Rouges crystalline basement (Western Alps). Different periods of active synsedimentary volcanism occurred during formation and evolution of the Salvan-Dorénaz basin. Basal rhyodacitic flows and autobrecciated products localized along the

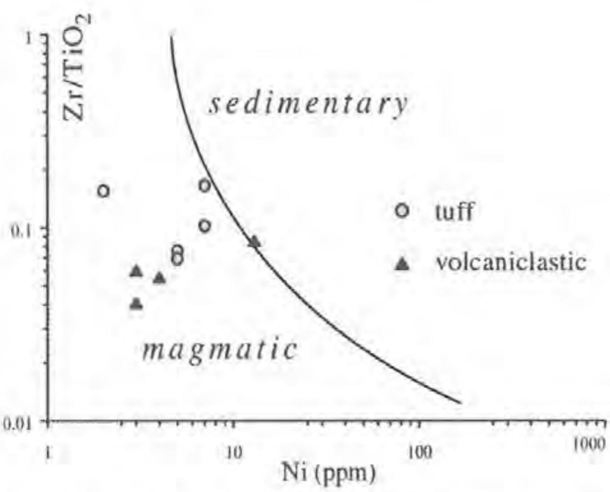

Fig. 13 – Discrimination diagram between sedimentary vs. volcanic origin of detrital material based on Ni vs. Zr/Ti trace element concentrations (after Winchester *et al.*, 1980); all tuffs and volcaniclastic layers plot concordantly into the magmatic field.

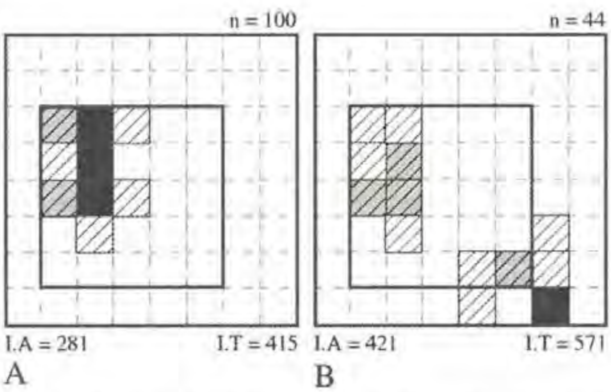

Fig. 14 – Typologic distribution of zircon for the syndepositional volcanism in the Salvan-Dorénaz basin, mean A/T values as in the classification diagram of Pupin (1980). A) basal rhyodacitic volcanism presenting an homogeneous population characteristic of anatectic magmas; B) tuff layer presenting a bimodal distribution of zircons with a maximum in the alkaline magma series.

northwestern margin of the basin were deposited during its initial stage of development. Zircon typology suggests crustal derivation for this subaerial volcanism, while petrographical studies provide evidence of contamination from disintegrating wall-rocks. An estimated age of emplacement at 308 ± 3 Ma was determined from U/Pb geochronological analyses on zircons (Capuzzo & Bussy, 2000). This lower volcanism is possibly associated with a local magmatic phase documented in the Aiguilles-Rouges massif by numerous subvolcanic dykes and by the shallow-seated, anatectic Vallorcine granite (Brändlein *et al.*, 1994), which intruded syntectonically along a dextral transcurrent shear-zone at 307 Ma (Bussy *et al.*, in press). On the other hand, ash-fall and volcaniclastic layers found within alluvial plain and palustrine sediments of Unit II and III testify for high-explosive volcanic eruptions from distant volcanic centers, and a consistent age of 295 +4/-3 Ma was determined from U/Pb geochronological analyses on zircons (Capuzzo & Bussy, 2000). Their zircon typology presents a bimodal distribution, which suggests derivation from alkaline magma series differently contaminated by crustal material. Coeval, highly explosive volcanism is known from the Aar massif in the Central Alpine basement (Schaltegger & Corfu, 1995), and tuff layers associated with this magmatic event have already been described in a Permo-Carboniferous basin located in northern Switzerland (Schaltegger, 1997).

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STRATIGRAPHICAL CONSTRAINTS ON MOLASSE DEPOSITIONAL SYSTEMS IN THE PERMO-CARBONIFEROUS SAAR-NAHE BASIN, GERMANY

ANDREAS SCHÄFER[1]

Keywords – European late Variscan sedimentary basin; Permo-Carboniferous; siliciclastics and volcanites; continental depositional environments; strike-slip basin model.

Abstract – The Saar-Nahe Basin formed as an internal rift within the Central European Variscides. It may have communicated with other freshwater basins, interconnected by depositional systems during the Westphalian and the Stephanian in tropical to subtropical latitudes, and in the Rotliegend in arid latitudes; marine deposits are not known. Basin fill of about 7.5 km thickness today, consisting of siliciclastic rocks together with rhyolitic/andesitic volcanic rocks, exhibits a well documented standard for continental stratigraphy of the Central European Permo-Carboniferous.

Parole chiave – bacino sedimentario tardo-varisico europeo; Permo-Carbonifero; rocce silicoclastiche e vulcaniche; ambienti deposizionali continentali; modello di bacino a *strike-slip*.

Riassunto – Il Bacino della Saar-Nahe si formò come un *rift* interno alle Varisidi centro-europee. Esso può aver comunicato con altri bacini d'acqua dolce, collegati da sistemi deposizionali durante il "Westfaliano" e lo "Stefaniano", a latitudini da tropicali a subtropicali, e nel Rotliegende a latitudini aride. Depositi marini sono sconosciuti. Il riempimento del bacino con depositi oggi potenti all'incirca 7.5 Km, che consistono di rocce silicoclastiche e vulcaniti riolitiche/andesitiche, rappresenta un ben documentato standard per la stratigrafia continentale del Permo-Carbonifero centro-europeo.

GENERAL SETTING

The Saar-Nahe Basin is one of the largest and best exposed sedimentary basins within the Variscan orogenic belt in Central Europe (Schäfer, 1989; Korsch & Schäfer, 1991, 1995; Schäfer & Korsch, 1998). It extends over an area of 300 x 100 km, within which it is exposed from underneath the Mesozoic cover by a size over 100 x 40 km. As drill-well evidence and seismic control is extraordinary, a three-dimensional basin model can be constructed.

About 7.5 km thick non-marine molassoid sediments, consisting of 0.5 km Namurian, 2 km Westphalian, 3 km Stephanian, and 2 km Lower Permian Rotliegend strata, were preserved in a wholly structure-controlled setting in the Saxothuringian Zone (Schäfer & Korsch, 1998). The Saar-Nahe Basin is one of the locations of Central Europe, where continental sequences are exposed internal to the Variscan orogen, providing a complete stratigraphic standard for major parts of the Permo-Carboniferous.

STRUCTURE OF THE BASIN

The basin is separated from the Rhenohercynian to the north by the South-Hunsrück Fault, which is part of one of the major Variscan suture zones, running through Central Europe (Fig.1). From the deep-reflection seismic line DEKORP 1C Korsch & Schäfer (1991) argued that the fault plane is more or less planar in shape and steep, cutting most of the continental crust. Also, there are opinions, that, from a previous continental collision (the Rhenohercynian in the N with the E Avalonia continental crust underneath versus the Mid German Crystalline Rise belonging to Armorica with the Saar-Nahe Basin above), the fault plane could be listric (Oncken, 1997).

The basin developed as a dextral strike-slip controlled half-graben with southwestward-oriented oblique transtension (Korsch & Schäfer, 1991, 1995). At the northeastern end of this fault with respect to the Saar-Nahe Basin, well exposed at the surface, shingled alluvial fans demonstrate the dextral strike-slip character of the basin – the youngest fans shingle on top of each other toward the NE (Schäfer & Korsch, 1998). In the subsurface, isopach maps compiled from thick stratigraphic sequences also show a shift of their depocentres from the SW towards the NE.

During its formation as a rift basin, internal to the Variscan orogen during the Permo-Carboniferous, the Saar-Nahe Basin is assumed to have been located in a mountain-

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ous area, (?) some hundreds of metres above sea-level. This could explain the rigid dewatering regime applied throughout the Westphalian (toward the S), the Stephanian (toward the NE), and the Rotliegend (toward the SW) - the latter two running in the Variscan axis. The distance from the sea during the Permo-Carboniferous could have easily been several hundreds of kilometres (Tethys, Biscay). Both topographic height and distance from the sea demand a freshwater regime in the basin. The perennial water-rich depositional environments of the basin formed greyish sediments in a tropical climate during the Westphalian and the Stephanian (from paleomagnetic results the latitude was 10° N in the Stephanian C; Schäfer & Stamm, 1989); in the Rotliegend about 30° N is assumed from the alluvial ephemeral environments and reddish sediments (for an overview see Schäfer & Korsch, 1998).

SEDIMENTS DEPOSITED

The sediment-fill of the Saar-Nahe Basin is siliciclastic in origin (Schäfer & Korsch, 1998).

During the Westphalian (Fig. 2), sediments derived from the Rhenohercynian Schiefergebirge in the north; as a consequence, greywackes formed. These were meandering-fluvial to deltaic sediments in the basin; close to the northern margin, alluvial fans are assumed, although they are not exposed.

From the Stephanian onwards, the source of the sediments was the Moldanubian in the south and southwest of the basin, in the Black Forest, in the Vosges, and in the Massif Central in France (Schäfer, 1986). All basement units provided detritus to form arkoses. The sediment input was fluvial, mostly of a meandering character. Clear-

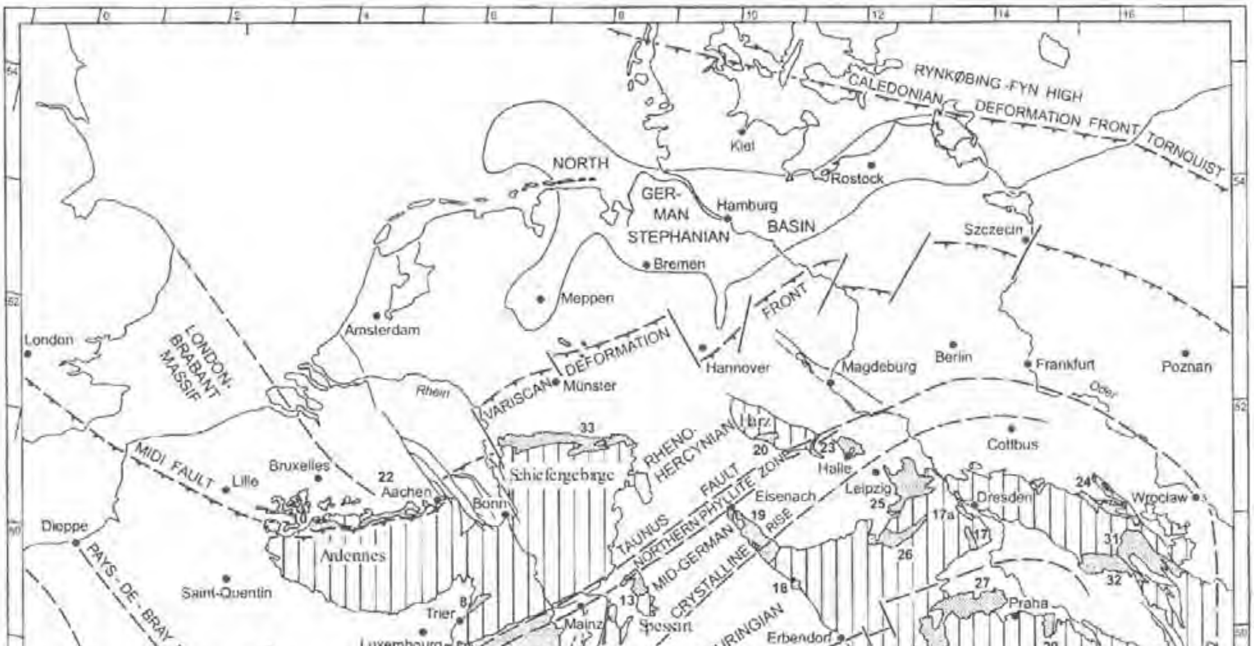

Fig. 1 – Map of Central Europe showing the Variscides together with Permo-Carboniferous continental rocks in surface exposures, mostly part of larger sedimentary basins (modified from Schäfer & Korsch, 1998). The shaded area of the North German Stephanian Basin (Ziegler, 1990) marks the central part of the depocentre of the North German Permo-Carboniferous Basin. The path of the Variscan deformation front is from Franke (1992). The codes B4, B5, C4, C5, D4, D5 are the sheet numbers of the International Geological Map of Europe 1:1,500,000 used for the compilation of this overview.

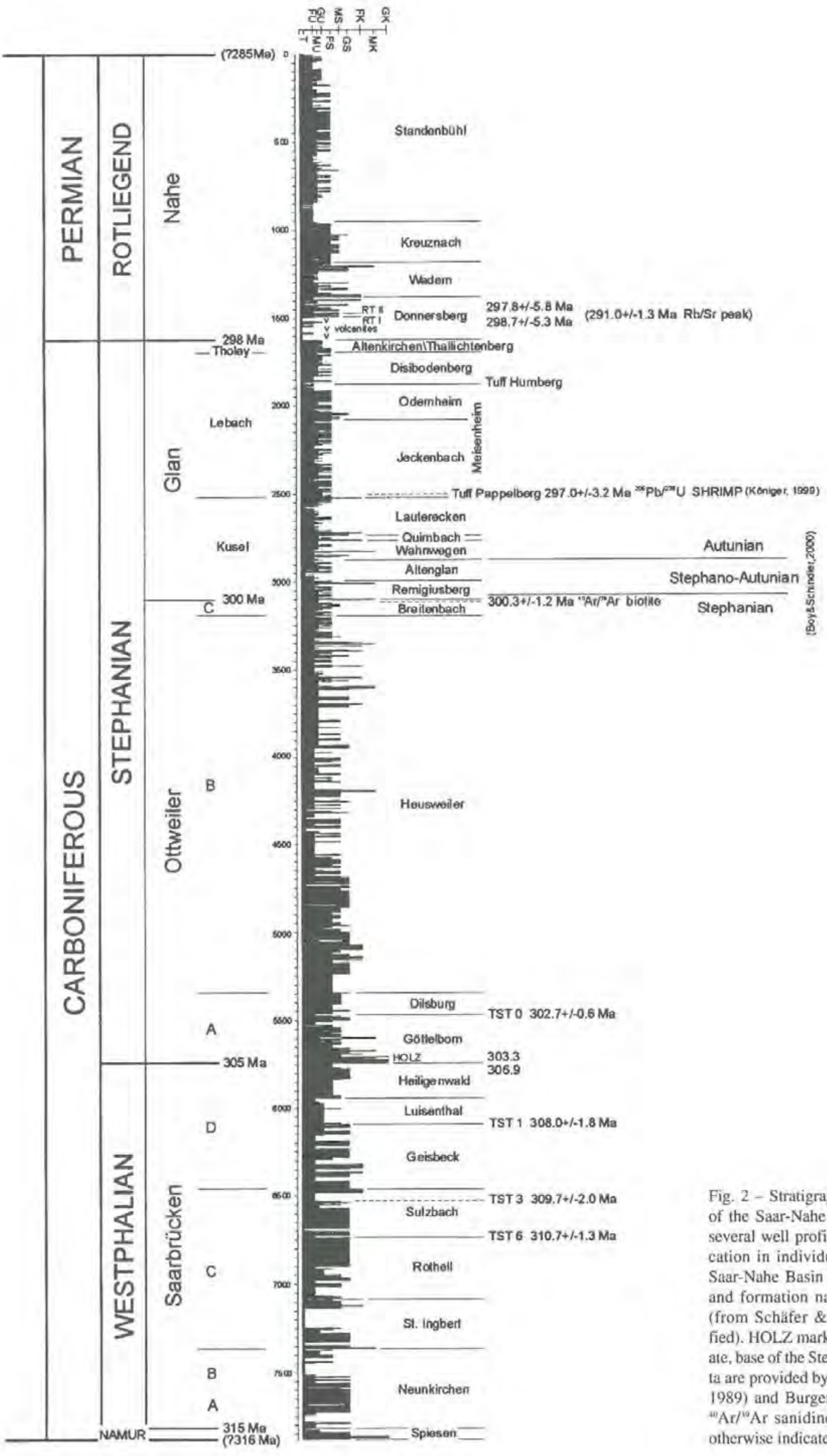

Fig. 2 – Stratigraphical standard section of the Saar-Nahe Basin, compiled from several well profiles, aligned to their location in individual depocentres of the Saar-Nahe Basin with group, subgroup, and formation names, currently in use (from Schäfer & Korsch, 1998; modified). HOLZ marks the Holz Conglomerate, base of the Stephanian. Numerical data are provided by Lippolt & Hess (1983, 1989) and Burger *et al.* (1997) and are $^{40}Ar/^{39}Ar$ sanidine plateau ages, unless otherwise indicated.

ly from orogenic pulses braided-fluvial conglomerate horizons invaded the basin in a stream-channel sedimentation style, providing well-defined lithostratigraphical markers. The Stephano-Autunian sediments were delivered from the Moldanubian in the S and the SW. Thus, they are rich in feldspars and granitic fragments. Also, they contain reworked durable siliciclastics from the Westphalian below. The basin dewatered towards the NE.

Intense Permo-Carboniferous volcanism produced lava flows, rhyolitic tuffs and intrusive rocks (Schwab, 1981, 1987); the volcanic rocks achieved a thickness of about 500 m. The rhyolites were mostly intrusives (Arikas, 1986; Hofmeister & von Platen, 1988; von Seckendorf & Chakraborty, 1993), but also formed volcanic tuffs showing a wide distribution and interacting with fluvial sedimentation (Stollhofen & Stanistreet, 1994). The volcanites were of the same chemistry as known from many other Permo-Carboniferous European continental basins (Plein, 1990; Plein *et al.*, 1995; Breitkreuz & Kennedy, 1999; Schaltegger, 1997). This volcanism was close to the base of the Permian Rotliegend.

Alluvial sedimentation in the Rotliegend above the volcanites (Stollhofen, 1994) was due to a change in basin geometry and the change of climatic conditions to more arid ones as before. Sediment input was from the north by rapid alluvial fans, and in addition from the E and SE where the Saxothuringian and the Moldanubian were eroded (Schäfer & Korsch, 1998).

After a folding event by the end of the Rotliegend, the basin was uplifted in the Cretaceous and the Tertiary. As a consequence, the Meso-Cenozoic overburden and part of the Rotliegend were removed (about 4000 m of bed thickness). The Saar-Nahe Basin became exposed from below the Mesozoic over a third of its surface, as it is known today (Schäfer & Korsch, 1998).

STRATIGRAPHY OF THE BASIN-FILL

Rich coal measures were preserved in the Westphalian and to some extent in the Stephanian as they were formed from the production and decay of tropical floras (Fig. 2). A structural disconformity with considerable loss of strata is obvious between the Westphalian and the Stephanian. The time gap comprises about 2 myr, and it is assumed that the missing Cantabrian stage is located here (Wagner & Winkler Prins, 1994, 1997; Korsch & Schäfer, 1995). The base of the Stephanian is marked by the Holz Conglomerate (HOLZ; Schäfer, 1986; Schäfer & Korsch, 1998).

It has been traditional to determine precise stratigraphic ages in the Saar-Nahe Basin by flora (Germer & Engel, 1989; Kerp, 1996), pollen and spores (Müller & Hoppe, 1996), and also by footprints of sauropods (Boy & Fichter, 1988; Boy & Martens, 1991). From this, the Carboniferous/Permian stratigraphical boundary was posted close below the Dirmingen Conglomerate at the base of the Glan Group. Right below this conglomerate in lacustrine beds of Stepanian C in age, a tuff was preserved, numerically dated to 300.3 +/– 1.2 Ma with a $^{40}Ar/^{39}Ar$ sanidine plateau age (Burger *et al.*, 1997).

The rhyolitic tuffs associated with the Permo-Carboniferous revealed $^{40}Ar/^{39}Ar$ sanidine plateau ages of 297.8 +/– 5.8 and 298.7 +/– 5.3 Ma (Lippolt & Hess, 1989). From initiation by Burger *et al.* (1997), Schäfer & Korsch (1998) assigned the stratigraphic boundary between the Carboniferous and the Permian to the Permian volcanites. Also, this age of roughly 298 Ma is about the same as was shown by Claoué-Long *et al.* (1995) and Roberts *et al.* (1995) using SHRIMP zircon $^{206}Pb/^{238}U$ ages, determining the Carboniferous/Permian boundary in Australia. This indeed contradicts the international trend to locate the Carboniferous/Permian boundary at 291 – 292 Ma (Wardlaw, 2000). On the other hand, Boy & Schindler (2000) discussed the palaeoecology of fishes and amphibians in lacustrine to fluvial environments close to the Carboniferous/Permian boundary in the Saar-Nahe Basin. On top of the Stephanian C (Breitenbach beds), they inserted a Stephanian D (Remigiusberg to Altenglan beds), a Stephanian D / Autunian transition in the Wahnwegen beds, and the Autunian from the Quirnbach beds onwards. Also, they found reason to correlate the Stephanian D / Autunian boundary with an absolute age of 296 Ma age (probably even 1 or 2 myr older).

DISCUSSION

The continental Saar-Nahe Basin developed as an internal molasse basin of the Permo-Carboniferous. Its subsidence was due to a dextral strike-slip regime that was active during the orogenic shortening of the Variscides, forming an asymmetric half-graben (Korsch & Schäfer, 1991). The calculation of its subsidence strongly depends on stratigraphic markers reliable for dating numerical ages. These markers for most of the Permo-Carboniferous are volcaniclastics, wind-born volcanic tuffs, produced external to the basin and assumed to be provided from the area of the northern Schwarzwald and/or the northern Vosges (Stollhofen, 1994). Some of them are numerically dated (Burger *et al.*, 1997; Königer, 1999). In addition to these, basin-derived rhyolitic tuffs originated from the Donnersberg (in the SE of the exposed part of the basin) and are associated with its intrusion process (Stollhofen, 1994). These tuffs provided the above cited numerical datum of about 298 Ma, as used by Schäfer & Korsch (1998). Yet, using this age, problems are obvious. The numerical age of the tuff in the Breitenbach Formation (Stephanian C) is about 300 Ma (Burger *et al.* 1997). This demands a time span of 2 m.y. for the strata of the Glan Group having a thickness of 2 km (Menning, 1995 a, b; 1999 pers. comm.). In any case, a rather high subsidence rate has to be calculated (Fig. 3; Schäfer & Korsch, 1998).

It should be considered that the Saar-Nahe Basin fill is extremely unbalanced during this period. Consisting of either a large amount of lake beds and meandering rivers versus coarse-grained conglomerates from stream-channels during the Kusel and Lebach Subgroups (Schäfer 1986; Boy & Schindler, 2000), subsidence must have been considerable in the pre-volcanic syn-rift phase (*sensu* Stollhofen, 1994). During the Tholey Subgroup, the basin regained a river-dominated environment with low-sinuosity meanders (Schäfer, 1986). Together with this, the fluvial input changed its direction from the SW to originate now from the Vosges in the south. No longer did it follow the basin axis from the Massif Central in the SW. A structural reorganisation is obvious, which may have been related to the volcanic syn-rift phase (*sensu* Stollhofen, 1994).

Permo-Carboniferous volcanic rocks cover large parts of Central Europe providing a volcanic signal rather short in duration. The rhyodacitic lavas in the North German Rotliegend Basin achieved a thickness of about 2500 m (Plein *et al.*, 1995; McCann, 1999). From these rhyo-

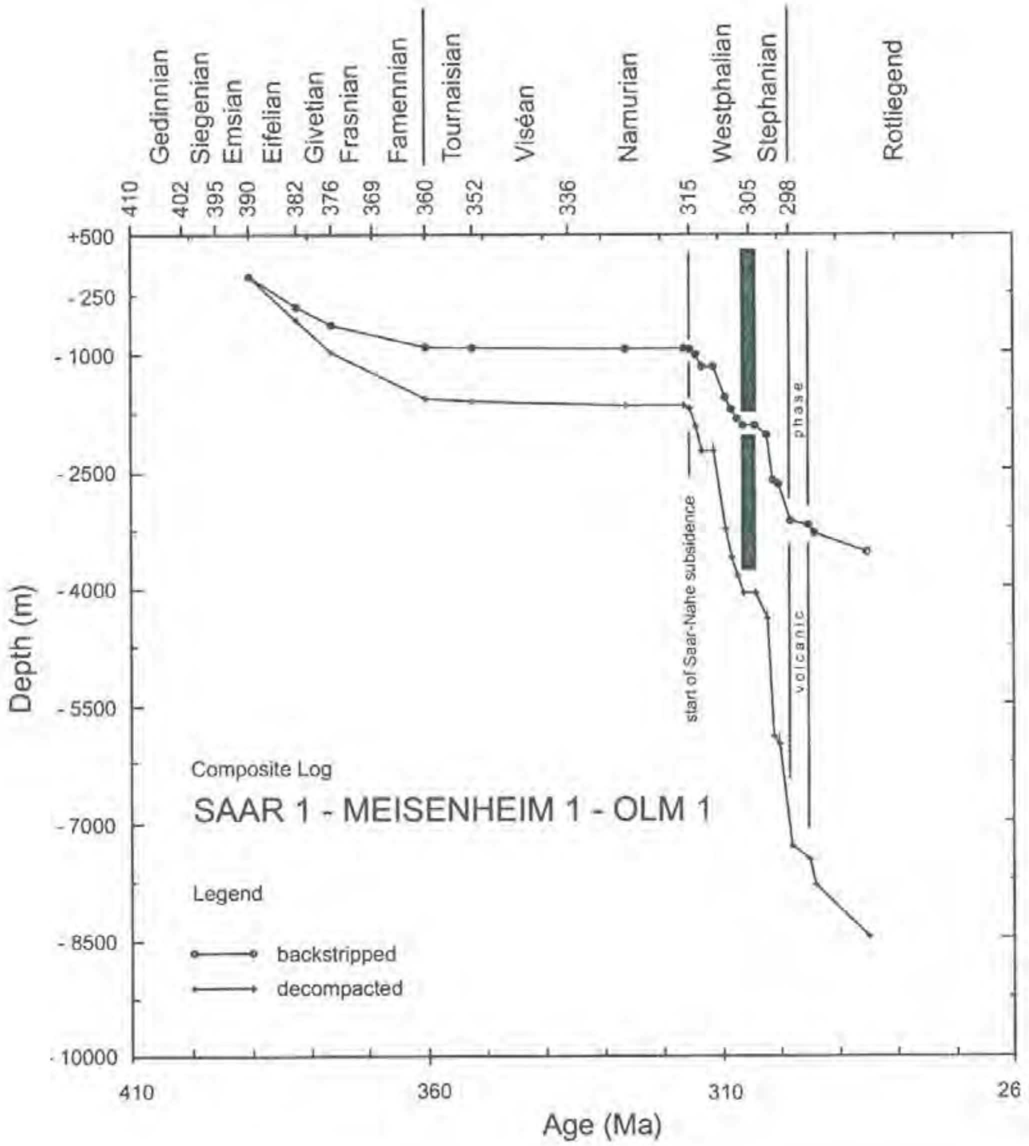

Fig. 3 – The Saar-Nahe Basin subsidence plot is calculated using the backstripping technique (Schäfer & Korsch, 1998). The curves are calculated from stratigraphic data from the wells Saar 1 (Emsian to Stephanian), Meisenheim 1 (Stephanian), and Olm 1 (Stephanian to Rotliegend). The lower curve shows the today's basin-fill, the upper one the structural subsidence of the basin (more details in Korsch & Schäfer, 1995). The horizontal segment at 305 Ma is the discordance due to deformation and erosion at the Westphalian/Stephanian boundary and is emphasised by the dark vertical bar.

dacitic lavas, Breitkreuz & Kennedy (1999) provided numerical ages, ranging from 300 +/– 3 Ma to 297 +/– 3 Ma (SHRIMP zircon ages) each from the centre and the top of the flows respectively, the sample sites being roughly 1.4 km apart from eachother (in well Mirow 1/74). In the North German Rotliegend Basin, from use of the oil industry, the volcanites mark the boundary between the Stephanian and the Rotliegend. In the Thüringerwald Basin, the Carboniferous/Permian boundary is located in a stratigraphical gap between the Möhrenbach and Ilmenau Group, whereby the effusive rhyolites in the Möhrenbach Group have a 296 +/– 5 Ma biotite age which is related to the intrusion of the Ruhla granite (Andreas & Wunderlich, 1998). Boy & Schindler (2000) focussed on a correlation between the Saar-Nahe Basin and the Thüringerwald Basin with palaeoecological findings, and they suggested the use Stephano-Autunian resp. Stephanian D to overcome the inconsistency of biostratigraphical and numerical data. As a consequence, their Stephanian D is different from that used by Schäfer & Korsch (1998) after the suggestion of Doubinger (1956).

Volcanic rocks in the continental Permo-Carboniferous of Central Europe provide stratigraphical markers. Yet, chronostratigraphical ages do not replace biostratigraphical zones. Nevertheless, they provide a chance to correlate between sedimentary basins, distant from each other, and also internal and external to the Variscides.

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VARIATION AND PRESERVATION OF *ICHNIOTHERIUM* IN THE TAMBACH SANDSTONE (ROTLIEGEND, THURINGIA)

SEBASTIAN VOIGT [1]

Keywords – tetrapod footprints; *Ichniotherium*; Rotliegend; Thuringia; trackway preservation; ichnotaxonomy.

Abstract – This contribution focuses on the variation of the main trackway characters of *Ichniotherium cottae* Pohlig, 1889 from the Tambach Sandstone. The investigation is based on the detailed description of 22 trackways with nearly 330 tracks of manus and pes, altogether.
As a result, it can be shown that there is strict and direct control of the trackway characters by substrate and gait. Knowledge of the inter-relationships of imprint morphology, trackway pattern, and its controlling factors is of great importance for ichnotaxonomy, and gives new insight into the ecology and environment of the trackmakers.

Parole chiave – impronte di tetrapodi; *Ichniotherium*; Rotliegende; Turingia; conservazione delle piste, icnotassonomia.

Riassunto – Questo contributo s'incentra sulla variazione dei principali caratteri delle piste di *Ichniotherium cottae* provenienti dall'Arenaria di Tambach. La ricerca è basata sulla dettagliata descrizione di 22 piste con quasi 330 impronte di *manus* e *pes*, complessivamente. Ne consegue che si può porre in evidenza che vi è uno stretto e diretto controllo dei caratteri delle piste a seconda della composizione del substrato e dell'andatura degli animali. La conoscenza dei rapporti tra morfologia delle impronte, disposizione delle piste, e loro fattori di controllo è di grande importanza per l'icnotassonomia, e offre nuove prospettive di studio sull'ecologia e sull'ambiente dei *trackmakers*.

INTRODUCTION

For more than 110 years the Tambach Sandstone has been well known as a source of excellently preserved tetrapod tracks. Pabst (1908) gave the first extensive description of tetrapod trackways from the Tambach Sandstone. All trackway slabs come from sandstone quarries in the Bromacker locality, 1.5 km north of Tambach-Dietharz village in the Thuringian Forest of central Germany (Fig. 1).

The Tambach Sandstone represents the middle unit of the Tambach Formation, which is stratigraphically positioned at the base of the Upper Rotliegend, Lower Permian. In addition to the tracks, the recently discovered tetrapod fauna at the Bromacker locality indicates a biostratigraphical position near the base of the Lower Permian, comparable to the Wolfcampian Series of North America (Sumida *et al.*, 1996).

The Bromacker tetrapod tracks are preserved as casts of original imprints at the base of bedded, fine- to medium-grained, reddish-brown sandstones. The sandstones are underlain by thin mudstones, which represent the original surface of trackway formation. The sandstone-mudstone interbedding of the trackway horizon is derived from repeated sheet floods followed by standing-water sedimentation in an intramontane undrained basin (Eberth *et al.*, 2000). The facies of the locality indicates a position near the basin centre (Fig. 1).

ANALYSIS AND MATERIAL

More than 200 trackway slabs have been discovered at the Bromacker locality so far. The biggest slab is 8 metres long. In spite of the abundance of trackway slabs, only five ichnospecies could be differentiated (Haubold, 1998). Approximately 70% of all trackways from the Tambach Sandstone are pentadactyl imprints of the ichnospecies *Ichniotherium cottae* Pohlig, 1885 (Fig. 2). This abundance of material and the obviously varying states of preservation were of prime importance to the analysis of the variability of the ichnospecies. The investigation focused on analysis of the existing interrelationships of the main trackway characters, and the features of the substrate upon which the trackmakers walked. Knowledge of the interrelationships between imprint morphology, trackway pattern and the substrate characteristics is essential in

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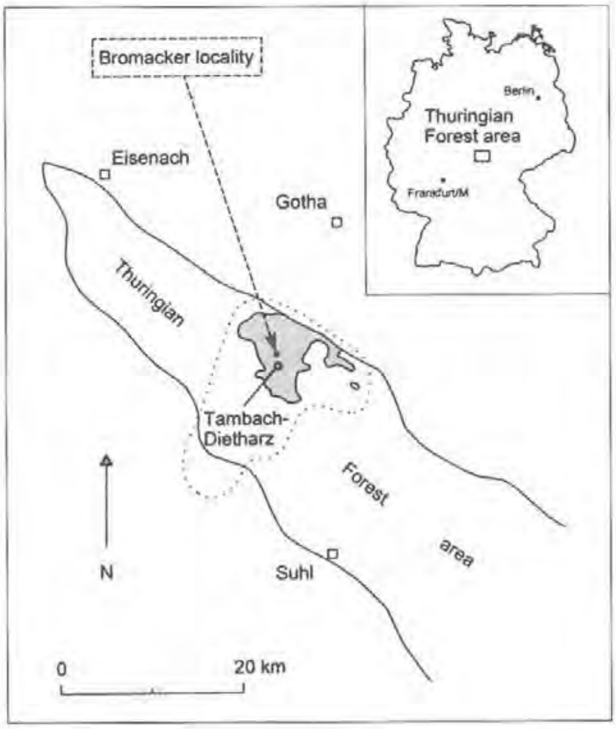

Fig. 1 – The outline of the Thuringian Forest area in central Germany and the Bromacker locality. The extent of the Tambach Formation is shown by the shaded area. The dotted line represents the hypothetical limit of the Tambach Basin (from Haubold, 1985).

Fig. 2 – I. cottae (MNG-1351): The first-discovered trackway slab from the Tambach Sandstone. The trackway is about 1.5 m long and one of the most representative examples of the ichnospecies from the locality.

avoiding the creation of new, unnecessary ichnotaxa. As shown below, there are new indications of the ecological situation of the trackmaker from knowledge of its locomotory behaviour. All results of this work are based on the investigation of 22 longer trackways of *I. cottae* with about 330 single imprints in total. About 2500 measurements have been recorded for a broad-based quantitative analysis (Voigt, 1999).

The vast majority of the investigated trackway material is stored at the Museum der Natur Gotha (MNG). Some other important trackway slabs are stored at the Museum für Naturkunde Berlin (MB).

IMPRINT MORPHOLOGY

Imprint morphology represents the most important characters of a trackway. *I. cottae* from the Tambach Sandstone shows two fundamental differences in imprint morphology. The first relates to the depth and completeness of the imprints. The other relates to certain dimensions of the imprint, especially the relative position and length of the digit V.

The variation in depth and completeness of the imprints is very broad. There are all stages of transition between deep, plantigrade imprints and shallow, unguligrade-like imprints (Fig. 3). The depth and completeness of the imprints are firstly a function of the porewater content of the substrate. This relationship is very close and could be identified on a 2.3 m long trackway which was discovered in summer 1995. The extent and orientation of mudcracks and raindrop impressions on the surface of this trackway slab allows the reconstruction of small-scale palaeotopography in the area of the slab. The extent and orientation of the sedimentary surface structures indicate that the trackmaker went from firm ground on to the soft sediment within a former pool or channel. Depth, completeness and the sharpness of the imprints change systematically in the direction of locomotion. Thanks to the preservation of this special trackway, it is possible to judge the relative consistency of the substrate of all investigated tracks at the time of trackway formation. As a result, the variation of depth, completeness and sharpness of the imprints is first of all controlled by the parameters of the substrate.

As mentioned above, there is variation in the proportions of imprints, too. Firstly, the pes imprints show different types of preservation with regard to the relative position and length of the digit V (Fig. 4). The remarkable lateral shift of the digit V as seen in the photograph on the right (Fig. 4) could be observed only on five of all 22 investigated trackways. An explanation of this kind of variation in imprint morphology is provided by the locomotory behaviour of the trackmakers. Figure 5 shows the locomotory efficiency of all 22 investigated trackways in order from low

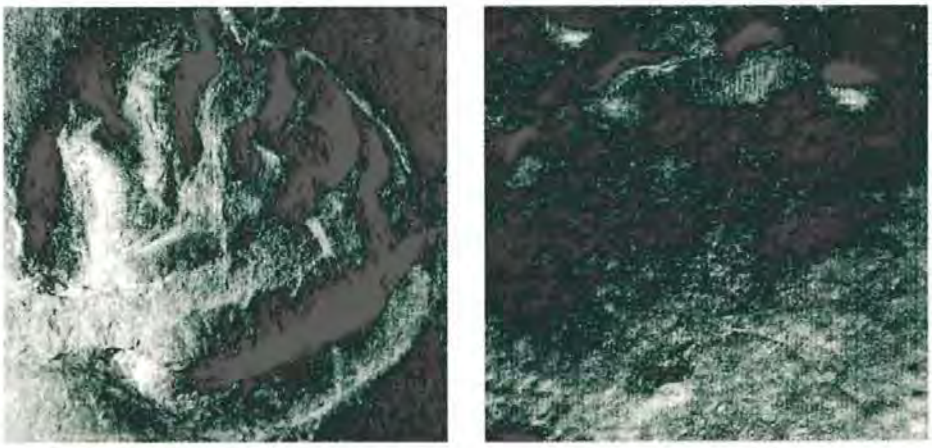

Fig. 3 – Left pes imprints of *I. cottae* to demonstrate the wide differences of preservation in the depth and completeness of the imprints. Illumination on both photographs is from the left. Scale: 5 cm. Left: MNG-10072, right: MNG-1840.

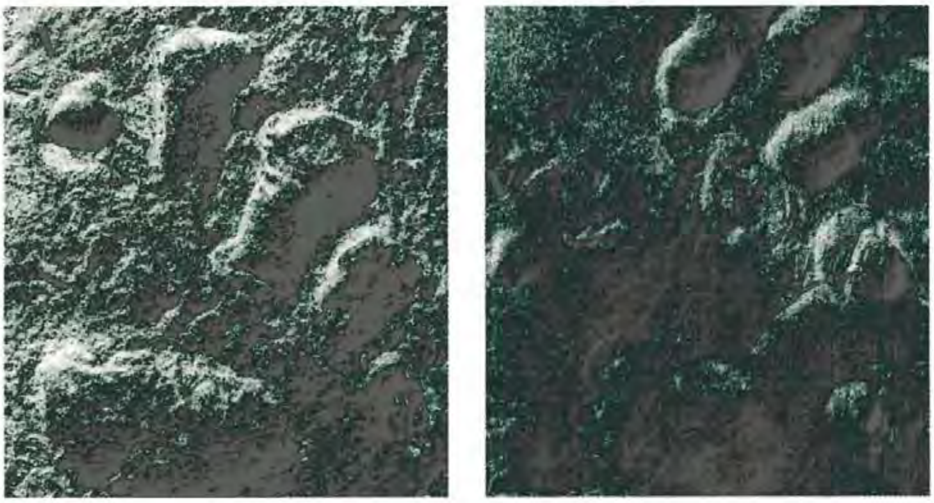

Fig. 4 – Left pes imprints of *I. cottae* to demonstrate the variability of imprint morphology with regard to the relative position and length of the digit V. The Roman numeral 'V' refers to the digit V. Illumination on both photographs is from the left. Scale: 5 cm. Left: MNG-1351, right: MNG-1352.

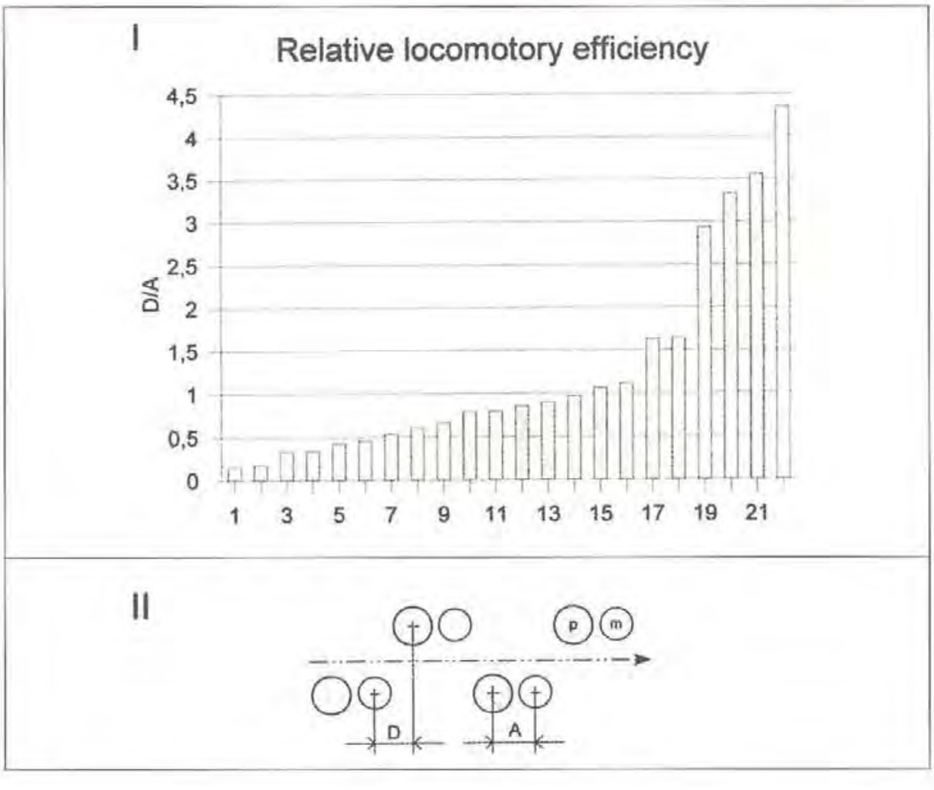

Fig. 5 – I. Graphical representation of the relative locomotory efficiency of all investigated trackways. The relative efficiency of the trackmaker's gait can be calculated from the trackway pattern. The relative locomotory efficiency is given by the ratio D/A. II. The meaning of the measurements D and A is graphically expressed in the lower part of the figure. D refers to the distance between the manus and pes imprint of two successive sets. The measurement A refers to the distance between manus and pes imprint within one set. Abbreviations: p, imprint of the pes; m, imprint of the manus.

to high. The five trackways characterized by a very lateral backward shift of the digit V in the pes imprints all lie at the higher end of the spectrum (numbers 18-22, Fig. 5). Therefore, I speculate that the trackmakers of *I. cottae* were able to change their gait from a more plantigrade to a partially digitigrade gait in order to achieve higher speed and an increased efficiency of locomotion.

TRACKWAY PATTERN AND MEASUREMENT DATA

The second important trackway character, the trackway pattern, is also variable. In the case of *I. cottae*, two extremes in trackway pattern can be observed (Fig. 6). In the first (type 1, Fig. 6), the imprints of manus and pes on the left and right side of the trackway lie nearly opposite one another. In the second (type 2, Fig. 6), the sets of manus and pes of both sides are clearly arranged alternately. All stages of the transition, from an opposite to an alternating arrangement of the trackway pattern, are present. The shift in the trackway pattern coincides with a shift in the trackway measurement data. The trackways of type 1 show a low pace angulation, a large distance between manus and pes, and a low stride-body length ratio. In addition, the type 1 trackways are more irregular. In contrast, the trackways with a clearly alternate pattern similar to type 2 are characterized by a high pace angulation, a small distance between manus and pes, and a high stride-body length ratio. Trackways of type 2 exhibit a more regular trackway pattern.

Moreover, there is a specific relationship between the parameters of the substrate and the trackway pattern type. Trackways with deep, plantigrade, completely preserved imprints show only the type 1 trackway pattern. In contrast, the trackway patterns of the second type could be observed only in trackways with shallow, digitigrade, commonly incomplete imprints. Thus, the gait of the trackmakers was limited by the characteristics of the substrate. The trackmaker achieved a very low speed with a careful and shaky gait on soft sediments, and a high-speed, more agile gait on firmer ground. It can be concluded that the trackmakers of *I. cottae* avoided walking on wet substrates. Most probably, the trackmakers normally lived on firmer substrates.

CONCLUSIONS

The variation analysis of *I. cottae* allows the differentiation between an input of substrate and an input of gait on the main trackway characters. As a result, the general relationship between imprint morphology, trackway pattern, and certain parameters of the substrate can be identified. All observed types of different preservation of *I. cottae* can be explained by the influence of extramorphological parameters, namely the substrate and gait.

Therefore, it is evident that the various preservation types were formed by animals with identical anatomical structures of the hand and foot. Consequently, the tracks of these animals should be assigned only one ichnotaxonomic name.

Moreover, the analysis yields indications to the locomotory behaviour and ecological situation of the trackmakers. The results of the investigation indicate that the trackmakers were fully terrestrial animals that walked preferentially upon firmer, drying, mudcracking surfaces. Consequently, the usual habitat of the trackmakers was not represented by the muddy, wet substrates of the central Tambach Basin plain. The trackmakers of *I. cottae* probably lived in flat upland areas around and away from the small pools and channels of the basin plain, and did not invade the basin plain until an advanced stage of dessication.

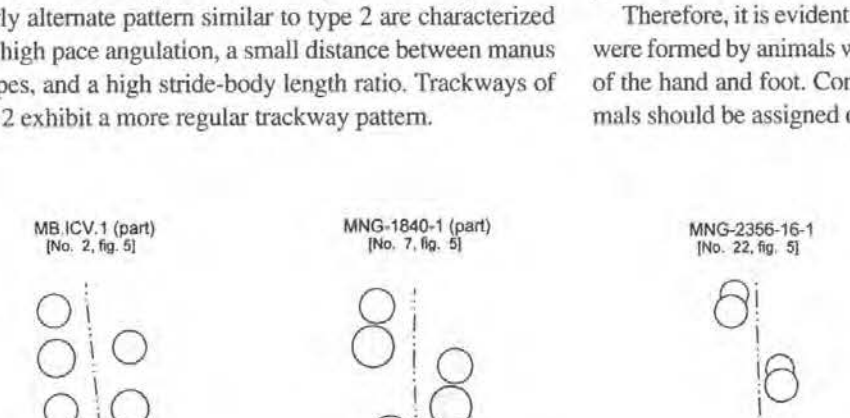

Fig. 6 – The range of the trackway pattern of I. cottae in the Tambach Sandstone represented by three characteristic trackways for each stage. The locomotory efficiency, the speed of gait and the regularity of the trackway pattern increase from left to right. For abbreviations see Fig. 5.

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PERMIAN EVOLUTION OF THE WESTERN CARPATHIANS, BASED ON THE ANALYSIS OF SEDIMENTARY SEQUENCES

ANNA VOZÁROVÁ[1]

Key words – Permian sedimentary evolution; crystalline basement; collision zones.

Abstract – The Permian sequences of the Western Carpathians are represented by continental, mainly coarse-grained volcanosedimentary formations, the origins of which were ascribed to collision-related transpressional/extensional and extensional sedimentary basins. They originated in time and space as a consequence of the collisional events of the Hercynian orogeny. The result was unevenly consolidated continental crust, which, after a short period of stability, was incorporated again into the Alpine orogenic cycle. The three crustal type fragments of crystalline basement in the Alpine Western Carpathian units were distinguished as: the Central Western Carpathian Crystalline Zone (CWCZ), the Northern Gemeric Zone (NGZ) and the Inner Western Carpathian Crystalline Zone (IWCZ). The main differences between the zones are in the chronological and spatial development of the Late Paleozoic sedimentary basins.
The relicts of two Hercynian zones of distinctly different geodynamic aspects are preserved in the Alpine structure of the Western Carpathians: the internal zone, with termination of collision during the Bretonian-Sudetian events; and the external, with termination of collision during the Asturian phase.

Parole chiave – evoluzione sedimentaria permiana; basamento cristallino; zone collisionali.

Riassunto – Le successioni permiane dei Carpazi occidentali sono essenzialmente rappresentate da formazioni vulcaniche e sedimentarie continentali, la cui origine fu riferita a bacini transpressivi/estensionali connessi a collisione ed a bacini estensionali. Esse si generarono nel tempo e nello spazio a seguito degli eventi collisionali dell'orogenesi ercinica. Il risultato di questi eventi fu la formazione di una crosta continentale non uniformemente consolidata, che, dopo un breve periodo di stabilità, fu di nuovo inclusa nel ciclo orogenetico alpino. Tre frammenti di tipo crostale di basamento cristallino nelle unità dei Carpazi occidentali alpini furono contraddistinti: la Zona Cristallina dei Carpazi centro-occidentali (CWCZ), la Zona Gemmerica settentrionale (NGZ) e la Zona Cristallina dei Carpazi occidentali interni (IWCZ). La principale differenza tra queste zone sta nello sviluppo cronologico e spaziale dei bacini sedimentari tardo-paleozoici.
I resti di due zone erciniche con aspetti geodinamici distintamente diversi sono preservati nella struttura alpina dei Carpazi occidentali: la zona interna, in cui la collisione si concluse durante gli eventi tettonici Bretone-Sudetico, e la zona esterna, in cui la collisione si estinse in concomitanza della fase Asturica.

INTRODUCTION

The kinematic evolution of the Western Carpathian orogenic system occurred during both Variscan and Alpine times. Fragments of newly-formed Epi-Variscan crust were incorporated in the Paleo-Alpine Western Carpathian units as evidenced by repeating subduction/collision and transform fault processes. Like most other collisional belts, the Western Carpathians have been divided into external and internal structural zones. The main difference between the traditionally distinguished zones is in the age of the main Alpine events as well as in the intensity of their deformational and metamorphic effects: (1) the internal zone – the HP/LT Late Jurassic subduction event and Lower/Middle Cretaceous collision, followed by nappe stacking; (2) the external zone – from the Upper Cretaceous/Lower Paleocene to the Oligocene/Lower Miocene subduction/accretion and collisional events. The fragments of the Late Paleozoic sedimentary basin fill are preserved only within the internal zone, as part of the principal Alpine crustal-scale superunits (Fig.1). Relicts of the Upper Carboniferous sedimentary sequences are proven by lithofacial and biofacial data in a relatively wide range of sedimentary environments, from shallow-water to paralic and continental. The Lower Permian sediments confirm only a continental environment as a whole. The Upper Permian facial associations confirm more variability, from continental to sabkha-lagoonal evaporitic formations, as a consequence of the start of a new sedimentary cycle, better developed during the Mesozoic.

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CHARACTERISTICS OF THE BASEMENT

In the Western Carpathians different types of Variscan basement were overstepped by the Upper Carboniferous/Permian sedimentary sequences. Regardless of the age of the metamorphic overprint, the Alpine-Western Carpathian basement can be subdivided into three zones:
1. *The Central Western Carpathian (CWCZ) crystalline zone* composed mainly of metamorphic rocks and huge masses of pre-Mesozoic granitoids. Fragments of pre-Hercynian metamorphic crust are probably also included in this zone. Several pre-Alpine terranes were identified in the CWC-Alpine nappe units (*the Tatra Plate*: the relicts are preserved within the Alpine Tatric, Northern Veporic, Southern Veporic and Zemplinic Units; these substratum fragments were formerly described as the Tatra Terrane, Kohút Terrane and Suspected Ipoltica Terrane – within the Hronic Unit; Vozárová & Vozár, 1996). Nearly all underwent the Hercynian metamorphism during the Early Carboniferous, postdating locally Silurian/Devonian and/or older metamorphic events. The magmatic activity took place in two stages as indicated by Rb-Sr and U-Pb dating, which gave 360-340 Ma and 320-300 Ma; (Cambel & Král, 1989; Cambel *et al.*, 1990). The age of rare post-orogenic A-magmatites corresponds to the Permian (Rb/Sr: 280-250 Ma; Cambel *et al.*, 1989). Geochemical data indicate the operation of subduction system and then collisional tectonics.
2. *The Northern Gemeric Zone (NGZ)*, with relicts of the Visean-Namurian flysch and thrust wedges of pre-Carboniferous oceanic crust. This zone represents a relict of a Hercynian collision suture. Two undated Hercynian terranes, which differ in tectono-metamorphic development (amphibolite vs. greenschist facies) and probably also in the age of protolith, belong to this domain. Their gradual amalgamation took place during the Early Carboniferous because the Uppermost Visean/Serpukhovian, shallow-water carbonate-clastic development was, after a stratigraphic hiatus, unconformably overstepped by the Moscovian marine "molasse". This marine "molasse" unconformably overlies part of the Lower Carboniferous flysch sequence (eastern part of the NGZ), to fix up the Late Variscan thrust sheet structure.
The higher-grade crystalline complex (the Klátov Terrane) consists mainly of amphibolites, subordinate gneisses and metaultramafic rocks. This complex is considered as a partly incomplete ophiolite suite (Hovorka *et al.*, 1984; Spisiak *et al.*, 1985). Findings of regressively overprinted eclogitic rocks lead to the assumption of a polyphase metamorphic P-T path, with high-pressure conditions dur-

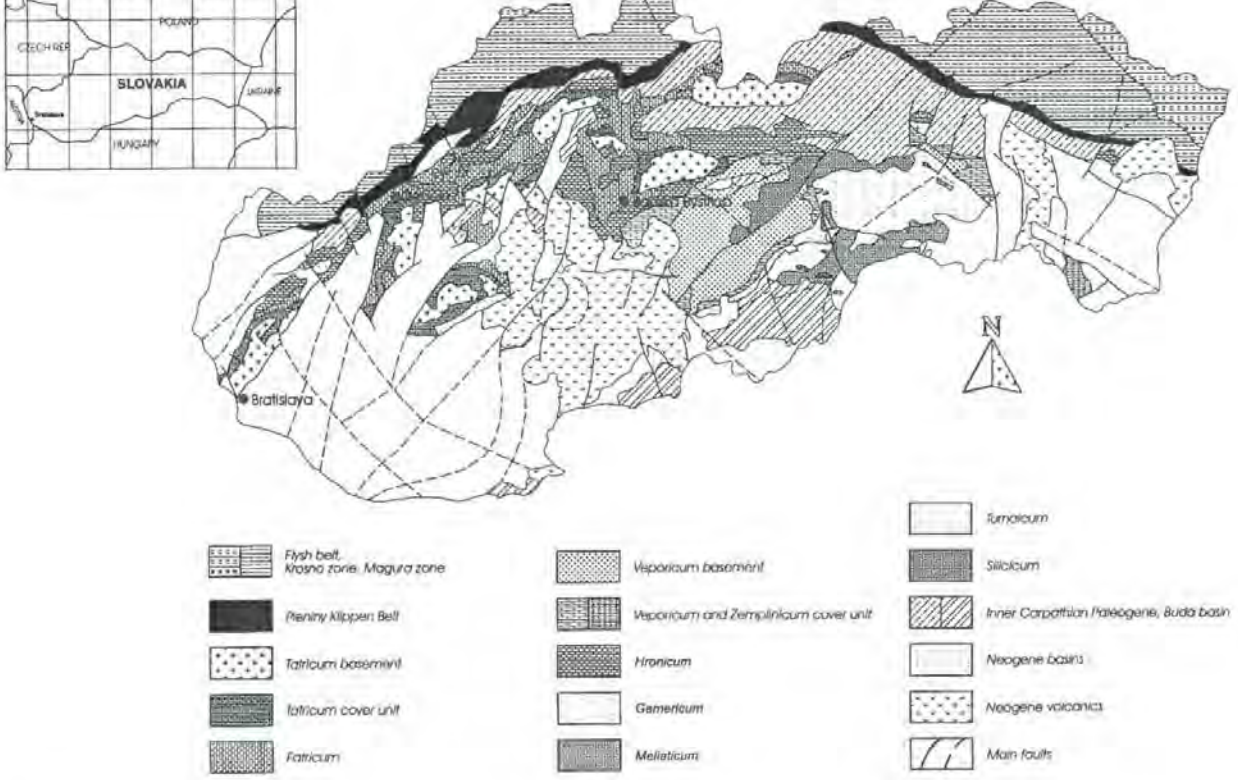

Fig. 1 – Tectonic sketch of the Slovak part of the Western Carpathians.

ing the climax of metamorphism. The low-grade crystalline complex (the Rakovec Terrane) is composed of tholeiitic metabasalts and matavolcaniclastics, associated with smaller amounts of sandy-shaly metasediments and small bodies of gabbro-diorites and metakeratophyres. The magmatic rocks show geochemical characteristics near to E-MORB/OIT basalts (Ivan, 1994).

3. *The Inner Western Carpathian crystalline zone (IWCZ)*, which is subdivided into two subzones – *the Southern Gemeric Subzone (SGS)* and *the Szendrö – Bükk Subzone (SzBS)*. The dominant part of the Southern Gemeric Subzone (corresponding to the *Gelnica Plate*) is composed of the Lower Paleozoic volcanigenic flysch formation and of its Permian-Triassic cover. An enormous mass of volcanogenic flysch comprises distinct turbidity current and other mass-gravity flow sedimentation features. Besides redeposited acid to intermediate volcaniclastic material, derived from the synsedimentary continental magmatic arc, detritus from the subduction complex and fragments of oceanic crust are also present. Regional metamorphism of the SGS basement did not exceed the low-pressure greenschist facies. The SGS Lower Paleozoic flysch sequence was interpreted as a relict of the fore-arc basin related to an active continental margin (Vozárová, 1993).

Within the Szendrö-Bükk Subzone the pre-Carboniferous complexes are preserved only locally. They are represented mainly by the shallow-water to basinal sequences of the passive continental margin of the northern Gondwana promontory (for a precise description see Kovács & Péro, 1983; Fülöp, 1994). The most typical features of both these subzones are the Bashkirian olistostroma flysch complexes. They were reported from the Turnaic Unit in the Inner Western Carpathians (Vozárová & Vozár, 1992) as well as from the Szendrö and Bükk Mts. (Kovács, 1988, 1992). Typically, the Variscan metamorphism of the SzBS Paleozoic complexes is either slight or absent.

LATE PALEOZOIC OF THE CENTRAL WESTERN CARPATHIAN CRYSTALLINE ZONE

The Permian deposits of the CWCZ either gradually develop from the underlying Upper Westphalian-Stephanian formations or unconformably overlie the crystalline basement rocks (Fig. 2). The Lower Triassic deposits disconformably rest on both the above structural stages. The Upper Carboniferous-Permian as well as the Permian sedimentary basins were founded on continental crust. These basins probably formed cratonwards of thin-skinned thrusting, on both the footwalls and hanging walls of crustal-scale reverse faults that ruptured the continental basement. The typical result of deformation was a set of non-marine basins connected by basement uplifts and thin-skinned thrust belts (broken-foreland basins). These retro-arc foreland basins occurred behind the compressional belt, on the thickened continental crust.

The Upper Westphalian-Stephanian sedimentary basins are represented by huge wedges of siliciclastic sediments deposited in continental, fluvial or fluvio-lacustrine and swamp sedimentary environments. These formations are associated with acid to intermediate calcalkaline volcanism. The Upper Westphalian and Stephanian ages are proved by macroflora: *Calamites cistii* Brongn., *Asterophyllites trichomatosus*, *Annularia pseudostelata* Potonie, *Pecopteris cyathea* Schotheim, *Cordaites borassifolius* Sternberg, *Asterotheca miltonii* Artis, *Asterotheca arborescens* Brongn., *Cordaites palmaeformis* Goepp. and *Callipteridium gigas* Guttb. (Nemejc, 1947; Sitár in Planderová *et al.*, 1981; Sitár & Vozár, 1973). The sedimentary regime continued uninterrupted in this basin during Permian times. Distinct changes in climatic conditions took place when the warm and humid climate was replaced by dry and arid conditions.

Generally, the Permian sequences are represented by continental, mainly variegated coarse-grained alluvial fan formations of red-bed type. Integral parts of these formations are volcanic rocks and their volcaniclastics, dominated by calcalkaline rhyolites-dacites, with subordinate andesites and continental tholeiitic andesite-basalts. The Permian sediments indicate very low structural and mineralogical maturity, with provenances from uplifted and tectonically rejuvenated crystalline basement or from dissected magmatic arc (inactive in Permian times, produced by pre-Carboniferous subduction). The prevalent sediments were deposited in alluvial or fluvio-lacustrine and ephemeral lake environments, in semi-arid to arid climatic conditions as indicated by the typical sedimentary facies and absence of fauna and flora. The biostratigraphical data are based on relatively poor finds of pollen and sporomorphs. The Early Permian age is generally documented by the presence of *Potonieisporites* and *Vittatina*. The Upper Permian age is proved by the following microflora assemblage: *Calamospora nathorstii* Klaus, *Klausipollenites* div. sp., *Lueckisporites parvus* Klaus, *L. virkkiae* (Pot.) Klaus, *Monosulcites minimus* Cookson, *Striatites richteri* (Klaus) Jizba, *Limitisporites rectus* Leschik and *Jugasporites lueckuides* Klaus (Planderová, 1973; Planderová *et al.*, 1981).

LATE PALEOZOIC OF THE NORTHERN GEMERIC ZONE

The continental Permian sequences overlie the slightly deformed relicts of the Westphalian peripheral basin fill, as well as all pre-Westphalian crystalline complexes of the NGZ (Fig. 2). The mostly coarse-grained clastic sediments derived from the collisional belt are associated with bimodal andesite-basalt/rhyolite volcanism. The characteristic features are: (1) varicoloured clastic sediments of violet and violet-red; 2) a gradual fining-upwards; 3) cyclicity manifested within the framework of small cycles as well as megacycles; 4) bimodal calcalkaline volcanism. The thickness of the basal formation is extremely variable (from several tens of metres to 350 m). It is composed of pebble material indicating source rocks from the immediate basement. The immature coarse-grained sediments represent fossil mudflows, partly reworked in some places, continued by alluvial, mainly stream channel deposits. An "Autunian" age is proposed because of their position below the main volcanic horizon. The widespread upper part of the Lower Permian ("Saxonian") polyphase volcanic activity manifested the spatial and temporal relationships to the large sedimentary cycles. Sediments are characterised by a low degree of maturity and by a mixture of syngenetic volcanic and non-volcanic detritus. The fining-upwards alluvial cycles, with channel lag, pointbar and floodplain lake facies alternating with playa subenvironments at the topmost part of large cycles, are among the most striking features. The age is inferred from the isotopic analyses of sulphides from the volcanigenic horizons: $^{206}Pb/^{238}U$ = 263 Ma; $^{207}Pb/^{235}U$ = 274 Ma (Novotný & Rojkovič, 1981). The upper part of the Lower Permian ("Saxonian") was also proved by poor microflora.

The "Autunian-Saxonian" formations are overlain by a relatively mature sandy-conglomerate horizon, with some pebble material derived from the immediate stratigraphic underlier. This could indicate a break in the sedimentation after the "Saxonian", but there is no biostratigraphical evidence to support this assumption. Alluvial, stream-channel deposits prograde upwards to the inland sabkha and near-shore sabkha/lagoonal facies, with the anhydrite-

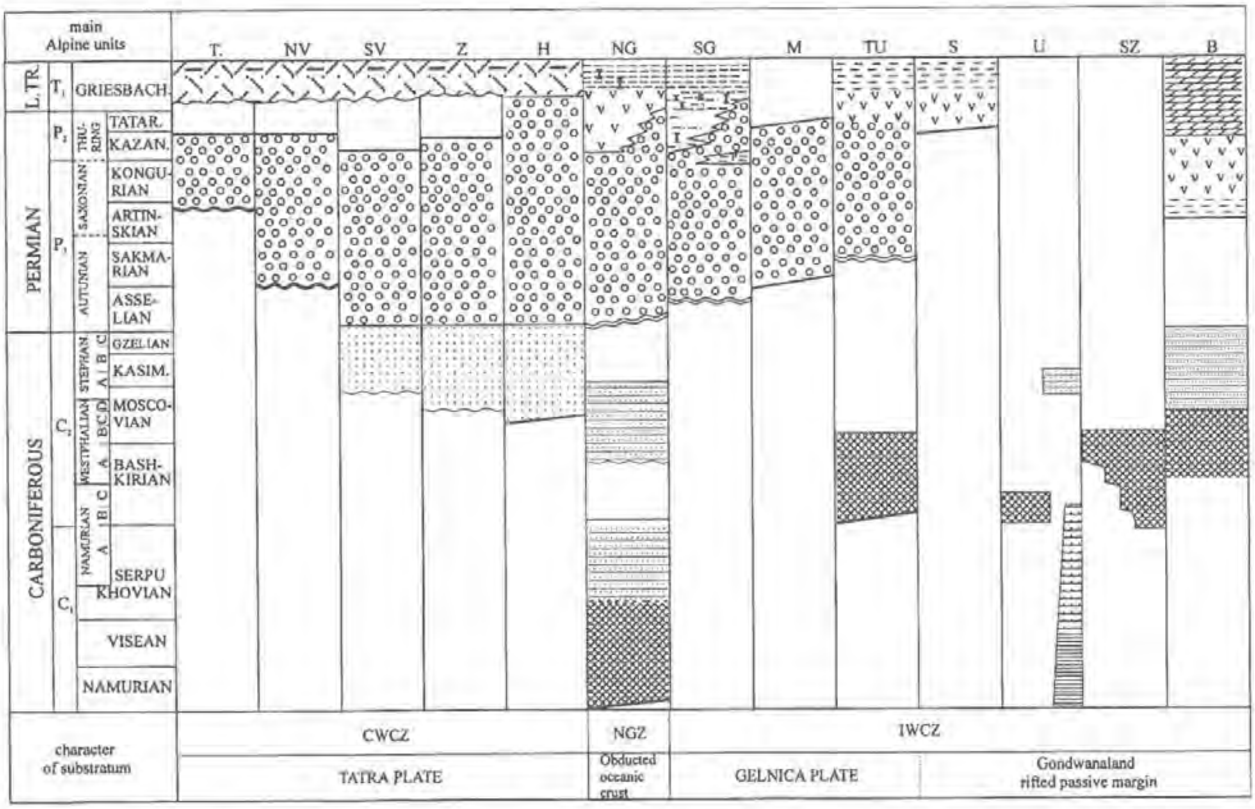

Fig. 2 – Scheme of sedimentary evolution of the Western Carpathians Carboniferous-Permian basins.

1 – continental, prevalently coarse-grained alluvial, alluvial-lacustrine sediments; 2 – continental lacustrine, fluvio-lacustrine sediments with thin coal seams in places; 3 – aeolian, braided alluvium to tidal clastic sediments of greater maturity; 4 – sabkha-lagoonal facies; 5a – lagoonal to shallow-water clastic sediments with thin intercalations of carbonates; 5b – shaley-sandy deposits with thin lenses of phosphorites; 6 – evaporites; 7 – deltaic and shallow-marine clastic deposits, with less carbonates; 8 – shaly-carbonate shelf sediments; 9 – basinal shaly-carbonate sequences; 10 – dolomites, limestones; 11 – flysch: turbidites, olistostrome sequences; disconformity: undulating line; unconformity: double undulating lines; overthrust planes: oblique line. Abbreviations of the main paleo-Alpine units: T – Tatric Unit; NV – Northern Veporic Unit; SV – Southern Veporic Unit; Z – Zemplinic Unit; H – Hronic Unit; NG – Northern Gemeric Unit; SG – Southern Gemeric Unit; M – Meliatic Unit; TU – Turnaic Unit; S – Silicic Unit; U - Uppony Mts.; SZ – Szendrő Mts.; B – Bükk Mts. Abbreviations of the substratum: CWCZ – Central Western Carpathian Crystalline Zone; NGZ – Northern Gemeric Zone; IWCZ – Inner Western Carpathian Crystalline Zone.

gypsum and salt breccia horizons. The isotopic analysis of sulphur (Vozárová, 1997) shows results similar to the data obtained from the Upper Permian to lowest Triassic. There are transitions up to the *Pseudomonotis (Claraia) clarai* (Emmr.) horizon (Griesbachian-?Nammalian).

The Permian sedimentary basin was part of the collision-related foreland basins at the suture zone. The development of this peripheral foreland basin was preceded by the consumption of oceanic crust and, during the first stage of oblique continental collision, by development of an Early Carboniferous remnant ocean basin. Later collisional shortening led to deformation and cannibalisation by uplift and erosion of the early accretionary wedges. The sedimentary infill of the marine Westphalian basin shows evidence of active thrust-faulting in the adjacent collision belt. This can be documented by "clastic wedges" of delta-fan conglomerates. Finally, the extensional faulting on the downgoing rifted continental margin resulted in widespread deposition of coarse-grained nonmarine clastics during the Permian. These clastic red-bed wedges document a large sediment supply from the collision fold-thrust belt. The continual collision during the Permian resulted in the development of a continental sedimentary basin with dominant strike-slip faulting.

LATE PALEOZOIC OF THE INNER WESTERN CARPATHIAN CRYSTALLINE ZONE

The Late Variscan, post-orogenic overstep sequences of the Southern Gemeric Subzone unconformably overlie the Lower Paleozoic crystalline basement of the IWCZ (Fig. 2). Generally, the basal parts of the Lower Permian volcanosedimentary complexes are characterised by a high content of mature mineral detritus. Conspicuous fining upwards is accompanied by a relative decrease in mineralogical maturity of sediments. The whole sequence is subdivided vertically into two large cycles, with the quartzose conglomerate horizons at the base of each and a sandstone-shale member between the two. The stream-channel and sheet-flood deposits with unimodal transport systems predominate. The upper conglomeratic horizon contains detritus from rhyolite-dacite synsedimentary volcanism.

The prograded Upper Permian horizon is a monotonous complex of cyclically alternating sandstones, siltstones and shales. Lenses of carbonatic sandstones and dolomitic limestones with intercalations of shales occur only in its upper part. Exceptionally, thin lenses of phosphatic sandstones and sediments with extremely high contents of albite (albitolites) occur. This sedimentary environment is interpreted as alluvial-lacustrine and lacustrine, with highly alkaline lakes in some places, prograding into the near-shore and lagoonal-sabkha facies at the top. The sediments represent the relicts of sedimentary basin infill, which first developed in a transpressional tectonic regime and prograded to the initial stage of post-Variscan rifting.

The extremely mineralogically mature detritus of the Southern Gemeric Permian formations, allows us to correlate them with other early-Alpine riftogenic tectofacies of the Alpine-Mediterranean domain.

The Early Permian age of the lowermost part of the basal formation is assumed on the basis of microflora, with predominant species of the genera *Potonieisporites, Striatodisaccites, Vittatina sp.* (Planderová, 1980). The Upper Permian date for the uppermost part of the formation was proven on the basis of a cone slice and twig of *Pseudovoltzia liebeana* (Geinitz) Florin, and leaves of the genus *Sphenozamites*, as well as remains of bivalve tests of the genus *Carbonicola* McCoy, 1855 (Šuf, 1963).

The continental red beds unconformably overlying the Bashkirian flysch in the Turnaic Unit are most probably, Upper Permian in age. Upper Permian sediments of the innermost, Szendrö-Bükk Subzone, are represented by the evaporite formation (Silicic Unit). They were spatially connected with a Permian epiplatform, characterised by shallow-marine facies of South Alpine-Dinaric type. The Serpukhovian-Bashkirian turbidite deposits and the Moscovian shallow-marine sedimentary sequences are characteristics of the Szendrö-Bükk Subzone.

DISCUSSION

The typical distribution of the Carboniferous-Permian sedimentary basins and, the lithofacial character of their infill, document the southern polarity of the Hercynian orogeny of the Western Carpathians, responsible for the opposite vergency compared with the Alpine branch (Table 1). Basins originated gradually as a consequence of different stages of collisional events. The beginning of collisional events was connected with the "Bretonian" movements and development of the Lower Carboniferous flysch remnant basin sequences preserved within the Northern Gemeric Unit. Collision continued with the closure of the Lower Carboniferous flysch basin and caused a hiatus during Namurian B-C. The "Sudetian" movements gave rise to the Moscovian marine peripheral basin, whose basal sequences fixed the Lower Carboniferous flysch and both the pre-Carboniferous complexes (fragments of crust with oceanic/supraoceanic affinities). Closure of this basin was connected with Asturian events and is reflected by a hiatus during the Stephanian. The North Gemeric Lower and Upper Permian continental red beds originated under transpressional and extensional tectonic regimes.

The variously evolved basins were established on formerly overthrust continental lithosphere, whose fragments

VARISCAN orogen	INTERNAL						EXTERNAL						
Relics in paleo-Alpine units	T	NV	SV	Z	H	NG	SG	M	TU	S	U	SZ	B
Character of substratum of Late Paleozoic basins	continental crust					thrust wedges of oceanic crust	consuming plate-boundary fore-arc basin filling		disrupted continental crust, attenuated in places				
Tectono-thermal development	Devonian–Carboniferous tectono-thermal activity with maximum deformation, magmatism and metamorphism						weak or no magmatism; metamorphism more advanced than anchimetamorphism/low-temperature greenschist facies						
Main collisional events	Bretonian–Sudetian						Asturian						
Depositional system	continental: Permian Upper Carboniferous marine: shallow-water Moscovian and Serpuchovian deep-water post-Bretonian flysch						continental to sabkha-lagoonal: Permian sabkha to shallow-water: Permian marine: shallow-water Stephanian deep-water post-Sudetian flysch						
Provenance	continental block prov. cut magmatic arc prov. orogen. provenance						continental block provenance mixed with recycled orogen provenance						

Table 1 – Scheme of tectono-sedimentary evolution of the Western Carpathian Carboniferous Permian basin. *Explanation of symbols:* T - Tatric Unit; NV - Northern Veporic Unit; SV - Southern Veporic Unit; Z - Zemplinic Unit; H - Hronic Unit; NG - Northern Gemeric Unit; SG - Southern Gemeric Unit; M - Meliatic Unit; TU - Turnaic Unit; S - Silicic Unit; U - Upponyi Mts.; SZ - Szendrö Mts.; B - Bükk Mts.

are dismembered within the several Alpine mega-units. Continental sedimentation under humid climatic conditions was characteristic of this area during the Late Carboniferous, with relicts preserved within the Zemplinic, Hronic and Southern Veporic Units. Generally, a characteristic feature for this development is gradual progradation into the Permian arid/semi-arid red-bed formations. The sedimentary basins were established in pull-apart and rift-related tectonic settings.

Particular features have been observed in the Permian formations of the Southern Gemeric Unit. Their extremely mineralogically mature detritus, compared with other contemporary sediments of the Western Carpathians, allows their correlation with other early-Alpine riftogenic tectonofacies of the Alpine-Mediterranean Domain. They were most probably areally connected with the Bashkirian flysch depositional zone of the Turnaic Unit (the Turiec Formation, in Brusník Anticline; Vozárová & Vozár, 1992), as well as with the corresponding facies in the SBsZ.

CONCLUSION

Two zones of Hercynian collision at different times were identified in the Western Carpathians on the basis of relict infilled Carboniferous-Permian basins:

– *the internal zone*, in which the collision terminated during "Bretonian-Sudetian" events and resulted in the formation of a syncollisional Lower Carboniferous flysch basin succeeded by a marine Upper Carboniferous and continental Permian peripheral foreland basin (NGZ) as well as the formation of continental Permian back-arc transpressional/transtensional retro-arc basins (CWCZ). With respect to these characteristics, this zone corresponds to the Mediterranean Crystalline Zone as defined by Neubauer & Raumer (1993), comprising the eastern Grauwackenzone and Gurktal thrust system and the Carboniferous of the Noetsch area in the Eastern Alps, the nappe thrust system of the Bucovinian and Ghetic Units in the Eastern and Southern Carpathians in Romania, and part of the Balkanides and Kraishtides in Bulgaria;

– *the external zone*, in which the collision terminated during the "Asturian" events (IWCZ). This zone is represented by the Bashkhirian flysch and the Upper Westphalian/Stephanian marine peripheral foreland basin (relicts preserved only in northern Hungary), and the post-orogenic continental Permian deposits (the Turnaic Unit in Slovakia). This zone was spatially connected with shallow-marine foreland deposits of the Bükk Mts. area in northern Hungary. Generally, this whole external zone could be correlated with the Noric-Bosnian and Betic-Serbian zones distinguished by Neubauer & Raumer (1993).

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OUTLINE OF THE PERMIAN PALEOGEOGRAPHY IN THE CENTRAL AND EASTERN PART OF THE BALKAN PENINSULA

SLAVCHO YANEV [1], LJUBINKA MASLAREVIC [2] and BRANISLAV KRSTIC [3]

Key words – Paleogeography; facies; sedimentology; terranes; Balkans; Carpathian; Upper Paleozoic; Carboniferous; Permian.

Abstract – Upper Carboniferous formations are preserved within the present territory of the Balkan Peninsula in a system of paleogeographical zones: (1) the very high "Paleo-Balkan" mountain range running in a WNW-ESE direction (Variscan orogeny); (2) the rolling plain north of the range; (3) intermontane hills west and south of the orogenic belt; (4) high mountains further west and south of the orogenic belt (to the present Serbian-Macedonian and Thracian massifs); (5) a shallow sea basin west and south of the massifs. Principal paleogeographical zones changed slightly in configuration through the Early Permian times. Sedimentation covered large, still isolated, areas.
The Thracian and Serbian-Macedonian massifs were preserved, as well as the high orogenic features, in and around which were deposited separate, often very thick cones of dominantly coarse proluvial-colluvial clastics. North of the orogenic belt, towards the Moesian region, layers of fine sediments continued to form. On the southern and western piedmont of the Variscan range, coarse clastics and volcaniclastics were deposited in intermontane zones, and finegrained beds at high levels more distant from range.
The Variscan mountain range was lower through the Late Permian, and locally covered with deltaic sediments that extended northeastwards into a larger continental basin. Terrigenous deposits in the Provadia Depression pass into evaporites formed in sabkha facies. Another large basin was formed in the intermontane zone. There is not a facial transition between the continental and marine Permian (and Upper Carboniferous) sedimentary rocks. This fact can been explained by a separation of the two sedimentation areas, because of paleorelief existing between them.

Parole chiave – paleogeografia; facies; sedimentologia; *terranes*; Balcani; Carpazi; Paleozoico superiore; Carbonifero; Permiano.

Riassunto – Il Carbonifero superiore è presente nella Penisola Balcanica all'interno di un sistema di zone paleogeografiche: (1) l'alta catena montuosa Paleo-Balcanica che decorre secondo una direzione WNW-ESE (dovuta all'orogenesi varisica); (2) la pianura ondulata a nord della catena; (3) le colline intramontuose ad ovest e a sud del *belt* orogenico; (4) le pronunciate montagne poste più oltre ad ovest e a sud del *belt* orogenico (sino agli attuali massicci Serbo-Macedone e Tracio); (5) un bassofondo marino sviluppato ad ovest e ad est dei suddetti massicci. Le principali zone paleogeografiche modificarono leggermente la loro configurazione nel corso del Permiano inferiore. I processi di sedimentazione si attuarono all'interno di ampie, ancora isolate aree. I massicci Tracio e Serbo-Macedone furono preservati, come pure i più pronunciati lineamenti orogenici entro e attorno ai quali si depositarono, tra loro separati, conoidi di materiali clastici spesso assai potenti e in prevalenza grossolani, di natura proluviale e colluviale. A nord del *belt* orogenico, in direzione della Piattaforma Moesia, continuarono ad aversi sedimenti fini. Nelle aree pedemontane a sud e ad ovest della catena varisica confluirono, in aree intramontuose, apporti detritici grossolani e prodotti vulcanoclastici, nonché, a livelli stratigraficamente alti e in luoghi relativamente più distanti dalla catena, sedimenti clastici fini. I lineamenti morfologici della catena varisica si attenuarono nel corso del Permiano superiore, e localmente furono ricoperti da depositi deltizi che si estesero verso nordest a formare un più ampio bacino continentale. Nella depressione di Provadia, depositi terrigeni passano a evaporiti di sabkha. Un secondo ampio bacino si generò nella fascia intramontuosa. Non si osserva alcuna transizione di facies tra Permiano (e Carbonifero superiore) continentale e marino. Ciò si può spiegare ammettendo una separazione tra le due aree di sedimentazione, determinata dalla presenza di un paleorilievo.

PRE LATE PALEOZOIC DEVELOPMENT

The Late Paleozoic development and paleogeography of the present central part of the Balkan Peninsula were largely predetermined by the Early Paleozoic evolution (Fig. 1). The eastern part of the region (on Bulgarian territory) consists of three large Lower Paleozoic tectonostratigraphic superterranes (composite terranes): Moesian, Balkan and Thracian (Yanev, 1993, 1997, and references therein). The Balkan superterrane is composed of several lower-rank

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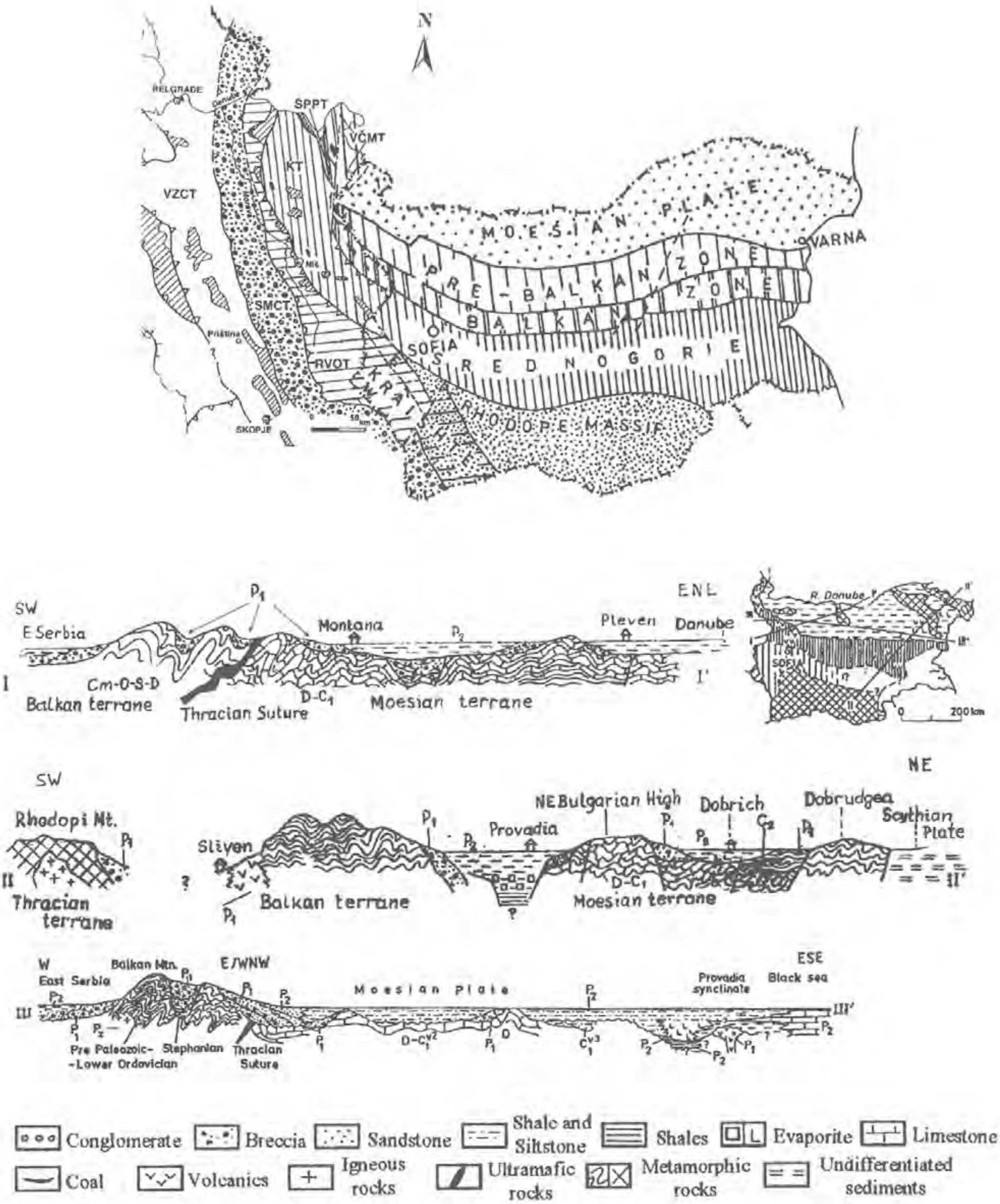

Fig. 1 – Scheme for the relationship between the main Bulgarian and Serbian tectonostratigraphic units. The abbreviations from east to west: MP - Moesian Plate; VCMT - Vrska Chuka-Miroc Terrane; SPPT - Stara Planina -Porec Terrane; KT - Kuchaj Terrane; RVOT - Ranovac-Vlasina-Osogovo Terrane; SMCT - Serbian-Macedonian Composite Terrane; VZCT - Vardar Zone Composite Terrane; E KRAI. and W KRAI.– Eastern and Western Kraishte, respectively. Moesian Plate – Moesian Terrane; Pre-Balkan Zone, Balkan Zone, Srednogorie and Kraishte – Balkan Composite Terrane; Phodope massif – part of the Serbo-Macedoniam-Thracian Composite Terrane. (The Serbian part after Krstic & Karamata, 1992; the Bulgarian part after Bonchev, 1986 – see Yanev & Cassinis, 1998).

Chronostratigraphical abbreviations in the sections: Cm - Cambrian; O - Ordovician; S - Silurian; D - Devonian; P_Z - Paleozoic *s.l.*; C_1^{v2} - Lower Carboniferous, Middle Visean; C_1^{v3} - Lower Carboniferous, Upper Visean; C_2 - Upper Carboniferous; P_1 - Lower Permian; P_2 - Upper Permian.

units, including the Western Kraishte, Eastern Kraishte and Sakar-Standja zones. Within the boundaries of ex-Yugoslavia, as well as the extensions of the Moesian and Balkan (Carpatho-Balkan) superterranes, are the Serbian-Macedonian, the Vardar and the Dalmatian-Hercegovinan, and between the latter two are the Jadar, Drina – Ivanjica and Dinaridic ophiolite belts, East Bosnian-Durmitor, and Central Bosnian terranes (Krstic & Karamata,1992; Karamata & Krstic,1996; Fig. 2).

The latter seven units extend over to Greece and partly to Albania. The Serbian-Macedonian superterrane is definitely connected with the Thracian one, the Balkan with Carpathian in Romania (Krautner *et al.*, 1981), and the Moesian superterrane forms the Paleozoic basement beneath the Walachian depression of Romania to the fault bounding the northern Dobrogea. Boundaries between the mentioned tectonostratigraphic units are tectonic and repeatedly active, including the significant Alpine overthrust. It is noteworthy that the sediment origin data for adjacent blocks sometimes distinctly indicate that these crustal blocks, although their contacts have been later tectonised or individual blocks rotated, were placed a similar order and only mutually displaced (drawn closer and in places thrust one over another). At the present time the terranes form a complex collage of peri-Gondwana blocks that successively docked from the south during the Late Paleozoic and Early Carboniferous (Visean).

LATE PALEOZOIC PALEOGEOGRAPHY

The most important explanation of the Late Paleozoic paleogeography of the region (Balkan Peninsula), and especially of its continental sedimentation, is the collision of the Balkan-Carpathian and Moesian superterranes, in the interval between the Early and Late Carboniferous epochs, which produced the Variscan range. The Thracian and Serbian-Macedonian massifs also collided with the Carpatho-

Fig. 2 – Schematic map of the terranes in the central part of the Balkan Peninsula between the Moesian Plate and the Adriatic Sea (after Karamata & Krstic, 1996) showing the distribution of the Carboniferous rocks (after Krstic *et al.*, in press)

Legend: 1 - fault, observed and covered; 2 - overthrust, observed and covered; 3 - unclear relationship; 4 - tectonised boundary; 5 - major strike-slip faults; 6 - marine Carboniferous; 7 - continental Carboniferous. Abbreviations for the Yugoslavian area (corresponding partly to that in Fig. 1) from east to west: MP - Moesian Plate; VCMT - Vrska Chuka-Miroc Terrane; SPPT - Stara Planina-Porec Terrane; KT - Kuchaj Terrane; RVOT - Ranovac-Vlasina-Osogovo Terrane; SMCT - Serbian-Macedonian Composite Terrane; VZCT - Vardar Zone Composite Terrane ; JBT - Jadar Block Terrane; DIT - Drina-Ivanjica Terrane; DOBT - Dinaric Ophiolite Belt Terrane; EBDT- East Bosnian-Durmitor Terrane (A - Lim and Tara catchment area, B - Praca region); CBMT - Central Bosnian Mts. Terrane (C - Central Bosnian Mts region, D - Sana region, E - Una region); DHCT - Dalmatian-Herzegovinian Composite Terrane.

Balkan superterrane, but this hypothesis cannot beverified due to a lack of Upper Paleozoic sediments on the massifs. Ophiolites of the pre-existing oceanic crust are compressed between these three superterranes in many places (Haydoutov, 1987; V. Quadt *et al.*, 1998).

The region is paleogeographically reconstructed using biostratigraphy, as well as the spatial distribution of facies and the thickness variations of series, stages, and lithic bodies, petrographic compositions with respect to the source area of derived material, and paleotransport measurements (Figs 3, 4, 5 and 6) (Janev, 1969; Yanev, 1970, 1979, 1981, 1989; Maslarevic & Protic, 1975). The principal criterion is the facial character of the Upper Paleozoic sediments. Information on the types and distributions of continental sediments and facies in the Upper Paleozoic series of Serbia is given by Maslarevic & Krstic (1999), and for Bulgaria by Yanev in Yanev & Adamia (1999), Cassinis & Yanev, 1999 and Yanev *et al.*, (1999) in the volume of abstracts of the Brescia congress, and in this book.

Late Carboniferous paleogeography

Upper Carboniferous formations are preserved on the present territory of the Balkan Peninsula in a system of different paleogeographical zones: (1) the high "Paleo-Balkan" mountain range running in a WNW-ESE direction (Variscan orogeny); (2) the rolling plain north of the range; (3) intermontane hills west and south of the orogenic belt; (4) high mountains further west and south of the orogenic belt (to the present Serbian-Macedonian and Rila-Rhodope massifs); (5) a shallow sea basin west and south of the massifs (Fig. 4). Theprevious zones, according to the available data, were areas only of denudation, not deposition. Isolated depressions of the paleo-Balkan zone contain continental, often coal deposits, varying in thickness from 0 to 1700 m in the Westphalian of Svoge; from 0 to 850 m. in the Stephanian between Ozirovo and Lyutadzik, NW Bul-

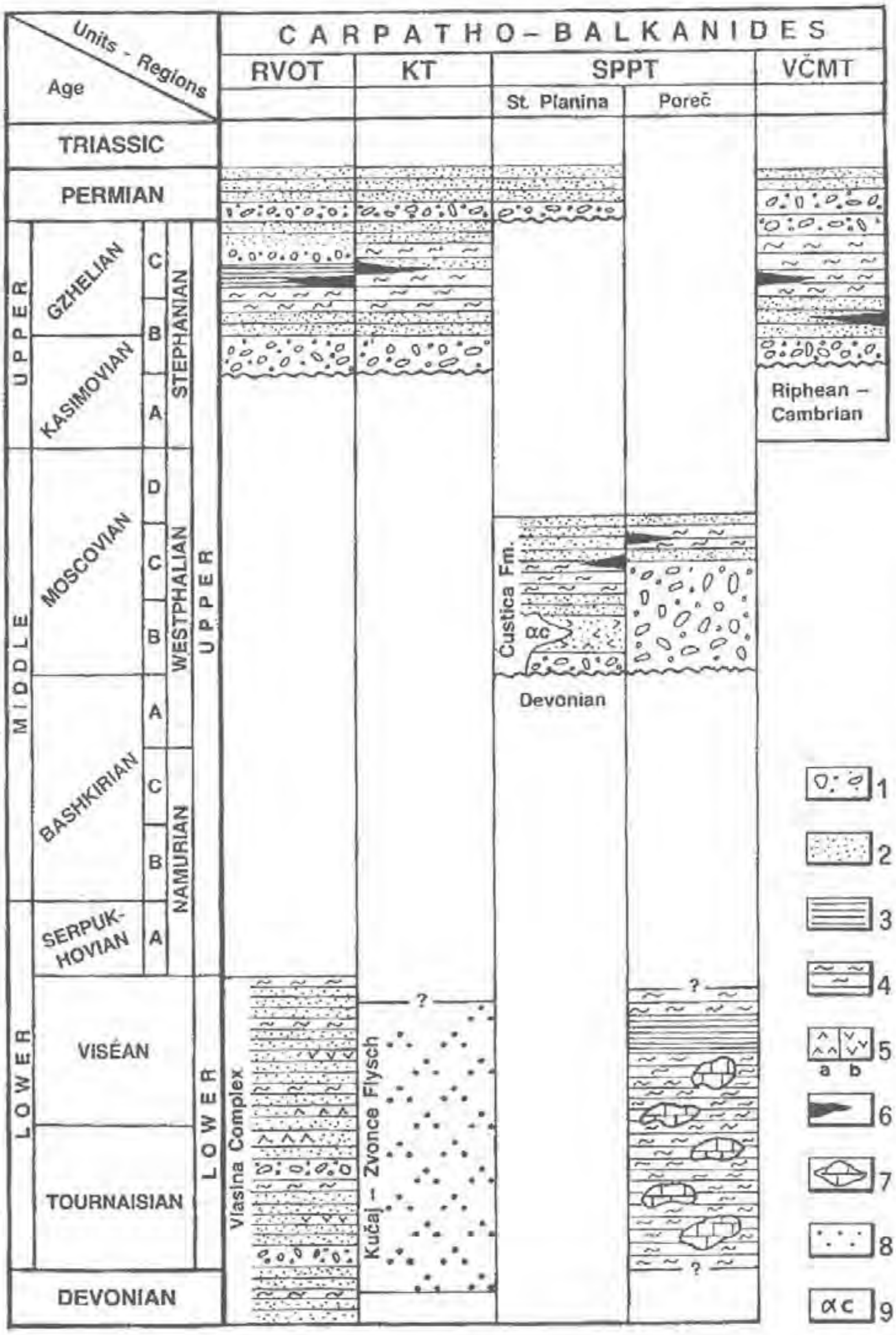

Fig. 3 – Schematic column for the Carboniferous successions in their main zones of distribution in East Serbia (after Krstic *et al.*, in press). Legend: 1. Conglomerate; 2. Sandstone; 3. Shale; 4. Siltstone; 5a. Basic volcanic rocks; 5b. Acidic volcanic rocks; 6. Coal; 7. Olistostromes; 8. Flysch; 9. αc: Paleoandesite.

garia; and from 0 to 360 m in East Serbia (Maslarevic, 1961; Yanev, 1985) (Fig. 5). Basins are supplied with rock clasts derived from the basement and adjacent areas (Lower Paleozoic sediments and low-crystalline metamorphic rocks). The paleo-rivers of the western Stara Planina Mountain drained to the north and northeast, and those of Eastern Serbia to the southwest (Vrshka Chuka). Late Carboniferous tectonic movements, mainly block displacements (Late Variscan?), transformed some of the Namurian-Westphalian depressions from depositional into denudational areas (Svoge), and formed new grabens (seven in Bulgaria and three in Serbia) during the Stephanian. A Westphalian volcano-sedimentary formation was formed in the Stara Planina-Porecka zone of Eastern Serbia only, unconformably overlain by Permian rocks (Krstic & Maslarevic, 1970). Westphalian deposits are lacking to the ENE and WSW of this zone, where Stephanian sediments lie unconformably over crystalline schists or over Lower Paleozoic rocks and pass upwards into Permian deposits (Krstic *et al.*, in press; Fig.3). The trough-bounding faults often were courses for subsequent volcanic flows. Reliable Upper Carboniferous sediments in the Moesian region are known from boreholes, but so far only at Kavarna-Balchik, Southern Dobrogea. More than 1600 m of Namurian-Westphalian clastic and coal deposits of alluvial plains are met by a marine basin to the east (or SE?). The larger part of the Moesian region was probably a major sediment bypass zone during the Late Carboniferous. Intermontane valleys in Bulgaria bear records

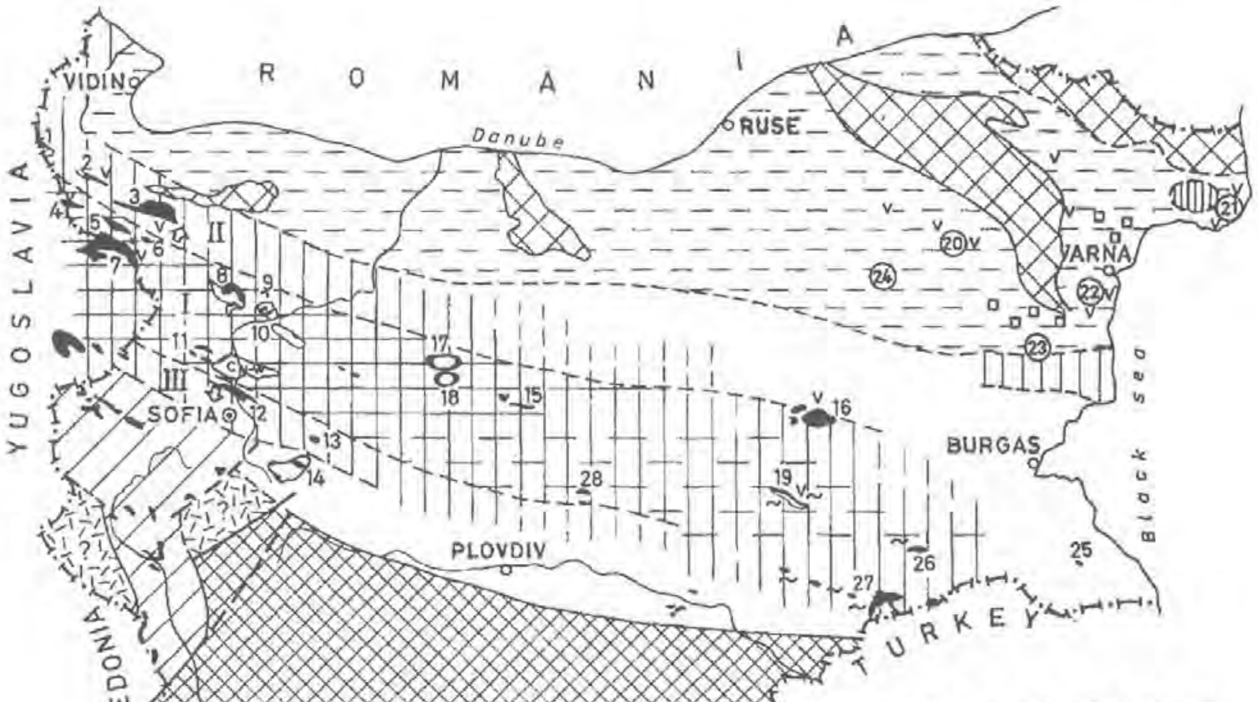

Fig. 4 – Map showing the distribution and paleogeography of the Upper Carboniferous and Permian deposits in Bulgaria.

Legend: 1 - Variscan Orogen: I - zones of deposition of intermontane molasses, II - zones of foremontane molasses, III - zones of intramontane molasses (former back-arc zones); 2 - probable continuation of the zones I, II and III; 3 - greatly uplifted part of the orogen; 4 - probably as 3; 5 - zone of distribution of intramontane molasses (during the Late Permian, mainly continental basin type); 6 - zones of molasse sedimentation with hilly relief (during the Late Permian mainly continental basin type); 7- coal-bearing Carboniferous deposits in the Dobrudja coal basin; 8 - zone with evaporite-bearing molasses; 9 - zones of most intense volcanism; 10-12 - sediment source regions: 10 - probably uplifted zones, out of the orogen, 11 - intrabasinal low terrains, 12 - hypothetical feeder zones; 13 - present-day outcrops: a - of the Upper Carboniferous, b - of the Permian; 14 - main paleotransport directions; 15 - Upper Paleozoic rocks, affected by low-grade metamorphism. Abbreviation: NW = Namurian-Westphalian deposits. Localities of Upper Paleozoic sedimentation: 1 - Kiryaevo; 2 - Belogradchik; 3 - Falkovets-Smolyanovtsi-Vinishte; 4 Stakevtsi; 5 - Prevala; 6 - Melyane; 7 - Midzhur-Kopren; 8 - Draganitsa-Bela Rechka-Ozirovo-Lyutadzhik; 9 - Zverino; 10 - Ignatitsa; 11 - Svoge (Namurian-Westphalian); 12 - Sofioter Stara Planina Mts; 13 - Bunovo; 14 - Lozen Planina Mts-Bakarel Hills; 15 - Troyan; 16 - Sliven; 17 - Glozhene; 18 - Vasilyovo; 19 - Sveti Ilia Hills; 20 - Borehole Vassil Levski; 21 - Borehole Kaliakra; 22 - Boreholes Varna-Ravna Gora; 23 - Boreholes Mirovo; 24 - Borehole Vetrino; 25 - Kondolovo - Strandja (marine Lower Permian in allochthonous position?); 26 - Klokotnitza (Upper Carboniferous?); 27 - East Rhodopi (redeposited pebbles from marine Upper Permian); 28 - Chernogorovo. (In circles - after drill-hole data; the others, after outcrop data).

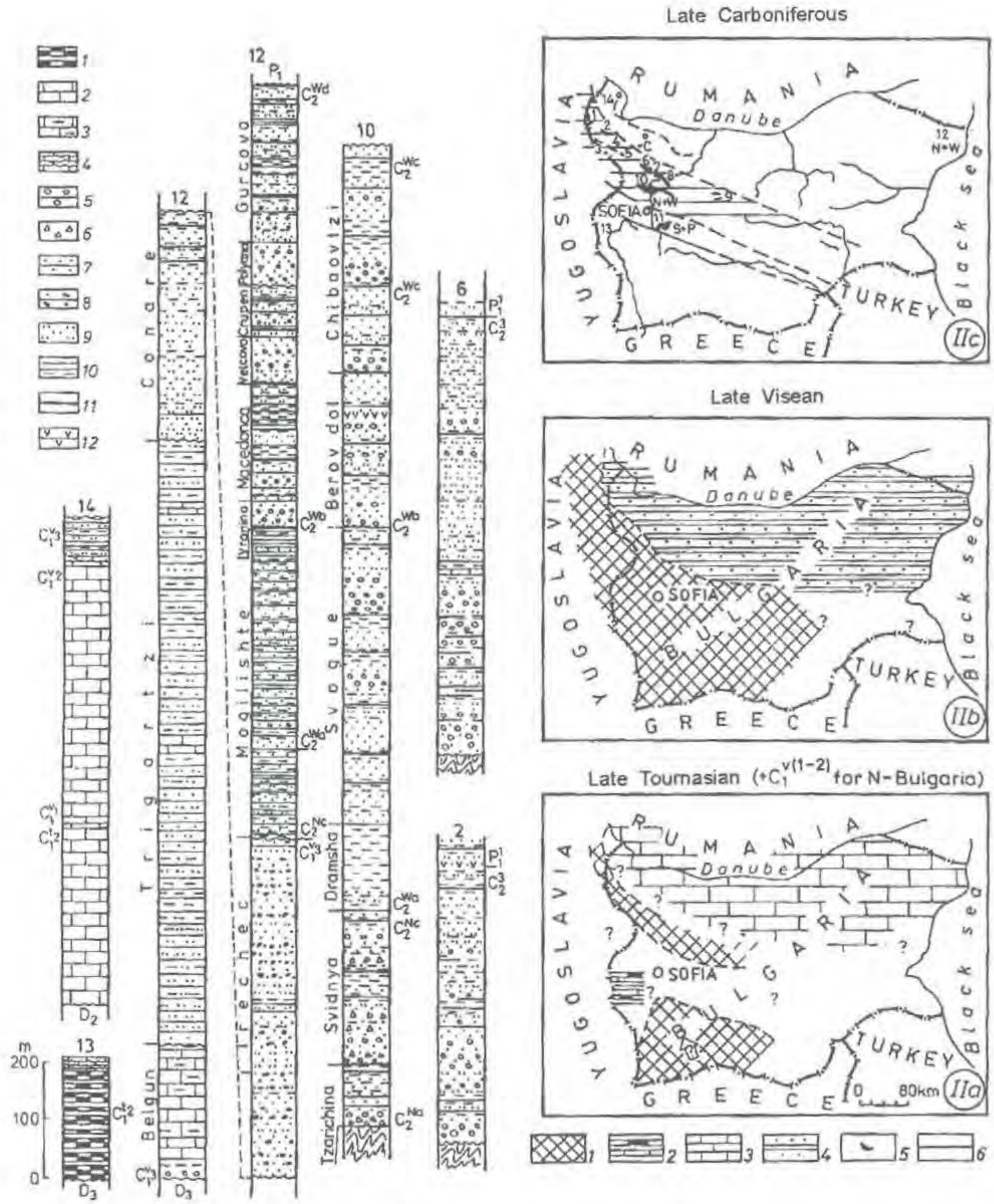

Fig. 5 - The Carboniferous System in the Bulgarian part of the Balkan Peninsula.

Legend: I. Lithological columns: 1 - Cherts; 2 - Limestones; 3 - Shaly limestones; 4 - Wave shaped and nodular limestones and marls; 5 - Conglomerates; 6 - Breccias; 7 - Sandstones; 8 - Sandstones with coal fragments; 9 - Siltstones; 10 - Shales; 11 - Coal.; 12 - volcanic rocks. II. Lithological-paleogeographical schemes for Bulgaria during the Carboniferous times. IIa - Late Tournaisian: 1 - dry lands; 2 - area with flint-carbonate-shaly sedimentation; 3 - area with shallow marine carbonate sedimentation. IIb – Late Visean: 4 - area with terrigenous-shaly sedimentation. IIc - Late Carboniferous: 5 - present day outcrops; 6 - distribution of continental sediments (see Fig. 4 for more detail). Abbreviations in the columns: D_3 - Upper Devonian; C_1^{t2} - Upper Tournaisian; C_1^{v1} - Lower Carboniferous, Lower Visean; C_1^{v2} - Lower Carboniferous, Middle Visean; C_1^{v3} - Lower Carboniferous, Upper Visean; C_2^{Na} - Upper Carboniferous, Namurian A; C_2^{Nc} - Upper Carboniferous, Namurian C; C_2^{Wa} - Upper Carboniferous, Westphalian A; C_2^{Wb} - Upper Carboniferous, Westphalian B; C_2^{Wc} - Upper Carboniferous, Westphalian C; C_2^{Wd} - Upper Carboniferous, Westphalian D; C_2^{3} - Upper Carboniferous, Stephanian; P_1^{1} - Lower Permian. Abbreviations in the schemes: $C_1^{v(1-2)}$ - Lower Carboniferous, Lower to Middle Visean.

of Stephanian, dominantly alluvial-lacustrine sedimentation on Lozen Mt. (south of Sofia). The deposited materials were derived from the Thracian massif. Stephanian sediments in the East-Serbian extension of this zone are interpreted as colluvial-alluvial and lacustrine deposits. There is no evidence of Late Carboniferous sedimentation in the Thracian and Serbian-Macedonian massifs.

Permian paleogeography

The principal paleogeographical zones changed slightly in configuration through the Early Permian. Sedimentation covered large, still isolated areas. The Thracian and Serbian- Macedonian massifs were preserved, as well as the high orogenic features in and around which were deposited separate, often very thick (up to more 3000 m) cones of dominantly coarse proluvial-colluvial clastics. North of the orogenic belt, toward the Moesian region, bands of fine sediments (cones tapering toward the periphery, often in playa facies) continued to form (or

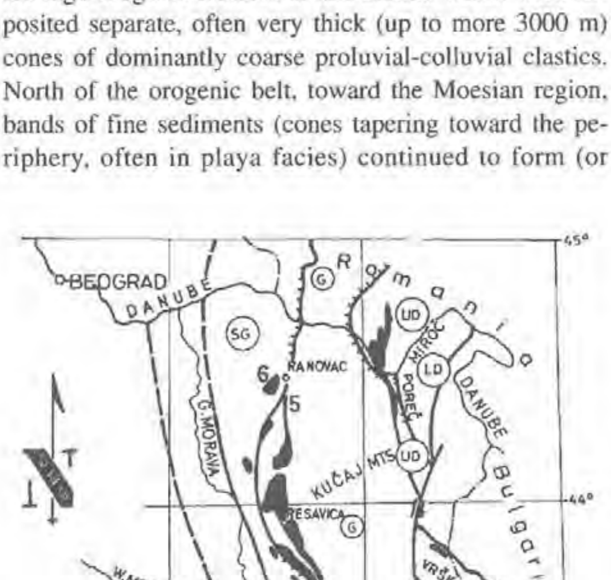

Fig. 6 – Distribution scheme of the continental Permian sediments in East Serbia (after Maslarevic & Krstic, Fig 1, this volume). The stratigraphy and lithologies of the various Permian successions are presented by Maslarevic & Krstic in Fig. 2, this volume). Abbreviations for the tectonic units: LD - Lower Danubicum; UD - Upper Danubicum; G - Gethicum; SG - Supragethicum; SMM - Serbian-Macedonian massif.

formed around new highs). On the southern (and western?) piedmont of the Variscan range, coarse clastics and volcaniclastics were deposited in intermontane zones (*e.g.* the Sofian Stara Planina Mt. and Sveti Iliya Heights, in Bulgaria, and the Serbian part of the Stara Planina Mts.), and finegrained beds at high levels more distant from the range. The Lozen Planina Mt. provides evidence (composition of fragments and paleotransport measurements) of material supply from the Thracian terrane (Kozhoukharov *et al.*, 1980). A connection with a sea basin could have existed on the extreme southeastern periphery of the zone (Strandzha Mt.) where the Thracian massif ended. The small outcrops of Lower Permian marine sediments here (Malyakov & Bakalova, 1978) could even be allochthonous. Material in Eastern Serbia was washed down from the Serbian-Macedonian massif, as indicated by conglomerates (fragments of gneisses and various schists, including muscovite and amphibole-bearing).

The Variscan mountain range was lower through the Late Permian and locally covered (*e.g.* between Kopren peak and Montana town) with deltaic sediments that extended northeastward into a larger continental basin. This basin covered a large part of the Moesian superterrane over Lower Permian and older rocks. Another large basin was formed in the intermontane zone. It is well studied in Bulgarian Kraishte (Yanev, 1979) and a large part of Eastern Serbia, the so-called Inner Belt of Eastern-Serbian red sandstones, where two zones of Permian rocks are recognised: the western, which is part of the Supragethicum (between the Mlava and Pek rivers), and the eastern, which belongs to the Gethicum (Gornjak-Suva Planina Mts. zone) (Maslarevic, 1967; Fig.6).

Deposits of these two continental basins are predominantly finegrained red terrigenous sediments; locally, coarser material was deposited or parts of the basins were temporarily dry. Coarse clastic sediments from alluvial fans prevail at the top of the Permian in the extreme north of the Supragethicum towards the Danube (Krepoljin) (Maslarevic & Krstic, 1999).

Terrigeneous deposits at Provadiya (in NE Bulgaria) pass into evaporites (mainly halite) formed in sabkhas facies, indicating a probable periodic seawater incursion on the basin periphery. One hypothesis is that the extreme southeastern parts of the zone, equivalent to the southern basin, had some communication with the Late Permian sea, because Mesozoic conglomerates near Mandrica (easternmost Rhodope Mts) include redeposited carbonate clasts with Upper Permian fauna (Trifonova & Boyanov, 1986). Data are lacking for Upper Permian, either continental or marine, sedimentary rocks over the Thracian and Serbian-Macedonian massifs. Extending west and south of the massifs, in the Jadar terrane of northwestern Serbia, are marine sediments of the Middle and Upper Permian (Pantic & Pesic, 1975).

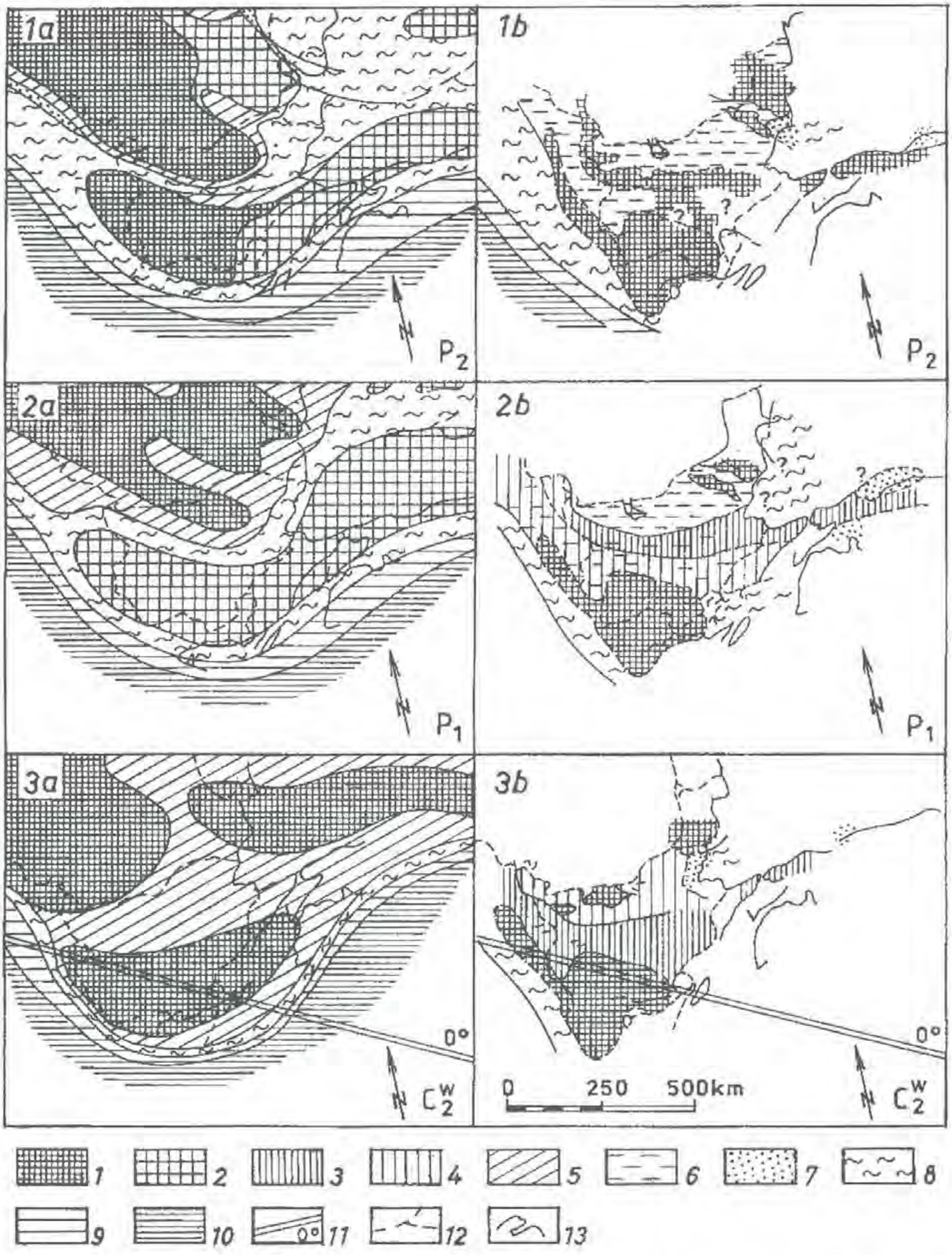

Fig. 7 – Comparative paleogeographical schemes for part of southeastern Europe during the "Westphalian" (C_2^w), Early (P_1) and Late Permian (P_2) times. 1a, 2a, and 3a: fragments from the maps of Ylmaz *et al.* (1996); 1b, 2b, and 3b: after Yanev (in print).

Legend: 1 - continental highlands (zones without deposition, sediment sources); 2 - continental lowlands (sediment bypass); 3 - continental highlands with intramontane (very limited) sedimentation; 4 - continental intermontane deposition; 5 - continental, fluvial, lacustrine deposition; 6 - continental basin deposition; 7 - deltaic and coastal plain deposition; 8 - marine shelf deposition; 9 - basin or slope deposition; 10 - deep ocean deposition; 11 - equator; 12 - state boundaries; 13 - block and coastline.

These shallow-water sediments, transgressive over the pre-existing rocks, consist of dolomites and clastics (Middle Permian) and limestones abounding with megafaunal associations of mixed Alpine, east-European and Indo-Armenian species (Late Permian). Shallow marine deposits are exposed in a limited area of the Thracian massif on some of the Greek islands (Lesvos, Skiros, etc), and further southeast in Turkey.

CONCLUSION

It should be mentioned in conclusion, that there is no facial transition between the continental and marine Permian (and Upper Carboniferous) sedimentary rocks. This fact can be explained by a separation of the two main sedimentation areas, due to the paleorelief existing between them. The position of this paleorelief corresponds of the location of the Thracian and Serbo-Macedonian massifs and is confirmed by the rock distribution and facies assemblages around the uplifted domains.

The revised paleogeographic picture of the Late Paleozoic of the Balkan Peninsula, based on numerous studies and regional generalisations (Maslarevic & Protic, 1975; Protic, 1978; Yanev, 1981; Yanev & Cassinis, 1998), essentially differs from any of the generalised reconstructions published to date (Vai, 1994, 1997; Yilmaz *et al.*, 1996; and others). The last-mentioned publication, for instance, describes a marine basin exactly where the range of the Variscan mountain chain was the highest (Fig.7).

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CONTINENTAL PERMIAN AND LOWER TRIASSIC RED BEDS OF THE SERBIAN CARPATHO-BALKANIDES

LJUBINKA MASLAREVIC[1] and BRANISLAV KRSTIC[2]

Key words – Continental sediments; Permian; Lower Triassic; East Serbia.

Abstract – Permian continental clastics in the Serbian part of the Carpatho-Balkan Mountains are post-orogenic sediments formed on the Hercinian land.
The clastics lie unconformably as overstep sequences on various pre-existing magmatic, metamorphic and sedimentary rocks, or are gradually derived from Upper Carboniferous (Stephanian) continental beds. Transitional between the Stephanian C and the Permian are the so-called Mottled Series with *Callipteris conferta*.
The Permian rock system is present as a component of four large Alpine geotectonic units: Vrska Cuka - Miroc (Lower Danubicum), Stara Planina-Porec (Upper Danubicum), Kucaj (Gethicum) and Ranovac-Vlasina-Osogovo (Supragethicum). Permian rocks of the Lower and Upper Danubicum contain andesite and quartz-porphyry extrusions, rarely including pyroclastics.
Generally, Permian clastics are immature deposits of ortho- and para-conglomerates, sandstones (arkoses and impure arkoses, rarely greywackes), siltstones and shales with organic markings. They are red or purple, rarely grey, in colour.
Permian clastics are deposited mainly in intramontane depressions with unstable tectonic regimes, in alluvial fans and meandering rivers; the Permian also includes thick alluvial plain, flood plain, and shallow lake deposits.
Palaeomagnetic data for Permian beds locate the primary depositional area at about latitude 8°N.
Continental deposits of the Lower Triassic lie unconformably over various units of the Permian red beds at a low angle (about 12°) and over Lower Paleozoic schists. The formation is composed of quartz and sub-arkose conglomerates (usually organised) and sandstones, then arkoses, siltstones, and shales mainly in the youngest member. Moreover, these clastics include fossil macroflora.
Lower Triassic rocks were deposited in braided rivers showing diverse sedimentary structures, in a more stable tectonic situation and lower-relief environment. The sediments are purple, white or red in colour.
The stratigraphic boundary between the Permian and Triassic Systems is not recognised everywhere, but is inferred mainly from sedimentological features.

Parole chiave – sedimenti continentali; Permiano; Trias inferiore; Serbia orientale.

Riassunto – I depositi clastici continentali permiani nella parte serba dei Monti Carpato-Balcani corrispondono a sedimenti post-orogenici formatisi sul basamento ercinico. Queste clastiti giacciono in discordanza come sequenze sovrapposte su varie rocce magmatiche, metamorfiche e sedimentarie preesistenti, o sono state gradualmente alimentate a seguito dell'erosione di strati continentali appartenenti al Carbonifero superiore ("Stefaniano"). La così chiamata "Serie Screziata" (*Mottled Series*), con *Callipteris conferta*, rappresenta una successione di transizione tra lo "Stefaniano" C e il Permiano.
Il Permiano è presente in quattro ampie unità geotettoniche alpine: Vrska Cuka-Miroc (Danubico inferiore), Stara Planina-Porec (Danubico superiore), Kucaj (Getico) e Ranovac-Vlasina-Osogovo (Supragetico). Rocce permiane appartenenti al Danubico inferiore e superiore contengono estrusioni di andesite e di porfido quarzifero, raramente commiste a depositi piroclastici.
Generalmente, i depositi clastici permiani sono depositi immaturi di orto- e paraconglomerati, arenarie (arcose e arcose impure, raramente grovacche), siltiti e shales con tracce organiche. Essi risultano di colore rosso o porpora, e raramente grigio.
Le clastiti permiane si sono soprattutto accumulate in depressioni intramontane controllate da regimi tettonici instabili, sotto forma di coni alluvionali e di depositi fluviali governati da meandri; il Permiano, pure, include potenti piane alluvionali, piane di inondazione e depositi lacustri poco profondi. Dati paleomagnetici raccolti negli strati permiani posizionano l'originaria area di deposione di quest'ultimi ad una latitudine di circa 8°N.
I depositi continentali del Triassico inferiore poggiano in discordanza, a basso angolo (circa 12°), sui vari membri che costituiscono i *red-beds* permiani e sopra scisti del Paleozoico inferiore. Essi consistono di quarzo, conglomerati sub-arcosici (in genere organizzati) e arenarie, inoltre di arcose, siltiti e *shales* soprattutto in corrispondenza dei livelli stratigrafici più recenti. Vi è anche presente una macroflora fossile. Il Trias inferiore si depositò in fiumi *braided* che mostrano varie strutture sedimentarie, nell'ambito di una situazione tettonica più stabile e in seno ad un ambiente caratterizzato da una topografia scarsamente pronunciata. I sedimenti sono di colore porpora, bianco o rosso. Il limite stratigrafico tra Permiano e Trias non è dovunque riconosciuto, ma è essenzialmente dedotto in base a caratteristiche sedimentologiche.

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INTRODUCTION

Permian and Lower Triassic rocks are found in two large Alpine orogenic systems in Serbia: the Carpatho-Balkanides and the Dinarides, which are very dissimilar (Fig. 1). The sequence in the Carpatho-Balkanides consists of Permian continental rocks (the so-called "Red Sandstone Formation"); in part of the Dinarides, the Permian system of rocks is lacking, so that Lower Triassic continental clastics lie unconformably over Lower Carboniferous flysch; Triassic marine limestones are transgressive over Lower Carboniferous volcanosedimentary units. Marine sedimentary rocks of the Middle and Upper Permian are found in other parts of the Dinarids,in NW and SW Serbia, transgressive over the pre-existing rocks.

Permian continental clastics in the Serbian part of the Carpatho-Balkanides are post-orogenic products laid over the Hercynian land. They are unconformable over various magmatic, metamorphic and sedimentary rocks, or derive from Upper Carboniferous ("Stephanian") beds. The Permian system crops out in four large Alpine geotectonic units: Vrska Cuka - Miroc, Stara Planina - Porec, Kucaj, and Ranovac-Vlasina-Osogovo.

Generally, Permian clastics are immature deposits of ortho- and para-conglomerates, sandstones (arkoses and impure arkoses, rarely greywackes), siltstones and shales with organic markings. They are red or purple, rarely grey, in colour. The sequence varies in thickness from 50 to 700 metres or more, formed in an arid or semi-arid climate.

Permian and Lower Triassic continental sediments were studied by Protic (1934, 1959, 1967, 1978), Petkovic (1937), Antonovic (1958), Pantic & Protic (1960), Maslarevic (1961, 1967, 1970, 1980), Maslarevic & Protic (1975), Bogdanovic (1980) and Maslarevic & Cendic (1995), and palaeomagnetism was investigated by Milicevic (1998).

PERMIAN CONTINENTAL SEQUENCES

The Vrska Cuka-Miroc Unit (Lower Danubicum)

This Unit is the extreme eastern part of the Carpatho-Balkanides of eastern Serbia (Figs 1, 2). Lower Paleozoic and Lower Carboniferous rocks are lacking there, whilst Permian rocks are largely eroded, preserved only over a small area, and corresponding to the Lower Rotliegend. Permian strata lie conformably over Upper Carboniferous ("Stephanian") continental deposits with coal beds, and consist of conglomerates, sandstones, siltstones and shales. Conglomerates lie lowest in the column and include pebbles of pre-Permian limestones, quartz porphyry, and Lower Paleozoic sandstones. Beds of conglomerates are quite thin and pass upwards into a sequence of sandstones and prevailing siltstones and shales which include quartz porphyry exstrusions. The sediments were deposited at first in river valleys, and later between river channels, mainly in shallow lakes.

The Stara Planina-Porec Unit (Upper Danubicum)

This Unit is found to the southwest of the Lower Danubicum (Figs 1, 2). Permian rocks in the Stara Planina part of the Unit lie unconformably over Riphean/Cambrian schists of the Crni Vrh Formation, Devonian rocks of the Inovo Formation, and "Westphalian" volcanosedimentary products of the Custica Formation, and are referred to in the geological literature as "the outer belt of Permian Red Sandstones" (Maslarevic & Protic, 1975).

Two localities, Sugrin and Topli Do, have outcrops of the Toplodolska Formation. The lowest-lying rocks are basal conglomerates, several metres thick at Sugrin; much coarser conglomerates and breccias up to 200 metres thick occur at Topli Dol (Babin Zub). These are para- and orthoconglomerates alternating with sandstone and rarely siltstones. The conglomerates are massive, rarely stratified, occasionally repeatedly graded, often of disorganised and unsorted subangular material which includes sills of porphyrite, quartz-porphyry and tuff.

Conglomerates are composed of porphyrite, quartz-porphyry, granite pebbles, limestone cobbles, schist, quartz and feldspar in a sandy matrix. Porphyrite fragments prevail in places. Upwards in the sequence, porphyrite disappears and the rock is arkosic in type. Conglomerates are clast- to matrix- supported. The sequence fines upwards; coarse-grained rocks are overlain by variably sized arkoses, a few tens of metres thick. Arkosic rocks are medium- to thick-bedded, graded in places, composed mainly of quartz, feldspar, and granite fragments; the minor constituents are porphyrite and rarely mica, schist, diabase, acid volcanic rocks and chert fragments. Arkoses are clast- to matrix- supported rocks with an illite matrix and hematite or calcite cement.

The rocks at Sugrin were interpreted as river-bed deposits and those at Topli Dol as alluvial fan deposits from proximal parts including debris-flow sediments and from the middle part with Gm and Sh lithofacies (Miall,1977), deposited in high relief on intermontane slopes.

The rapid, irregular deposition of the lowermost Permian ("Lower Rotliegend") is followed by deposition of better-sorted finegrained sediments. Upwards, finegrained sandstones and siltstones prevail, and then shale to a total thickness of nearly 400 metres (Sugrin). There are also thin beds of medium-grained sandstones or arkoses. The finegrained rocks are often horizontally laminated bearing symmetrical ripple-marks, organic markings, and limestone, rarely dolomite concretions. Mud chips in parallel orientation are common. Shales are marly in places with $CaC0_3$ to 28%. These are dominantly interchannel, flood plain, lake, and playa sediments. Scarce thin beds of coarser rocks are the probable remains of crevasse splays.

Finegrained rocks of the Topli Dol Formation are overlain by coarse deposits of a new sedimentary cycle in the Upper Permian. These are immature arkosic conglomerates and coarse, rarely medium-grained, arkoses of subangular unsorted material, either in beds of less than 60 cm or massive, about 30 metres in thickness. Horizontal and gently inclined tabular cross-bedding and asymmetrical ripple-marks can be found, and the series fines upwards. Arkosic rocks contain intraclasts from the base in calcite and haematite cement with scanty dolomite and clay matrix, deposited in the middle part of alluvial fans, possibly from braided channels and river-beds. Magnetite and ilmenite, locally epidote, are the dominant heavy minerals in Permian rocks, whereas stable tourmaline, rutile and zircon are scanty, all subangular to subrounded.

Palaeotransport is dispersed, but mainly towards the east, less to the west.

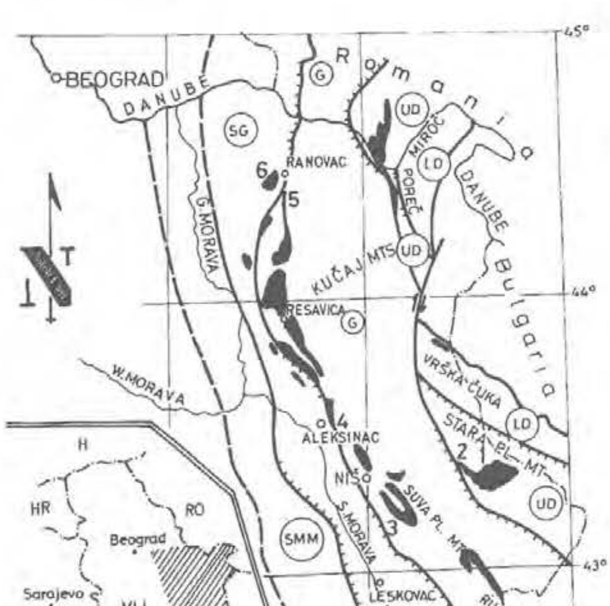

Fig. 1 – Main geotectonic units of eastern Serbia: SMM - Serbian-Macedonian massif; SG - Ranovac-Vlasina-Osogovo Unit (Supragethicum); G - Kucaj Unit (Gethicum); UD - Stara Planina - Porec Unit (Upper Danubicum); LD - Vrska Cuka - Miroc Unit (Lower Danubicum). Lithological columns: 1. Temska (UD); 2. Sugrin (UD); 3. Suva Planina (G); 4. Budina (G); 5. Zdrelo (G); 6 Ranovac (SG).

Similar Permian deposits are situated in the northern, Porec part of the unit. It is composed of conglomerates, sandstones and shales, alternating with pyroclastics and volcanic rock flows, which lie transgressively over pyroxene gabbro of Glavica (at Donji Milanovac), and green schists on the left side of the Porecka reka. Bogdanovic (1980) distinguished two Permian horizons: sedimentary and volcanosedimentary. The terrigenous horizon forms the lower part of the Permian sequence, from conglomerate and conglomerate breccia at the base to red sandstone and shale intercalated with freshwater limestone at the top. The volcanosedimentary horizon is composed of red sandstone, shale, volcanic breccia, tuff, and quartz porphyry and porphyrite extrusions.

The Kucaj Unit (Gethicum)

This Unit has a central and western position in eastern Serbia (Figs 1, 3). It includes several continental basins of graben or semi-graben type. Permian (and Lower Triassic) continental rocks extend discontinuously from the Danube in the north to Ruj Mt. in the south, over 200 km. It is referred to in the literature as "the Inner belt of Permian Red sandstones of eastern Serbia"(Maslarevc & Protic,1975). The system is unconformable over the older metamorphosed rocks at the base, or passes from Upper Carboniferous ("Stephanian") beds. It is believed to form the complete Permian sequence between the underlying ("Stephanian") and overlying (Lower Triassic) units.

Permian rocks make up a thick sequence of clastics, from 50 to 700 metres or more in thickness. Its constituents are sandstones and conglomerates, siltstones and shales. Dolomite and limestone are scanty. Rocks are red or purple, rarely grey or green, in medium to thin beds, rarely thick or massive. Horizontal bedding and lamination, thick tabular cross-bedding and asymptotic bedding are common, and current lamination can be seen. Conglomerate-sandstone grading with lower bed-surface erosion and clay chips also occur. Bed surfaces show fine symmetrical and asymmetrical ripple-marks. Finegrained rocks bear concretions of calcite and dolomite, rarely hematite, organic markings, raindrop imprints, and rarely desiccation cracks. There are many bleached marks around active centres (organic matter). Rocks show characteristic fluvial fining-upward sequences.

The southernmost Permian rocks of this Unit are those of Ruj Mt. and at the Vlaska and Greben mountains, unconformably overlying the older amphibolites or Devonian-Lower Carboniferous flysch. The entire Permian sequence (and the basal Unit) is fining upwards: disorganised, quite thin basal conglomerates that alternate with coarse sandstones, medium-grained sandstones which pass into a succession of finegrained sandstones and siltstones, for a total thickness of more than 100 metres.

These are unconformably overlain by Lower Triassic quartz conglomerates and sandstones. The Permian sequence is interpreted as alluvial fan clastics passing upwards into sediments filling river channels, topped by plain rivers and lake (playa) deposits.

The Permian stratigraphic sequence in this Unit is almost complete in the Suva Planina region (Fig. 2). It begins with pebbly sandstones and partly conglomerates, unconformable over Upper Devonian-Lower Carboniferous flysch or progressively developed from Upper Carboniferous ("Stephanian") rocks. Upwards follow coarse and medium-grained sandstones, and subsequently hundreds of metres of a monotonous succession of siltstones alternating with fine-grained sandstones. The coarsest constituents are interpreted as river channel deposits that rapidly pass upwards to finer-grained sediments of plain rivers and further to shallow lake and playa deposits, which dominate.

Although thin (about 100 m), the sequence is complete in the Budina stream near Aleksinac, between the underlying "Stephanian" and the overlying Lower Triassic rocks with megafloral remains (Fig. 2). Constituent rocks are siltstones and shales with scarce sandstone beds and lenses and a few coarse-grained sandstone and conglomerate interbeds: deposits of meandering rivers and fine-grained interchannel deposits of alluvial plain, flood plain, and playa sediments deposited from suspension.

Two zones of Permian rocks, divided by regional dislocations, are recognised to the north around Resavica and between Donja Mutnica and Rujevica, southwest of Samanjac.

Rocks mostly from the middle of the Permian sequence, composed of medium- to finegrained sandstones, siltstones and coarse-grained sandstones, rarely conglomerates and shales, are found in the Resavica-Josanica zone.

These rocks contain limestone and dolomite concretions. The sequence consists of thin, medium, and locally thick beds (tens of metres) beds. The bed-

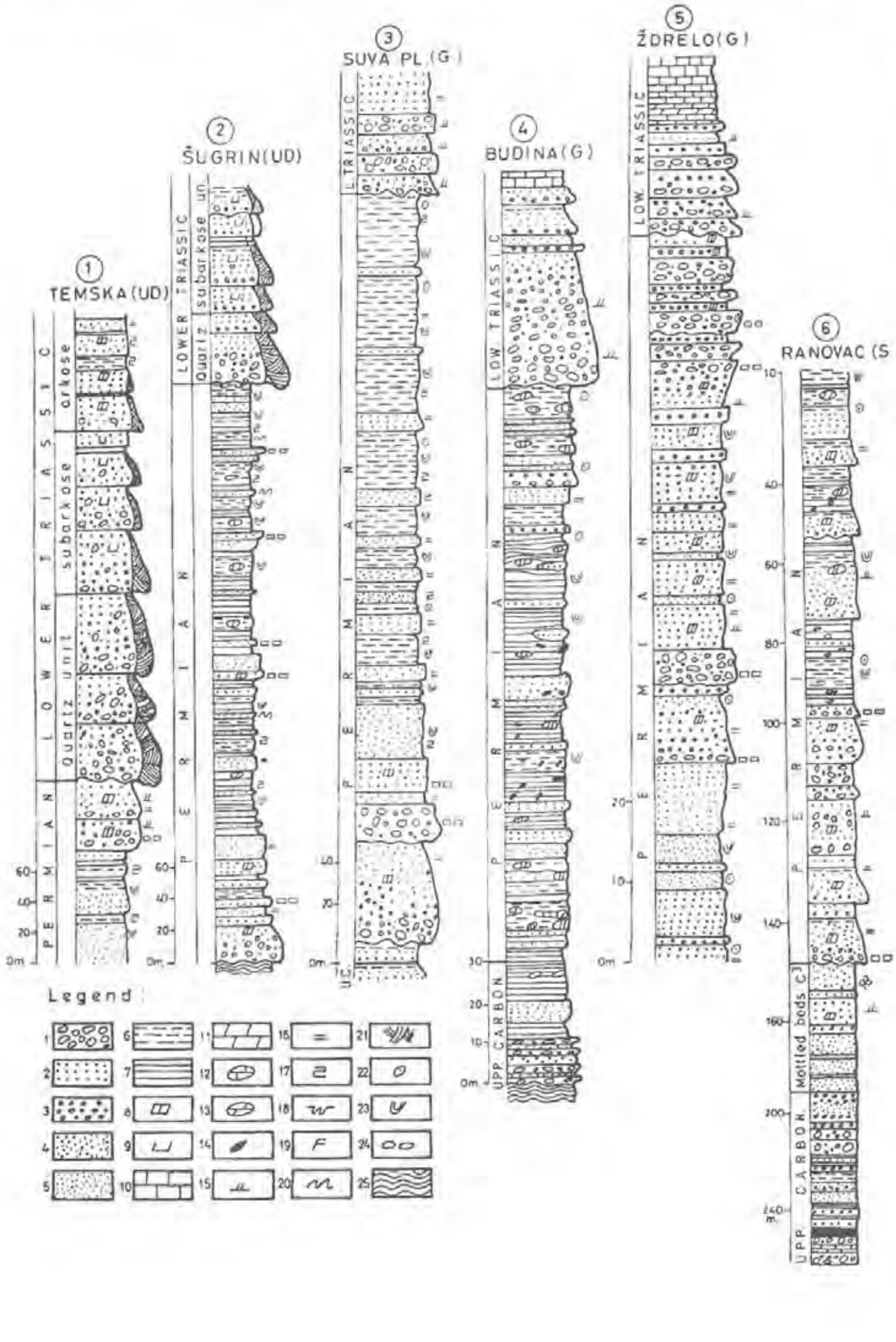

Fig. 2

Legend: 1. Conglomerate; 2. Coarse-grained sandstone; 3. Pebbly sandstone; 4. Medium-grained sandstone; 5. Finegrained sandstone; 6. Siltstone; 7. Shale; 8. Subarkose; 9. Arkose; 10. Limestone; 11. Dolomite; 12. Concretion of limestone; 13. Concretion of dolomite; 14. Concretion of hematite; 15. Cross-bedding; 16. Horizontal bedding; 17. Horizontal lamination; 18. Convolute lamination; 19. Cross-lamination; 20. Wave marks; 21. Trough cross-bedding; 22. Trace of raindrop imprints; 23. Organic markings ("bioglyphs"); 24. Concretion of spheroidal siderite; 25. Schists.

ding is horizontal, rarely cross-bedded (unidirectional fluvial sets), and the lamination in finegrained rocks is horizontal, cross-laminated or convolute. These are fining-upwards sequences from meandering rivers and bars, with gravel lag and cross-bedded sandstones. Permian rocks in this zone are transgressively overlain by Middle Jurassic rocks.

Rather thin conglomerates, locally with cobbles, and coarse-grained sandstones in the Donja Mutnica-Rujevica zone, associated with the lower part of the Permian stratigraphic sequence, correspond to scanty alluvial fans (sometimes with debris-flow sediments). These rocks pass upwards to thin river-channel deposits and further into quite widespread flood plain and playa deposits with dolomite concretions and scarce beds.

The Upper Permian sequence is well developed in the northern part of the Unit, *i.e.* in the Zdrelo and Krepoljin region (Fig. 2). It is characterised by a succession of ortho- and paraconglomerates, and of coarse-grained and rarely medium-grained sandstones. Rocks are stratified, graded with distinctly erosional lower surfaces, or else form large homogeneous masses. They show gently inclined tabular and asymptotic bedding, through to cross-bedding and horizontal bedding. The sequence is either fining or coarsening upwards. The coarsening sequence ranges from a quite thick succession of medium- and finegrained sandstones to ortho- and paraconglomerates and coarse sandstones (progradation of proximal over distal sediments). The fining sequence ranges from orthoconglomerate to sandstone with scattered pebbles, and tabular cross-beds gently inclined to horizontally bedded medium-grained sandstone. Besides fluvial transport, the sediments also indicate eolian transport desert-polished dreikanter pebbles.

These rocks are interpreted as deposits of an alluvial fan proximal to mid-slope, and partly of a braided river with bars. The imbrication, cross-bedding, and often erosional base suggest traction currents that carried pebbles as the bed load.

The Ranovac-Vlasina-Osogovo Unit (Supragethicum)

This Unit is part of the western Carpatho-Balkanides of eastern Serbia (Figs 1, 2), where unmetamorphosed Upper Carboniferous ("Stephanian") rocks lie unconformably over metamorphosed pre-existing rocks. The Permian system (Ranovac Formation) is conformable over "Stephanian" coal-bearing rocks (Fig. 2). Stephanian deposits, over the so-called "Mottled Series" (alternating grey, green, and red sandstones), 30 to 100 m thick, gradually pass into a formation of Permian red beds. The Carboniferous/Permian boundary is marked by beds with *Callipteris conferta*. Permian rocks are best studied in the Ranovac area (Maslarevic, 1961).

The Lower Permian-Rotliegend sequence is represented by conglomerates and breccia-conglomerates, coarse- to finegrained sandstones, and pyroclastic breccia of andesite-dacite type (Fig. 2). The upper part of the sequence includes siltstone and shale, and scanty thin limestone and dolomite beds. Rocks are stratified, thick or massive, exhibiting horizontal and cross-bedding produced by unidirectional flow, wavy bedding, and trough cross-lamination. Bed contacts are often sharp and erosional at the base of the coarser bed.

These rocks are channel with bar deposits in a fining-upward sequence. The upper part of the sequence consists of alluvial plain, flood plain, lake, and playa suspension deposits, composed dominantly of fine- to medium-grained sandstones, siltstones and shales with structures similar to those of the Lower Permian. The siltstones and shales are thinly bedded. Limestone concretions, raindrop prints, rarely desiccation cracks and bleaching around small plant remains are found throughout the sequence. Thin limestone and dolomite beds are scanty.

To the south, at Plazane near Despotovac, the Upper Carboniferous is conformably overlain by a Lower Permian unit of coarse- to finegrained sandstones, finegrained conglomerates, and scantier siltstone.

Permian clastics of the Gethian and Supragethian regions are immature, poorly sorted and subangular with a significant unstable component and a clay matrix. These are mainly arkoses or impure arkoses, or feldspathic greywackes and feldspathic subgreywackes (*sensu* Folk, 1954) as in the Budina brook. Arkoses and impure arkoses consist of quartz, feldspar, albite, oligoclase, andesine, potassium feldspar, rare mica, abundant granitic rock fragments (which contain purple feldspar in the Ranovac Formation), some quartzite, low-crystallinity schists, green schist (Ranovac), rarely gneiss, Lower Palaeozoic sandstone, diabase, tuff, keratophyre, lydite, andesitic and rhyolitic rocks, amphibole schist, and common fragments from the same sequences. Schist fragments are common in the Donja Mutnica-Rujevica area. Sedimentary rocks are clast- to matrix-supported, clasts being bound by a mixture of illite, montmorillonite, occasionally kaolinite in calcite, (secondary) quartz, and rarely dolomite cement.

Permian rocks are rich in heavy minerals, only some of which are important for the correlation. Dominant in older Permian rocks are the stable heavy minerals: tourmaline, zircon and rutile, occasionally also magnetite (and ilmenite), and garnet in younger beds.

Authigenic minerals include hematite, barite, pyrite, rarely gypsum, and anhydrite.

PROVENANCE AND PALEOTRANSPORT

The mineral compositions of rocks, rock fragments, and the directions of paleotransport indicate the pre-Hercinian Serbo-Macedonian Massif, and partly the Hercinian land of the Kucaj and Homolje Mts (Lower Paleozoic sedimentary and volcanic rocks), as the provenance of the deposited materials. The angularity of clasts in the Permian sediments of the Supragethicum suggests short transport distances, and those of the Ranovac Formation as deriving from underlying rocks of Permian/Carboniferous and the granite massif of Neresnica and Brnjica. In the northern and southern parts of the Kucaj Unit, paleotransport was northward from the south and SSW-NNE, respectively. The paleotransport and dispersal for the Ranovac Formation was dominantly westward (from Hercynian land).

PALAEOMAGNETISM

Paleomagnetic data for the Permian beds of the Carpatho-Balkanides (Milicevic, 1998) locate the primary depositional area at about latitude 8°N. The general magnetisation direction is $D = 350°$, $I = 5\text{-}30°$. Coordinates of the palaeopole are: latitude $\varphi = 52{,}6°N$, longitude $\lambda = 227°E$ (radius of confidence circle $\alpha_{95} = 13{,}4°$).

IMPLEMENTATIONS OF CLIMATE AND TECTONISM

The climate prevailing in the Permian was arid or semiarid, with storm rainfalls and flash floods. As a result of high rates of evaporation, basins in arid areas were alkalic, and saw the formation of concretions and thin beds of calcite and dolomite, with occurrences of barite, gypsum, and anhydrite. The clay component contains 3000 ppm strontium, indicating increased salinity in the basins. The cementation of syntaxial quartz is due to precipitation of silica from alkaline ground-water. Subaerial features such as raindrop prints are common, with occasional desiccation cracks. The environment was oxidising, intermittently reducing (grey sandstone), with conditions favouring the presence of authigenic pyrite.

Deposition of the thick Permian sequence can be associated with the Hercynian post-orogeny that formed strong relief and contributed to rapid erosion, short transport distances and, often, rapid deposition in intermontane depressions. The coarsening-upwards sequences indicate local uplift of the source areas and depositional slopes during tectonic events. Very thick fining-upwards megasequences might result from gradual decreases in basin-margin faulting. Both fining- and coarsening-upwards sequences were subjected to less violent tectonic fluctuations. Alluvial fans are often associated with normal faults and the evolution of grabens and subgrabens. Thick, finegrained deposits (interchannel, flood plain, etc.) are generated from suspension in stable, subsiding tectonic situations (Reineck & Singh, 1973).

LOWER TRIASSIC CONTINENTAL CLASTICS

Continental deposits of the Lower Triassic are found in the Stara Planina - Porec and Kucaj Units. They are best developed at Stara Planina Mountain, where they are identified as the Temska Formation (Maslarevic & Cendic, 1995).

Sedimentary rocks of this formation lie unconformably over various units of the Permian red beds at a low angle (about 12°) and over Lower Palaeozoic schists. The formation is dated using fossil macroflora: *Equisetites mougeoti, Schisoneura paradoxa, ?Neuropteridium intermedium, Woltzia heterophylla, Yuccites sp.*, and so on, characteristic of the Lower Triassic (Pantic & Protic, 1960). The formation is divided into three members: the lower quartzite, the middle subarkose, and the upper arkose.

The Temska Formation marks a new megasequence of fluvial sedimentation that began in the Permian. It is equivalent to gravel-sand and sand deposits from braided rivers, and at the beginning possibly also from the middle part of an alluvial fan. The contributing rivers were mostly large with lower flow regimes, and sometimes with slight bends. Younger finegrained constituents of the arkosic member correspond to meandering rivers. The major facies are channels with bars (longitudinal and tranverse), and interchannel finegrained facies (flood plain, natural level, crevasse, and crevasse splay) in younger beds. Sediments form beds 10-50 cm thick with diverse structures (horizontal bedding, variable cross-bedding: through, tabular, planar, asymptotic, and in finegrained rocks: horizontal and cross-laminations and convolution, showing bioturbation and raindrop prints). Sedimentary material was carried by traction currents, seldom being suspensed load.

The formation is made up of quartz and subarkose conglomerates (organised, rarely disorganised) and sandstones, then arkoses, siltstones, and shales mainly in the youngest member. Basal conglomerates often contain large (50 x 40 cm) fragments of Permian red siltstone. Sediments on the whole fine upwards in several similar sequences. These are supermature and mature clastics, with the textural maturity decreasing upwards to the younger beds. Clastics are characterised by mineralogical maturity (dominant quartz and stable heavy minerals: tourmaline, zircon, rutile), comprising rounded grains without or with a low clay matrix. The maturity is a result of prolonged fluvial transport and effective redeposition. The sediments are clast-supported, cemented by quartz and hematite, and in upper layers have a hydro-mica matrix with calcite and hematite cement. Depending on the amount of hematite cement, the rocks are red, purple or white in colour.

Lower Triassic sediments were deposited in semi-arid or arid climate with storm episodes of heavy rain. The paleotransport analysis shows dispersion in two directions, to the SE and the SW. The formation exceeds 400 metres in thickness. It is overlain by Lower Triassic shallow marine sediments of the Zavoj Formation.

Lower Triassic continental sediments in the Gethian domain do not lie over Permian rocks in all Permian localities. In many places, the overlying deposits are Middle Jurassic. Lower Triassic sequences are found in Ruj Mt., Suva Planina Mt., the Budina brook near Aleksinac, in Zdrelo and Krepoljin, where they unconformably overlie (at a low angle) various Permian members. Lower Triassic rocks are represented by basal quartz conglomerates and white or pink sandstones that pass upwards into a subarkosic rock member. Upper parts of the Lower Triassic sequence contain fossil macroflora. The sequence fines upwards (conglomerate to coarse- and medium-grained sandstone) and has features characteristic of gravel- and sand-dominated braided rivers. It resembles the Temska Formation. Continental clastics are overlain by Lower Triassic shallow-marine limestones and dolomites.

PERMIAN / TRIASSIC BOUNDARY

The stratigraphic boundary between the Permian and Triassic Systems is not recognised everywhere, but is inferred mainly from sedimentological features:

- Permian rocks are mainly red, while Lower Triassic rocks are purple, white or red in colour.
- Permian rocks are immature, poorly sorted with subangular grains, dominantly arkosic in type, with many unstable constituents such as feldspar and granitic rock fragments. Lower Triassic clastics are mature; supermature quartz conglomerates and subarkose sandstones have well-rounded or rounded grains and are well-sorted. There is evidence of humidisation during the Early Triassic.
- Grains of Permian rocks are set in a clay matrix and cemented by hematite and calcite. Lower Triassic rocks are cemented by quartz with hematite, rarely with clayey material.
- Dominant heavy minerals are garnet, sometimes also magnetite, in the uppermost Permian, but stable tourmaline, zircon and rutile in the Lower Triassic.
- Upper Permian rocks are fossil-free, and Lower Triassic rocks contain fossil macroflora.

Permian and Lower Triassic rocks were the products of two separate megacycles of fluvial sedimentation. Permian clastics were deposited mainly in intermontane depressions, in unstable tectonic situations, in alluvial fans and meandering rivers; they are thick alluvial plain, flood plain, and shallow lake deposits. Lower Triassic rocks are braided river products with diverse sedimentary structures, deposited in a lower relief environment with more stable tectonics. The Permian/Lower Triassic stratigraphic boundary consists of Lower Triassic basal quartz conglomerates, similar to western Bulgaria and Romania. The Lower Triassic rocks overlie the Permian at a low-angle unconformity. The boundary layer surface is often erosional, with cobbles in erosional pockets.

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UPPER CARBONIFEROUS AND PERMIAN CONTINENTAL DEPOSITS OF BULGARIA AND ITALY: A REVIEW

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Key words – Late Carboniferous; Permian; continental and marine deposits; geological events; Bulgaria; northern Italy.

Abstract (extended) – We give a short but updated synthesis of some Upper Paleozoic continental, sedimentary and volcanic, and locally also marine successions in Bulgaria and Italy.

– *Bulgaria*. Following the Variscan collision, the assembled and varied terranes of the Bulgarian basement generally shared the same Late Paleozoic evolution. A basin-and-swell topography occurred locally from the Early Carboniferous, or almost so, and spread gradually throughout the present territory. Namurian, Westphalian and Stephanian fossiliferous, coarse- to fine detrital sediments have been recognised, at intervals with calcalkaline intermediate-to-acidic volcanic products. Extensional and transtensional tectonic regimes seemingly began in the "Stephanian".

The Lower Permian is characterised by the alluvial-to-lacustrine clastic German Rotliegend, of which the lower part also consists, in places, of very thick igneous, mainly volcanic deposits. In contrast, the upper Rotliegend (locally interpreted as subsequent to the Saalian phase) seems to be marked by the disappearance of this effusive activity, and by the development of red beds. During the Late Permian, these red beds spread laterally and vertically, giving rise to a marked unconformity with the older rocks.

Towards the east, near the Black Sea, sabkha and marine fossiliferous sediments also occur. In Strandzha, however, the shallow-water carbonates of Kondolovo have been ascribed so far to Early Permian times (due to the presence of algae such as *Epimastopora piae*, *E. alpina*, etc.); in contrast, in the eastern Rodhope, an Upper Jurassic-Lower Cretaceous olistostrome highlighted, in the clasts, some Late Permian foraminifers. But the present position of these outcrops is still a matter of controversy; according to some authors, they are interpreted as resulting from tectonic movements from southern undefined sectors.

The boundary between the Permian succession and the overlying Lower Triassic Buntsandstein is sealed by a very important unconformity, which is marked by a gap. From a primary position slightly above the P/T boundary, the latter unit stepped down unconformably, by erosion, on to all the older rocks, including the Variscan basement.

– *Italy*. The Southern Alps, western Ligurian Alps, and Tuscany, thanks to a large number of studies, display the best examples of Late Carboniferous and Permian interregional correlation with continental areas. The South-Alpine domain clearly shows two main well-differentiated tectonosedimentary cycles, separated by a marked unconformity and a gap of as yet unknown duration. The lower one (1), as much as 2000 m thick, consists of acidic to intermediate volcanics and fluvial to lacustrine sediments, both infilling intramontane fault-bounded subsiding basins delimited by metamorphic and igneous structural highs. Normally, these deposits range from early to late Early Permian, but in some places (*e.g.* in the Tregiovo Basin) they also rise up to early Late Permian. However, as in the "Lake Volcanic District" (western Lombardy-Canton Ticino), the onset of this first cycle could also be ascribed to older times. In fact, it is noteworthy that the local Logone, Manno and other "aporphyric" and non-metamorphic molasses are interpreted as pertaining to the "Westphalian". In contrast with the above continental conditions, in the Carnic Alps, where development clearly took place from the Middle Carboniferous (late Moscovian) up to latest Artinskian or slightly younger (Bolorian), the cycle in question is made up only of shallow-marine and transitional deposits. The Lower Permian basins were controlled by strike-slip tectonics accompanied by a progressive thinning of the Variscan crust.

The Upper Permian cycle is characterised by the continental clastic Val Gardena/Verrucano Lombardo Fms and, east of the Adige Valley, by the coastal to shallow-marine Bellerophon Fm.; both units are in turn followed by the Lower Triassic Werfen/Servino deposits, which mark an overall transgression on the Alpine domain. This cycle was widespread outside the previous Permian basins, and stepped down unconformably on to the older rocks, to the Variscan basement.

In the Ligurian Alps also, two second-order cycles have been generally recognised, from Namurian-Westphalian to Early Triassic times. The older cycle consists of siliciclastics and volcanics; the latter has been subdivided into three episodes, of which the main one is related to the Early Permian. On the top, through a distinct unconformity, the Upper Permian cycle consists of Verrucano clastics.

The Upper Paleozoic of Tuscany displays continental and marine environments. The former, represented by siliciclastic sediments and volcanic products, are mostly developed in northern areas, while the latter show a more widespread distribution along the Farma Creek, near Siena. However, the age assessment of all these deposits is still debated. In general, paleontological evidence seems to assign the onset of the northern continental sedimentation to "Westphalian" times, stretching up to the Early Permian (Artinskian?). The shallow-marine sedimentation of the southern sector probably began in the late Early Carboniferous, but certainly developed up to latest Carboniferous times; the additional presence of Lower Permian clasts within the Triassic Verrucano, and of early Late Permian foram-bearing strata in the

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M. Amiata drill-cores, seem to indicate that marine conditions persisted locally throughout the Permian. However, further research is still required in these Tuscan areas.
In conclusion, from the above overview, we can identify some striking affinities between the Late Paleozoic evolution of Bulgaria and northern Italy. Consequently, we need to emphasise the validity of some events in a wider geological context.

Parole chiave – Carbonifero Superiore; Permiano; depositi continentali e marini; eventi geologici; Bulgaria; Italia settentrionale.

Riassunto (esteso) – È data una breve, ma aggiornata sintesi di alcune successioni continentali, sedimentarie e vulcaniche, nonché localmente anche marine, relative al Paleozoico superiore della Bulgaria e dell'Italia.
– *Bulgaria*. Dopo la collisione varisica, i basamenti bulgari in cui affiorano condivisero di norma un'evoluzione tardo-paleozoica piuttosto simile. Bassi ed alti strutturali si attuarono localmente in concomitanza, o quasi, al Carbonifero inferiore, e portarono a condizioni topografiche che si estesero progressivamente a gran parte del territorio. All'inizio s'ingenerarono sedimenti clastici da fini a grossolani, con fossili attribuibili al Namuriano, Westfaliano e Stefaniano, e con saltuarie intercalazioni di prodotti vulcanici, ritenuti per lo più calcalcalini, a composizione da acida a intermedia. Questi depositi vulcano-sedimentari presero probabilmente posto, a partire dallo "Stefaniano", nell'ambito di un regime tettonico estensionale/transtensivo.
Il Permiano inferiore è caratterizzato da facies ricollegabili al Rotliegende tedesco, cioè da clastiti per lo più alluvio-lacustri, che includono nella porzione inferiore masse ignee, anche cospicue, di origine prevalentemente vulcanica. Viceversa, il Rotliegende superiore (che è considerato, in Bulgaria, posteriore alla cosiddetta fase tettonica Saaliana) risulterebbe contraddistinto dalla scomparsa di questa attività estrusiva e da una sedimentazione a *red-beds*. Nel corso del Permiano superiore, questi *red-beds* si svilupparono ampiamente nella regione, raggiungendo anche notevoli spessori, e dando origine ad una marcata discontinuità stratigrafica con i preesistenti depositi.
Verso est, in prossimità del Mar Nero, sono inoltre presenti sedimenti fossiliferi evaporitico-marini. In Strandzha, i carbonati di mare basso posti in corrispondenza del villaggio di Kondolovo sono stati ascritti al Permiano inferiore (per la presenza di alghe quali *Epimastopora piae*, *E. alpina*, ecc.), mentre nella parte orientale dei M.ti Rodope un olistostroma riferito al Giurassico superiore e al Cretaceo inferiore ha evidenziato, all'interno dei clasti, la presenza di alcuni foraminiferi tardo-permiani. Tuttavia, l'attuale posizione di questi affioramenti marini è ancora oggetto di controversie; secondo alcuni autori, infatti, essi sarebbero da riferire ad unità tettoniche provenienti da settori meridionali, non meglio precisati.
Il limite tra le successioni permiane e il sovrastante Buntsandstein del Trias inferiore è sigillato da una discontinuità stratigrafica anch'essa assai significativa, in quanto include una lacuna di varie proporzioni; da una posizione stratigrafica che può essere ritenuta inizialmente più o meno prossima al limite P/T, il Buntsandstein viene tuttavia spesso a contatto, a seguito di una nutrita attività d'erosione permotriassica, con rocce relativamente più antiche, sino a raggiungere il basamento coinvolto nei movimenti collisionali varisici.
–*Italia*. Le Alpi Meridionali, le Alpi Liguri occidentali e la Toscana, grazie a innumerevoli studi, offrono ottimi esempi per una correlazione interregionale tardo-carbonifera e permiana tra aree prevalentemente continentali. Il dominio sudalpino è rappresentato da due maggiori, ben differenziati cicli tettono-sedimentari, separati da una decisa discontinuità stratigrafica e una lacuna di durata ancora sconosciuta. Il ciclo inferiore (1), potente fino a 2000 m, consiste di vulcaniti acide-intermedie e di sedimenti fluvio-lacustri, entrambi localmente accumulatisi in bacini intramontani subsidenti e delimitati da faglie, che li affiancano ad alti strutturali metamorfici e ignei. Normalmente questi depositi appartengono al Permiano inferiore, ma in talune aree (ad es. a Tregiovo) essi includerebbero anche parte del Permiano superiore. Tuttavia, come nel "Distretto Vulcanico dei Laghi" (Lombardia occidentale-Canton Ticino), l'inizio di questo primo ciclo potrebbe essere anche ascritto a tempi precedenti. Infatti, i depositi molassici a piante di Logone, Manno e di altre vicine località, che poggiano sul basamento metamorfico varisico, sono riferiti (almeno nelle prime due citate località) al "Westfaliano". In contrasto con le suddette condizioni ambientali, nelle Alpi Carniche il primo ciclo è rappresentato unicamente da depositi di mare basso e di transizione ad altri ambienti, sviluppatisi tra il Carbonifero medio (Moscoviano superiore) e l'Artinskiano sommitale, o in tempi leggermente più recenti (Boloriano). Secondo alcuni autori, la sedimentazione nei suddetti bacini sarebbe stata controllata da una tettonica a *strike-slip*, concomitante ad un progressivo assottigliamento della crosta ispessita dall'orogenesi varisica.
Il ciclo relativo al Permiano superiore è caratterizzato dalle formazioni clastiche del Verrucano Lombardo e/o dell'Arenaria di Val Gardena, che, ad est della Val d'Adige, s'interdigita ed è sovrapposta dalla Formazione a Bellerophon, di ambiente costiero e marino. Tutte queste unità sono a loro volta ricoperte dai depositi triassico-inferiori delle formazioni di Werfen e del Servino, che segnano l'inizio di una generale trasgressione marina sul dominio alpino. I prodotti di questo secondo ciclo si estesero al di là dei limiti dei precedenti bacini permiani, sovrapponendosi in discordanza su rocce relativamente più antiche, sino a raggiungere il basamento.
Nelle Alpi Liguri sono stati analogamente riconosciuti, tra il Namuriano-Westfaliano e il Trias inferiore, due maggiori cicli. Il primo tra essi consiste di depositi silicoclastici e vulcanici, quest'ultimi suddivisi in tre episodi, di cui l'episodio maggiore è quello più recente, assegnato di norma al Permiano inferiore. Più sopra, il ciclo pertinente al Permiano superiore, che è delimitato da una marcata discontinuità stratigrafica col sottostante ciclo, è caratterizzato anch'esso dal Verrucano clastico.
Il Paleozoico superiore della Toscana è contraddistinto da ambienti continentali e marini. I primi, costituiti da sedimenti silicoclastici e prodotti vulcanici, sono soprattutto evidenti nelle aree settentrionali, mentre i secondi affiorano essenzialmente lungo il Torrente Farma, presso Siena. Tuttavia, l'inquadramento cronostratigrafico di tutti questi depositi è ancora dibattuto. In genere, le evidenze paleontologiche porterebbero ad assegnare l'inizio della sedimentazione continentale al "Westfaliano", ed a farla proseguire, anche se interessata plausimibilmente da discontinuità stratigrafiche, sino al Permiano inferiore (Artinskiano?). La sedimentazione di mare basso nella Toscana meridionale probabilmente iniziò nel tardo Carbonifero inferiore, ma certamente si sviluppò sino al Carbonifero più tardo; l'ulteriore presenza di clasti del Permiano inferiore all'interno del Verrucano triassico, così come il rinvenimento di foraminiferi attribuibili alla base del Permiano superiore nei sondaggi del M. Amiata, sembrano inoltre voler indicare che le condizioni marine persistettero localmente, almeno in parte, anche nel corso del Per-

miano. Tuttavia, da questo incerto e discontinuo quadro cronostratigrafico emerge chiaramente la necessità di attuare più accurate ricerche nella successione toscana indagata.
In conclusione, in base alla rassegna svolta, è possibile riconoscere alcune strette affinità tra l'evoluzione tardo-paleozoica della Bulgaria e dell'Italia settentrionale. Ciò, pertanto, ci induce a verificarne il grado di validità di alcuni eventi riscontrati nell'ambito di un contesto geologico più ampio.

The paper gives a short but updated outline of the Upper Carboniferous and Permian continental, sedimentary and magmatic rocks in Bulgaria and Italy. These deposits are widespread in the former country, while in the latter they crop out only in the Alps, in some parts of Tuscany, and, much further south, in Calabria and northeast Sicily. Sardinia is also exclusively made up of continental deposits; however, as the evolution of this island and of the "Calabro-Peloritan arc" pertains to far western paleogeographical domains, the omission of both these regions from the present review appears justified in order to understand better the most suitable Bulgarian and Italian continental successions, which is the aim of this work.

BULGARIA

The Basement

The Variscan basement of Bulgaria has been interpreted as a collage of two large and considerably different tectonic blocks: the Protomoesian and the Thracian microcontinents. Protomoesia consists of two different parts: a nucleus, *i.e.* the Moesia terrane (with a Precambrian metamorphic substratum and an incomplete Paleozoic sedimentary cover), and a southwestern outer zone, *i.e.* the Balkane terrane (with Precambrian ophiolite and a Cambrian island-arc assemblage basement, unconformably capped again by an almost regular succession of Paleozoic sediments that differ significantly from the Moesian ones). Sedimentological, paleontological, biostratigraphical, paleobiogeographical, paleoclimatological and paleomagnetic research has supported a peri-Gondwanian origin for both these terranes (Yanev, 1990, 1993 b; Lakova, 1993; Boncheva, 1997; Haydoutov & Yanev, 1997). Generally, the Moesia microplate came into contact with the Laurussia continent at the end of the Late Devonian, whereas the Balkan terrane collided with Moesia in later times, probably between the Visean and Namurian.

The Thracian microcontinent, which corresponds to the present-day Rhodope and Serbo-Macedonian massifs, is made up of metamorphic and migmatised rocks, in which an ophiolite association again occurs (Kozhoukharova, 1984; Kolceva & Eskenazy, 1988), but with differing features from that of the Balkan complex.

We also wish to point out that the Protomoesian and Thracian superterranes were separated by a marked Variscan suture, which was in turn involved in the Alpine orogeny (Haydoutov & Yanev, 1997).

The (late) post-Variscan cover

Upper Carboniferous

At the end of the Variscan collision, the assembled and varied block fragments of the Bulgarian basement generally shared the same Late Paleozoic evolution. However, on the basis of the available dates, we can also state that the pre-Variscan plate configuration played an important role in the definition of the younger fundamental structural and sedimentary lineaments of the country. In the relatively stable Moesia, the Upper Carboniferous, found only by drilling, consists of some more or less complete and sparse clastic successions. The clearest example is represented, in the east, by the "Dobrudgea Coal Basin", where a Devonian or Lower Carboniferous basement is unconformably overlain by Namurian A (partly marine) and C sediments, up to a thick Westphalian plant-bearing clastic succession, intercalated with volcanics. Like everywhere in Moesia, there is no evidence of "Stephanian" rocks (Haydoutov & Yanev, 1997).

The Balkan Upper Carboniferous is well-developed in intramontane fault-bounded subsiding basins. The Svoge Trough includes Namurian to B/C Westphalian alluvial-lacustrine, fine to coarse detrital sediments, up to about 1700 m thick. They contain lithics of the non-metamorphic substrate and subordinately of Variscan granitoids in the upper part (Yanev, 1965). Coal layers and calcalkaline andesitic products also occur. Westphalian B is documented by the presence of *Calamites* and *Lepidophytes* (Tenchov, 1966).

In the Balkan Mts, there are also Stephanian plant-bearing clastic sediments, up to a maximum of 1000 m in thickness, which infill a number of narrow basins. In places, these deposits are associated with calcalkaline andesitic-dacitic volcaniclastics and lava flows (Yanev, 1981). Generally such igneous activity, which is greatly evidenced in some areas (Belogradcik, Berkovica, etc.), developed mainly during the Early Permian.

It is also noteworthy that the Stephanian basins of the Balkan region do not generally coincide with the slightly older ones, which appear less widespread. Normally, their sediments lie directly on a folded and metamorphic basement and pass gradually to the Permian deposits.

In the Srednogorie region, the post-Variscan succession of Mount St. Iliya shows some basal conglomerates, which have been dubiously ascribed to the Carboniferous-Permian transition (Catalov, 1985).

In Kraishte (SW Bulgaria), "Stephanian"-"Autunian" slightly metamorphosed, grey and red clastic sediments, which unconformably rest on a Lower Paleozoic or older basement, have also been recorded in the Vukovo area (Yanev, 1982).

In conclusion, from the above overview, we are led to believe that a fault-bounded basin-and-swell framework developed from the early Late Carboniferous, but most of the examples originated in slightly younger times, spreading progressively throughout Bulgaria. As in other parts of Europe, this geological setting seems to be connected to tensile tectonism, apparently lacking in marked compressional stresses. In an interregional context, the opening of the "Stephanian" basins could have been linked to transtensional movements, which were consistent with, and followed by, during the Early Permian, an extensional regime. Probably, the depicted geological evolution of the Moesia and Balkan

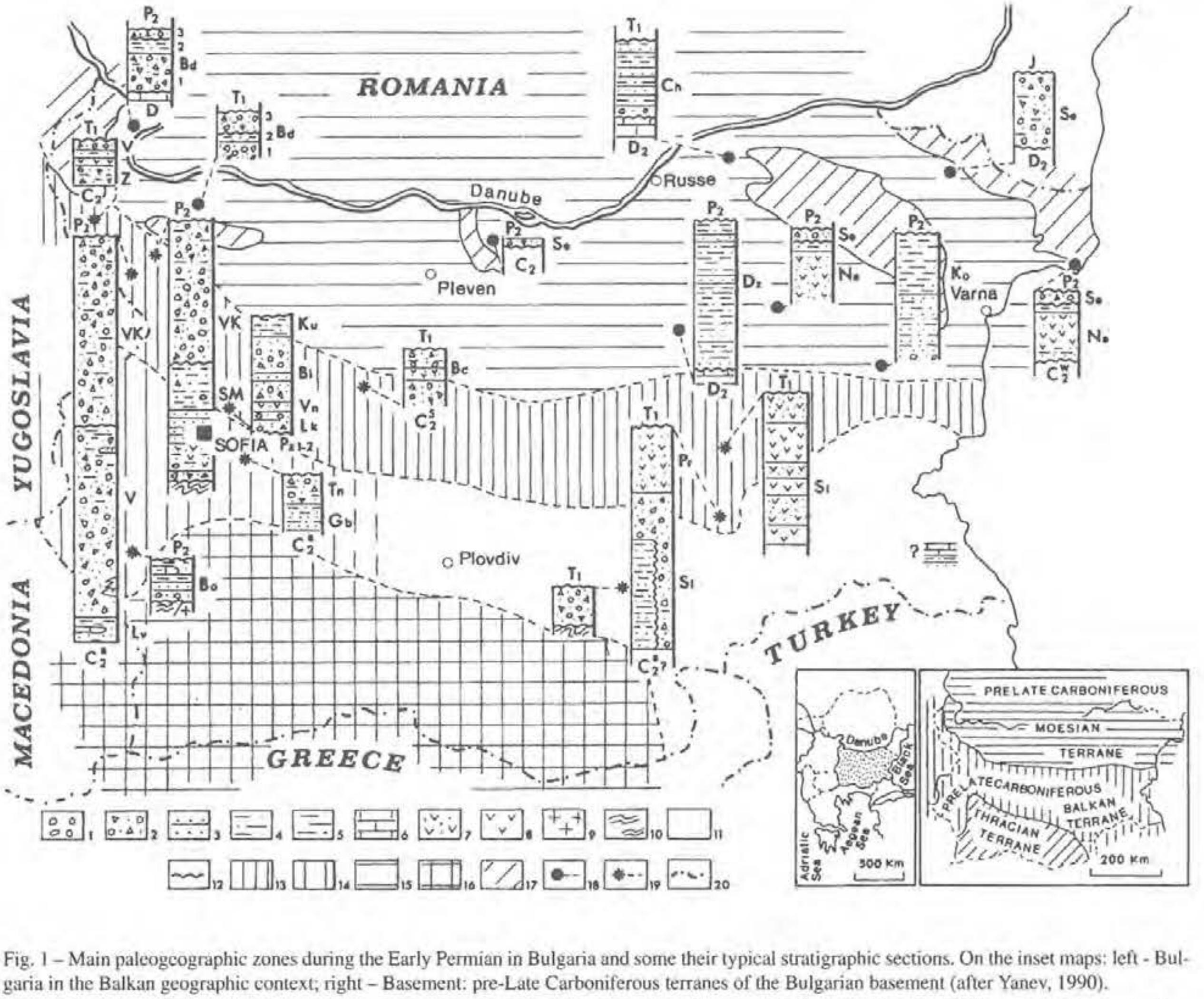

Fig. 1 – Main paleogeographic zones during the Early Permian in Bulgaria and some their typical stratigraphic sections. On the inset maps: left - Bulgaria in the Balkan geographic context; right – Basement: pre-Late Carboniferous terranes of the Bulgarian basement (after Yanev, 1990).

Legend. Lithology: 1. conglomerates; 2. breccia-conglomerates; 3. sandstones; 4. mudstones and siltstones; 5. shales; 6. limestones; 7. pyroclastics; 8. volcanics; 9.plutonics; 10. metamorphic rocks. Paleogeographic zones: 11. Areas lacking in Permian data (Meso-Cenozoic cover); 12. Erosive contact; 13. High dry land with isolated intramontane basins infilled by coarse clastic deposits; 14. Hilly terrains with isolated basins (generally sediment by-pass zone); 15. Isolated basins covered by Meso-Cenozoic deposits (from drilling data); 16. Dry land lacking in Lower Permian sediments; 17. Low dry land with Lower Permian deposits (Meso-Cenozoic cover); 18. Position of stratigraphic columns based on drilling data; 19. Position of Lower Permian sequences based on outcrop data; 20. State boundary. *Abbreviations* (in alphabetic order). – Lithostratigraphic units: Bc, Vasilyovo Fm.; Bd, Bdin Fm. (1, Bononia Mb., 2, Deleina Mb., 3, Rasovo Mb.); Bi, Birimirtzi Fm.; Bo, Boboshevo Fm.; Dz, Dalna Zlatitza Fm.; Gb, Gabra Fm.; Lk, Lokovsno Fm.; Lv, Levitza Fm.; Ko, Komunari Fm.; Ku, Kurilo Fm.; Na, Nanevo Fm.; Pr, Prohorovo Fm.; Si, Sveti Iliya Fm.; Sl, Sliven Fm.; Sm, Smolyanovtzi Fm.; Sv, Severtsi Fm.; Tn, Tarnava Fm.; V, Vranska Fm.; Vk, Vranski Kamak Fm.; Vn, Voinezhka Fm.; Z, Zelenigrad Fm. – Chronostratigraphic units: D, undivided Devonian; D_2, Upper Devonian; C_2, undivided Upper Carboniferous; C_2^w, Westphalian; C_2^s, Upper Stephanian; P_2, Upper Permian; T_1, Lower Triassic; J, Jurassic.

Upper Carboniferous outcrops could also be interpreted as inherited by the differing primary conditions of both these terranes, as well as by the weakness of their joint lines.

Permian

Throughout the Permian, a Rotliegend-type clastic sedimentation spread progressively over vast regions of Bulgaria, and was locally accompanied by significant volcanism (Fig. 1). However, according to the literature (Yanev, 1992; Yanev & Cassinis, 1998; etc.), this eruptive activity took place only during the Early Permian, and was mainly concentrated at the beginning of this epoch. Most of these examples are recorded in the Balkan areas; in particular near Sliven, in the east, this igneous activity gave rise to volcanic, subvolcanic and plutonic products. Generally, the volcanics consist of calcalkaline rhyolitic rocks, whereas the relatively deeper bodies are made up of granophyres, microgranites and granodiorites (Zhukov *et al.*, 1976).

In Moesia, Permian rocks have so far been detected only by drill-cores (Fig. 1). Some sections of the eastern area (Kaliakra, Targovishte and other places), similarly to a very large part of Bulgaria, can be divided into two sedimentary cycles. Only the first cycle (I), which coincides with the "North Bulgarian Lower Group" of Yanev (1992), is defined as typical Rotliegend, in turn subdivided by the same author into two parts, respectively indicated as "Lower Rotliegend" (P_1^1) and "Upper Rotliegend" (P_1^2), and both are generally ascribed to the Early Permian.

Along coastal Dobrudgea, the Permian begins with the sedimentary and volcanic Nanevo Fm., which is also recorded in the southern part of the same region and around the Permian "paleo-horst" of northeast Bulgaria (Fig. 1). In contrast, in these and nearby areas, and in the extreme NE part of the country, the unconformably overlain coarse detrital deposits of the Severci Fm. lack volcanic products. Generally, the fanglomerates of the two aforementioned Rotliegend-type units include Lower Paleozoic metamorphic and Devonian-Lower Carboniferous carbonate rock-fragments, deriving from the same Moesia Platform.

Locally, in the Bulgarian "Dobrudgea Coal Basin", the Lower Permian appears to be confined between two main unconformities, respectively, below, with Devonian rocks, and above with the Lower Triassic Buntsandstein or Jurassic sediments (Fig. 1).

In central and western Moesia, the Lower Permian clastic deposits are grouped as the Bdin Formation (Fig. 1). This unit spread throughout the northwestern part of the present Bulgaria, approximately from Vidin to Pleven, where it unconformably rests on a Lower Carboniferous (Visean) or Devonian substratum. On the top, near the village of Rasovo, the Bdin Fm. comes directly into unconformable contact with the Lower Triassic Buntsandstein or similar red beds.

Along the southern margin of Moesia, the Lower Permian is well developed south of the "Tarnovo Depression" and a little to the west. It consists of fanglomerates, breccias and finer clastics. These latter deposits, known as Dolna Zlatitsa Fm., pass laterally, in the eastern Vetrino area, to the coarse sediments of the Komunare Fm., which also occur near Varna (Fig. 1).

In the Balkan region, the Lower Permian is made up again of continental Rotliegend-type sediments and volcanics (Fig. 1). The latter products, which crop out in the pre-Balkan (Belogradcik, Teteven) and western and central Balkan (Berkovica, Levskj peak) areas, generally display a calcalkaline acidic composition, are irregularly distributed in the field, and can reach about 1000 m in thickness. The Rotliegend fine- to coarse-grained sediments infilled a number of fault-bounded basins, and locally onlapped outside. The lithoclasts commonly consist of metamorphic, Variscan and older intrusive rocks, as well as Permian volcanic fragments.

In the Balkan region, as in other Bulgarian areas, the presence of an unconformity between the Lower and Upper Rotliegend is probably related to the disappearance of the Early Permian magmatic activity. Furthermore, in some places, the latter unit is missing and the Lower Rotliegend comes directly into contact with the Lower Triassic Buntsandstein (Petrohan Group). In the Svoge area, the Permian was probably never deposited.

In Srednogorie, just above a Lower Palaeozoic and older basement, the Rotliegend of Mount St. Iliya, which consists of coarse- to fine clastics topped by volcaniclastic products both of presumed Early Permian age (Fig. 1), is unconformably capped by the red beds of the Upper Rotliegend.

In southeastern Bulgaria, the presence of Permian deposits is still the subject of controversy. In Strandzha, the Kondolovo area displays some algae-bearing carbonate rocks (Fig. 1), including *Epimastopora piae*, *E. alpina* and *Mizzia velebitiana*, which have been related to the Early Permian (Malyakov & Bakalova, 1978; Bakalova, 1988). However, these local marine, shallow-water sediments have also been interpreted as the result of tectonic displacement from southern sectors.

As already stated, in southwestern Bulgaria (Kraishte) a narrow fault-bounded basin near Vukovo was infilled by Upper Carboniferous ?-Lower Permian grey and red clastics, which unconformably overlie a Lower Paleozoic or older basement (Yanev, 1979; Ellenberg *et al.*, 1980; Fig. 1).

A second Permian cycle (II), corresponding to the "Lower Danube Upper Group" of Yanev (1993 a), is clearly developed in North Bulgaria, especially in the Moesia region (Fig. 2). This Group is generally represented by alluvial, partly deltaic massive clastic red beds (Targovishte Fm.), which unconformably overlie the Lower Permian or older rocks, reaching more than 1000 m in thickness. In the Provadia syncline, these deposits are affected by intercalation of evaporites and carbonate fossiliferous bodies (Vetrino Fm.; Fig. 2), which seem consistent with the presence of marine conditions towards the east, in the position of the present-day Black Sea.

Palynological data (Schirmer & Kurze, 1960; Pozemova *et al.*, 1972 in Yanev, 1993 a), such as the discovery of *Lueckisporites virrkiae*, *L. platysaccoides*, *Klausipollenites schlaubergeri* and *Falcisporites zapfei* in basal pelitic levels (Mirovo Fm.), suggest that this younger cycle (II) developed during Late Permian times. These Upper Permian deposits are more widely distributed than those of the Lower Permian cycle. In some places, however, they lack evidence.

Over a large part of Moesia, the Targovishte Fm. is overlain by well-bedded varicoloured, finegrained sediments, which yield palynomorphs (Pozemova *et al.*, 1972 in Yanev, 1993 a) related to the latest Permian (in part equivalent to the Upper Tatarian of the Russian platform). This Totleben Fm. is unconformably capped by the Lower Triassic red clastics of the Buntsandstein, which covered all the country's rocks and led to a new sedimentary cycle (Fig. 2). This regional discontinuity marks a time gap of unknown duration which, according to some authors (*e.g.* Yanev, 1981), seals the Permian-Triassic (P/T) boundary.

The influence of transitional to marine environments towards the Black Sea has also been recorded in the eastern part of the Rhodope Massif, near the Bulgarian-Greek bor-

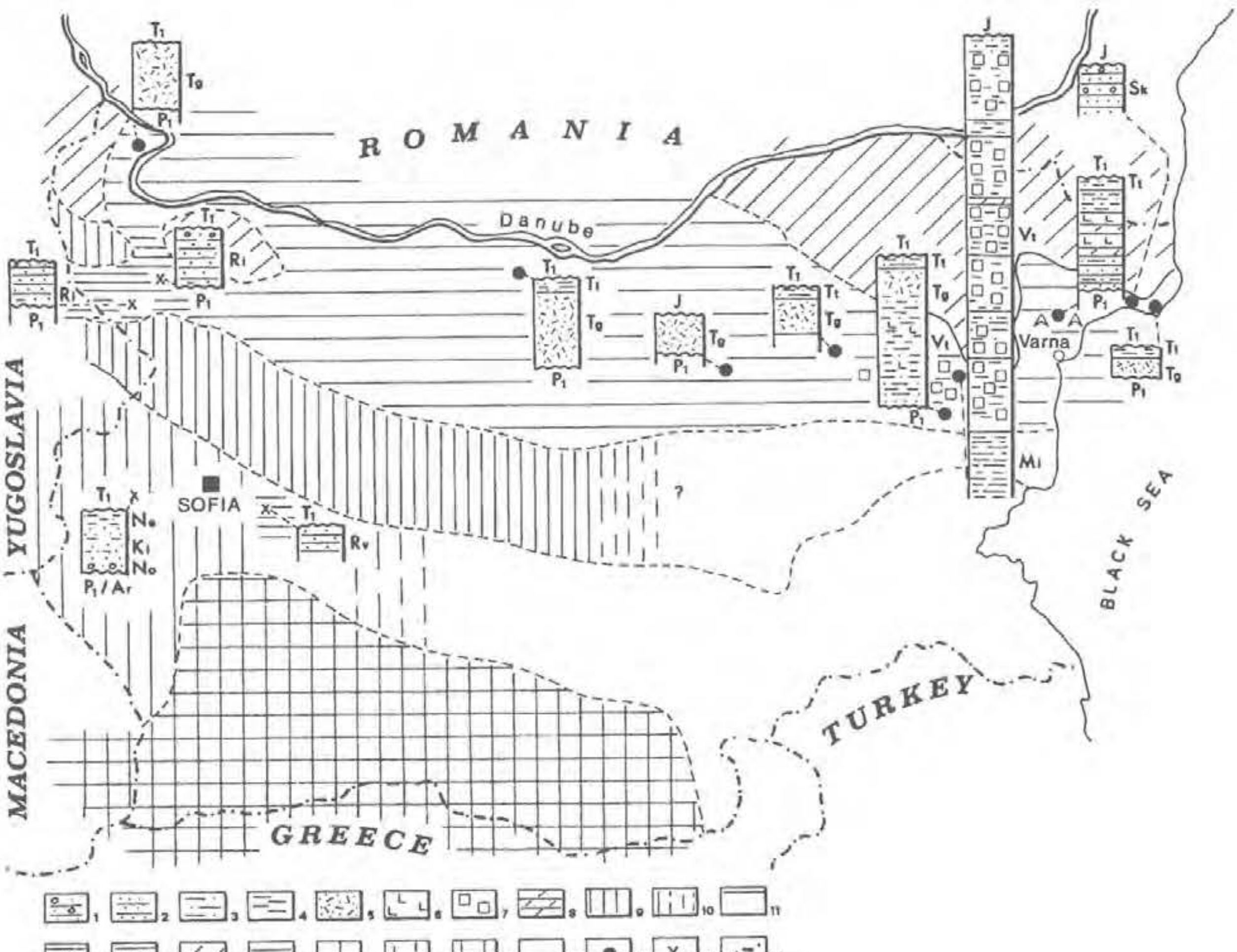

Fig. 2 – Main Late Permian paleogeographic zones of Bulgaria and some typical stratigraphic sections.

Lithology: 1. conglomerates; 2. sandstones; 3. mudstones and siltstones; 4. shales; 5. Very low sorted sediments; 6. anhydrites; 7. halites; 8. dolostones. Facial and paleogeographic zones: 9. Dry land with moderate relief; 10. Probably as 9; 11. Northern continental basin covered by Meso-Cenozoic deposits (drilling data); 12. Anhydrite-bearing zone pertaining to the same basin; 13. "Sabkha" sedimentation; 14. Lower dry land; 15. Delta connected to continental basin; 16. Southern continental basin, 17. Probably as 16; 18. Dry land lacking in Upper Permian deposits; 19. Zone without any information on the Permian sedimentation (Meso-Cenozoic cover); 20. Local section based on drilling data, 21. Local section based on outcrop data; 22. State boundary. *Abbreviations* (in alphabetic order) – Lithostratigraphic units: Ki, Kiselichka Fm.; Mi, Mirovo Fm.; Ne, Nepraznentzi Fm.; No, Noevtzi Fm.; Ri, Rinovska Fm.; Rv, Pavulya Fm.; Sk, Sokolare Fm.; Tg, Târgovishte Fm.; Tt, Totleben Fm.; Vt, Vetrino Fm. – Cronostratigraphic units: Ar, Archean; P_1, Lower Permian.

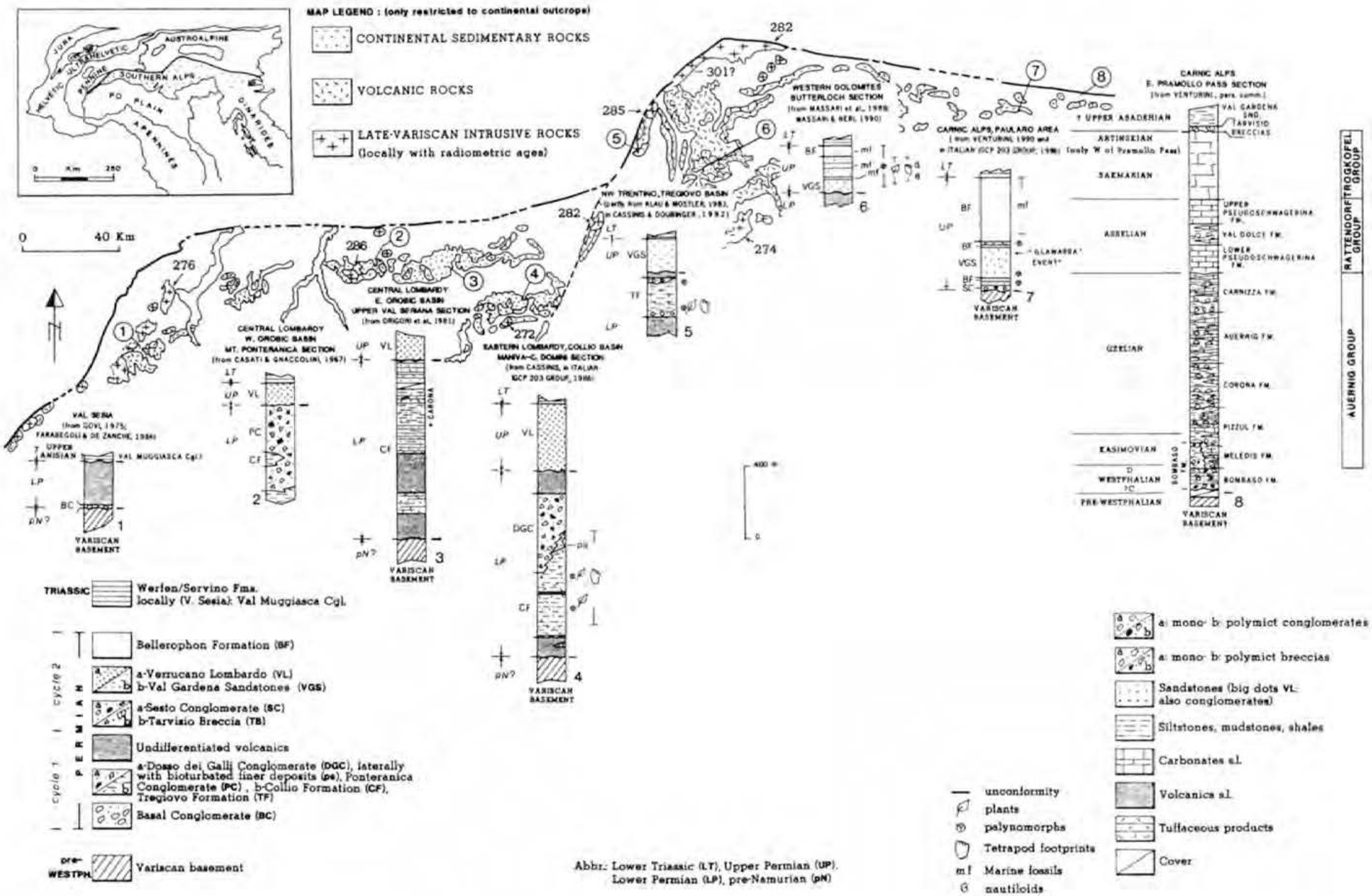

Fig. 3 – Selected and schematic Upper Carboniferous-Permian stratigraphic sections in the Southern Alps. Data from the authors cited above the columns; radiometric ages of intrusive bodies from A. Del Moro (Pisa, CNR) and G. Liborio (Milano Univ.), pers. comm.; Permian, very simplified continental map from Cassinis, unpublished. (After Cassinis, 1996).

der. Silicified carbonate rock fragments, reworked into an Upper Jurassic-Lower Cretaceous terrigenous olistostrome cropping out to the north of Dolno Lukovo, uncovered Upper Permian foraminifers (Trifonova & Boyanov, 1986), such as *Agathammina pusilla*, *Bradyina novizkiana*, *Neoendothyra parva* and *Colaniella* sp.; however, as for the Lower Permian of Strandzha, the paleogeographical-structural source-area of this unit could be envisaged to the south.

ITALY

The Alps and Apennines show a very differentiated and complex history for their generally folded and/or metamorphic substrata, but further research and careful evaluation are still necessary for a better understanding of these rock basements. Therefore, as this paper focuses on the most significant features of some Upper Paleozoic continental areas of Italy, our overview starts from this time. An inheritance of the substratum on the post-Variscan cover is, as in Bulgaria, very plausible in the light of the following data.

The Alpine Upper Paleozoic

Upper Carboniferous and Permian

In the Southern Alps, excluding Carnia, Permian sedimentary and volcanic continental rocks rest above a crystalline basement affected by a Variscan metamorphism, of which the last event can be assigned approximately to the Visean/Namurian boundary (Sudetian phase), or to slightly more recent times. The Carboniferous crops out only between Lakes Como and Maggiore, with some "aporphyric" fluviolacustrine-to-fluviopalustrine sediments bearing middle Westphalian (Venzo & Maglia, 1947; Jongmans, 1950, 1960) up to perhaps, locally, Stephanian megafloras. Intrusive bodies also occur, generally concentrated along important Alpine tectonic lines or close thereto (Fig. 3).

Therefore, from the Late Carboniferous (Moscovian), but mainly during the Early Permian, a basin-and-swell topography dominated the region. This structural framework can be interpreted as the result of a general collapse of the Variscan orogen, accompanied and followed by the outpouring of significant volumes of magma, as well as by transtensional-transpressional movements acting in a progressively extensional regime (*e.g.* Arthaud & Matte, 1977; Ziegler, 1984, 1988; Vai, 1991; Broutin *et al.*, 1994; Cassinis & Perotti, 1994, 1997; Cassinis *et al.*, 1997; Cortesogno *et al.*, 1998).

In the western and central sectors of the Southern Alps, the Lower Permian consists of continental acidic-intermediate volcanics and alluvial-lacustrine deposits (Collio and Tregiovo Fms, Ponteranica and Dosso dei Galli Cgls, etc.), both infilling intramontane fault-bounded subsiding basins delimited by metamorphic and igneous structural highs. Basal polymict conglomerates and breccias, together with fine siliciclastic products (Basal Cgl., Ponte Gardena Cgl., etc.), may be present (Fig. 3).

Paleontological data from the macroflora, palynomorphs and tetrapod footprints generally indicate an Early Permian or, locally, a slightly younger age (Haubold & Katzung, 1975; Remy & Remy, 1978; Kozur, 1980 a; Cassinis & Doubinger, 1991, 1992; Conti *et al.*, 1991, 1997; Pittau, 1999 a, b; Cassinis *et al.*, in press). Radiometric investigations also agree with the above dating (in Cassinis *et al.*, 1999 and in press).

During the Late Permian, new paleogeographical and structural conditions occurred, due to a plate reorganisation probably connected with the opening of Neotethys (Ziegler & Stampfli, this volume). In the Southern Alps, this new cycle includes the fluvial red clastics of the Verrucano Lombardo and the Val Gardena Sandstone as well as, to the east of the Adige Valley, the sulphate evaporite to shallow-marine carbonate sequences of the Bellerophon Fm. (Fig. 3). The above-mentioned red sediments, which in places are preceded by mono- or polymict ruditic units (Daone and Sesto Cgls, Tarvisio Breccia), at least in part regarded as scarp-foot fan deposits, form a widespread blanket up to 600 m thick which covers the Early Permian basins and the surrounding highs. The contact with these underlying rocks is delimited by a gap of as yet unknown duration. Moreover, the new cycle marks the extinction of the volcanic activity.

Paleontological data suggest, for this upper succession, a very Late Permian age, generally beginning from Tatarian times (*e.g.* Italian IGCP 203 Group, 1986; Massari *et al.*, 1988, 1994; Cassinis *et al.*, 1995, 1998, 1999).

On the basis of recent research, the boundary between the Upper Permian red beds and the overlying Lower Triassic Werfen/Servino Fms, in the area from central Lombardy to Slovenia, is seemingly devoid of a consistent stratigraphic gap.

In the Carnic Alps, where Variscan metamorphism is lacking, or almost so, the tectonic climax linked to this orogeny occurred during Moscovian ("Westphalian") times (*e.g.* Venturini, 1990). As a consequence, the overlying succession (Pontebba Supergroup) made up of a cyclic alternation of marine, deltaic and paralic sediments is very different from the coeval succession west of Comelico and resembles that of nearby former Yugoslavia. In contrast, the Upper Permian of the Dolomites shows continuity, even if in Carnia and Cadore the relative sections are characterised by basal clastics, and probably a different temporal shift of the formations.

In the other sectors of the Italian Alps, the only useful and updated studies on the Upper Paleozoic have been carried out in western Liguria (Vanossi, 1991; Fig. 4). In this area, the "internal Briançonnais" is represented by Namurian and younger fluvial to lacustrine clastic sediments associated with volcanic products, both infilling fault-bounded basins. As regards the volcanics, early calcalkaline rhyolitic ignimbrites are followed upwards by andesitic pyroclastics and lavas with subalkaline affinities, and, mainly during the Early Permian, by calcalkaline acidic ignimbrites and tuffs. Normally these latter products, ending with subalkaline, high-K rhyolites, represent the most important post-Variscan magmatic episode of the investigated sector, with deposits estimated at over 1000 m in thickness. The overlying Upper Permian Verrucano Fm. is marked by an erosive surface and a gap of uncertain duration. On the top, Lower-to Middle (*p.p.*) Triassic fine detrital sediments crop out. The "external Briançonnais" deposits of the Ligurian Alps are not substantially different to those of the internal sectors.

The Apennine Upper Paleozoic, up to the overlying Triassic

Upper Carboniferous, Permian and younger Triassic times. In the northern Apennines, only Tuscany includes Upper Palaeozoic outcrops. Conforming to a differing geographical distribution, they consist of continental and marine deposits (Fig. 5). As a consequence, for easier correlation with the previously described Alpine framework, we wish to initiate this regional review from the former domains, which display a wider development in the north.

Following the Variscan orogeny, of which the main tectonometamorphic event seems to be connected with the Sudetian phase, a basin-and-swell topography probably began during Moscovian (late "Westphalian"?) times and lasted up to as-yet-undefined Permian times (Remy in Rau & Tongiorgi, 1974; Vai & Francavilla, 1974; etc.). Fluvio-deltaic and lacustrine sediments accumulated in some troughs, later overstepping their boundaries and taking on a reddish colour (Rau & Tongiorgi, 1974, 1976). The onset of this event is unknown, due to a lack of paleontological data and the difficulties of stratigraphic reconstruction within a region strongly affected by the Alpine orogeny; however, a Permian post-"Autunian" time (more or less corresponding to the "Saxono-Thuringian" of the French Authors) may be reasonably assumed.

Volcanic products occurred during, and perhaps also af-

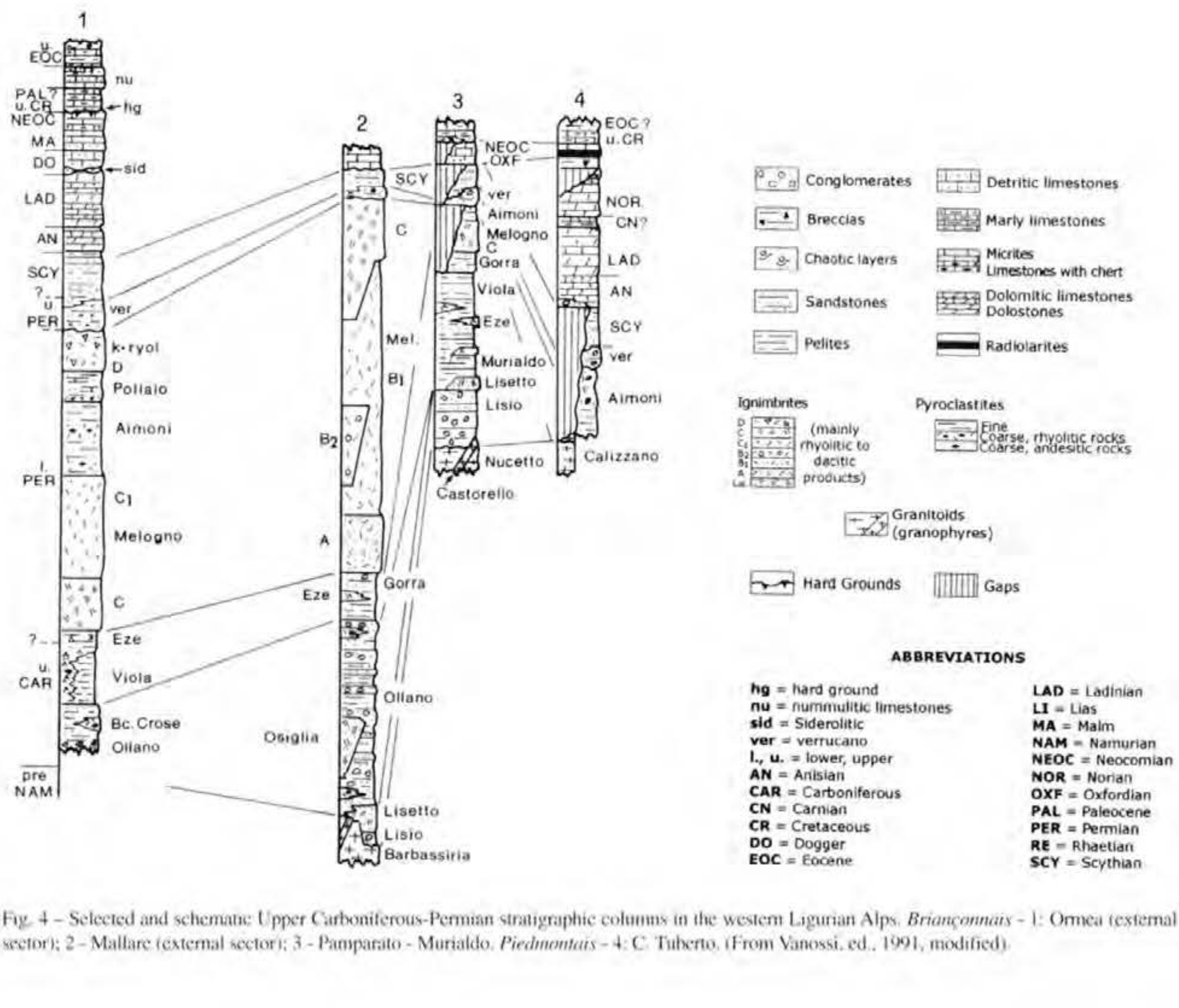

Fig. 4 – Selected and schematic Upper Carboniferous-Permian stratigraphic columns in the western Ligurian Alps. *Briançonnais* - 1: Ormea (external sector); 2 - Mallare (external sector); 3 - Pamparato - Murialdo. *Piedmontais* - 4: C. Tuberto. (From Vanossi, ed., 1991, modified).

ter, the Early Permian (*e.g.* Barberi, 1966; Bagnoli *et al.*, 1979; Costantini *et al.*, 1991, 1998; Pandeli, 1998). However, in places, the clasts deriving from some covering units (such as the Castelnuovo sandstones at Larderello) testify to a wider development of this activity. In any case, the impressive aspects of the Corsican-Sardinian plutonism and volcanism are lacking in Tuscany.

The Verrucano, of which the early sedimentation is generally related to the Middle Triassic but must be regarded as diachronous, "covers" a gap of as yet undefined duration. Between Punta Bianca and the Argentario promontory, it rests on deposits of different ages, locally laid down directly on the Variscan crystalline basement (Rau & Tongiorgi, 1972, 1976; Pandeli, in press; others).

In Tuscany, marine sediments crop out within or near the above-mentioned continental deposits in the form of thin intercalations or very thick sequences authors (Vai, 1978; Cocozza *et al.*, 1987; Costantini *et al.*, 1988). The latter are present near Siena, showing facies of shallow- and deeper water, including turbidites and olistoliths (Farma Fm.). The ages of all these marine sediments, as well as of others which are found either in the boreholes of the Amiata geothermal field or among the lithic clasts from the Monticiano-Roccastrada tectonic unit, generally extend, according to some authors (*e.g.* Costantini *et al.*, 1988; Engelbrecht *et al.*, 1989; Pasini, 1980, 1991; Pandeli & Pasini, 1990; Elter & Pandeli, 1991), from a presumed late Visean age up to the beginning of the Late Permian (Kubergandian *Cancellina* Zone). However, to be precise, this marine sedimentation seems to be discontinuous in Tuscany from the latest Carboniferous. Post-Kubergandian deposits have not been recorded so far in the country.

LATE PALEOZOIC SCENARIOS FOR BULGARIA AND ITALY: A COMPARISON

From Late Carboniferous to Permian times, the investigated areas of Bulgaria and Italy show some affinities, which can be summarised as follows.

Depositional events

Carboniferous p.p.

In both countries, post-Variscan sedimentation began in some places from late Visean or the Namurian times. It consisted of continental and local marine deposits. On the basis of paleontological data, the best examples are found in some Moesian ("Dobrudgea Coal Basin") and Balkan (Svoge) areas, in the Carnic and western Ligurian Alps, and in southern Tuscany (Farma Valley, near Siena). The earliest deposits unconformably overlie metamorphic or nonmetamorphic rocks, generally related to the Devonian/Early Carboniferous or, in places, to older times.

– Bulgaria

In this region, sedimentation spread progressively during the Late Carboniferous and Permian. Thanks again to floral and faunal investigations, "Westphalian-Stephanian" and Lower Permian siliciclastic deposits have been identified in a large number of regions, which were affected by a different evolution. Once more, Moesia and the Balkan Mts are the only Bulgarian areas where the Upper Carboniferous is widespread and well developed within intramontane fault-bounded narrow and subsiding basins. The Dobrudgea and Svoge areas represent the clearest examples. However, as already stated, there is no evidence of "Staphanian" deposits in Moesia.

– Italy

In Italy, the Upper Carboniferous is, often paleontologically, recorded in the southern and western Alps, and in the northern Apennines. Continental siliciclastic deposits bearing Westphalian and/or Stephanian plants occur in western Lombardy (Logone, etc.) and in the nearby Canton Ticino (Manno, etc.), in the Ligurian Briançonnais (associated with volcanics) and in Tuscany (Pisan Mts. and Jano). In contrast, in the Carnic Alps the onset of the Pontebba Supergroup is related to Moscovian times, and the basal, transitional to marine deposits (essentially pertaining to the Auernig Group) have yielded some indicative foraminifer assemblages, Kasimovian and Gzhelian in age (see Venturini in Cassinis *et al.*, 1998).

Permian

– Bulgaria

The Lower Permian of Bulgaria consists of, as in Italy, Rotliegend-type alluvial-lacustrine, coarse to fine clastic sediments. The basal part, which is marked by consistent volcanic products, passes upwards to more or less similar sediments, locally through an unconformity. These relatively upper deposits, in which volcanics seem missing, have so far been regarded as Lower Permian (Yanev, 1981 and other works).

Upwardly, the abundant clastic red beds of the Targovishte Fm. or of other lateral deposits unconformably cover all the preceding rocks, and locally step down on to the pre-Lower Carboniferous basement. We also wish to point out, in eastern Bulgaria, that some Upper Permian deposits of Moesia (Vetrino Fm.) containing evaporite and carbonate bodies seem to be connected with marine conditions in the position of what is now the Black Sea.

On the basis of palynological data and regional correlations (Yanev, 1993 a), the upper cycle mentioned above is ascribed to Late Permian times. Furthermore, as already stated, the presence of a new regional unconformity by the overlying Lower Triassic Buntsandstein (or Petrohan Group) supports the opinion that the P/T transition is marked by a gap of uncertain duration. The direct overlapping in some places of such Lower Triassic deposits on the Variscan basement of Bulgaria greatly increases the importance of this unconformity, from which a relatively wider sedimentary cycle began.

– Italy

As in Bulgaria, some central and western Alpine areas of Italy were also generally affected by a roughly similar evolution of the Permian succession. In fact, the underlying alluvial-lacustrine varicoloured sediments and the associated volcanic products are unconformably followed by the Verrucano and Val Gardena Sandstone fluvial red beds (Italian IGCP 203 Group, 1986; Cassinis *et al.*, 1999). In many places, the above volcanics appear widespread and well developed, so that, locally, the lower Rotliegend clastics are very subordinate or almost missing (such as in the "Bolzano-Trento porphyric plateau", the "Volcanic Lake District", and so on). Also the Ligurian Alps were not substantially different, during the Permian, from this stratigraphic framework in which the Verrucano was deposited unconformably on abundant volcanics.

According to palaeontological and radiometric data, the units underlying the Verrucano or the more or less coeval Val Gardena Sandstones have been so far related to Early Permian and locally also to slightly later times. However, some contrasting chronostratigraphical results obtained from the above research, and in part undoubtedly emphasised by the different time-scales currently in use, are still the object of controversy (Schaltegger & Brack, 1999; Cassinis *et al.*, 1999 and in press). Consequently, further investigations are necessary in order to interpret this age difference better.

In contrast, the overlying deposits of the Verrucano-Val Gardena Sandstone agree with Late Permian times. In fact, the macroflora, palynomorphs and tetrapod footprints discovered in the latter formation, which, to the east of the Adige Valley, passes laterally and upwards into the marine

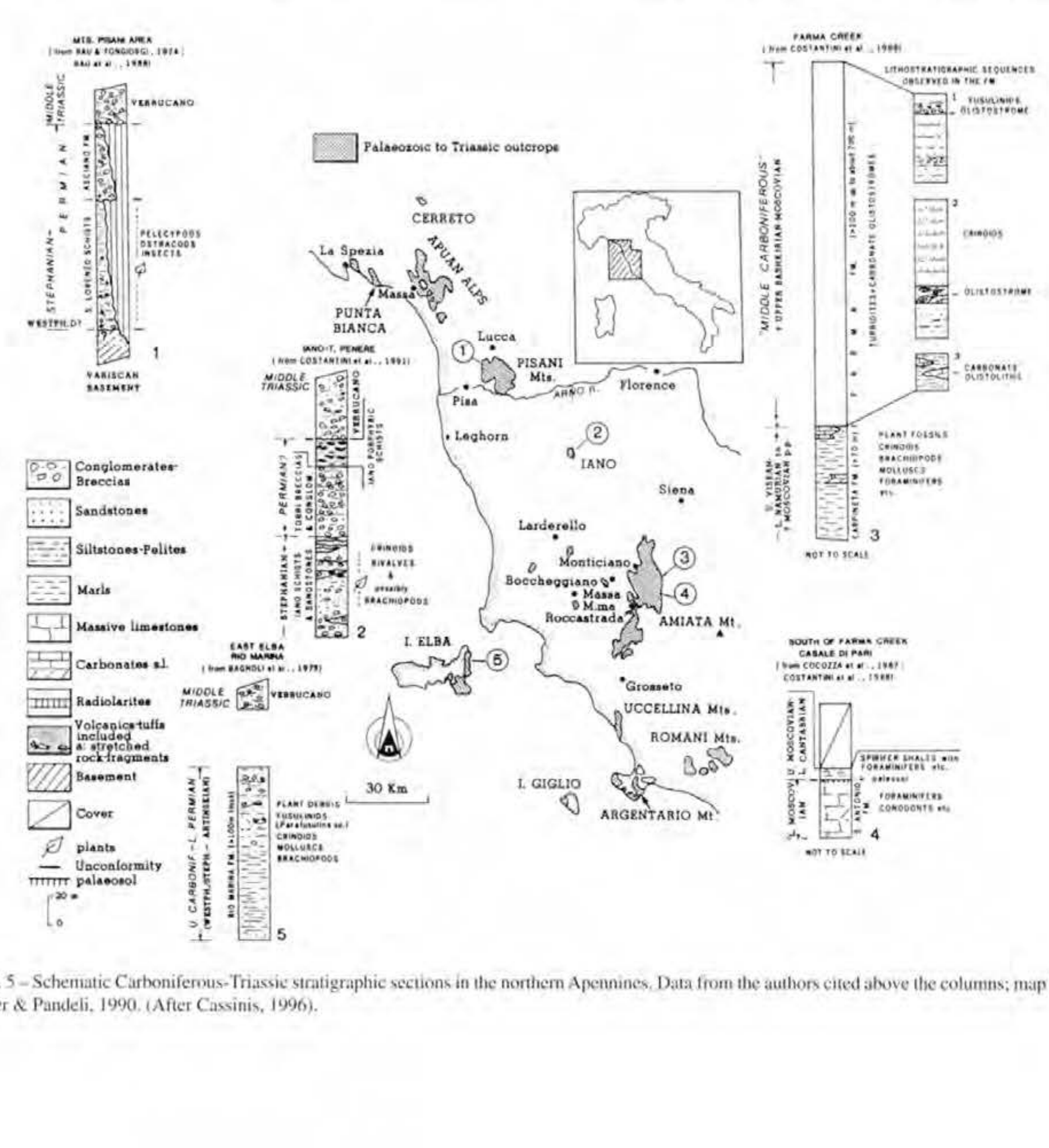

Fig. 5 – Schematic Carboniferous-Triassic stratigraphic sections in the northern Apennines. Data from the authors cited above the columns; map from Elter & Pandeli, 1990. (After Cassinis, 1996).

Bellerophon Fm. yielding algae, molluscs and *Paratirolites* sp. (Posenato & Prinoth, 1999), on the whole affirm resolutely this age assessment. According to Massari *et al.* (1988, 1994) and Pittau (in Conti *et al.*, 1997 and Cassinis *et al.*, 1999), this Upper Permian succession might be considered coeval with the Central European Zechstein or, in part, with the Tatarian of Russia.

The stratigraphic break between the above two Permian successions shows significant chronological differences (in the central-eastern Southern Alps, probably from approximately 14 to 27 Ma; Cassinis *et al.*, in press). The most consistent values are reached where the Verrucano-Val Gardena blanket lies directly on the pre-Upper Carboniferous basement, due to the progressive spreading of the post-Variscan sedimentation.

In Tuscany, the Alpine Carboniferous and Permian conditions could be again highlighted in some continental areas, in spite of their discontinuity and not being well known. In the Pisan Mts area, the S. Lorenzo schists and the Asciano breccias generally show good affinity with some units found in Liguria and in the central South-Alpine region between the end of the Carboniferous and the onset of the Early Permian. The lack of volcanic rocks, which are only present in the form of clasts within the overlying unconformable Verrucano, could be caused by erosion of magmatic products now occurring in Jano and other sites (*e.g.* Pandeli, in press).

As already stated, Permo-Carboniferous marine strata are also recorded in the Tuscan region (Elba Isl., Jano, etc.), but the bulk of these shallow-water marine deposits seem to match the one cropping out along the Torrente Farma and in nearby areas. In southern Tuscany, on the basis of few but significant paleontological data, the marine sedimentation developed, perhaps irregularly, throughout the Early Permian up to the beginning of the Late Permian.

In this context, therefore, we are forced to suggest that an important gap exists between the Lower Permian (or even slightly younger) deposits and the overlying ones, probably encompassing diachronously the Late Permian and perhaps, in part, the earliest Triassic. The Tuscan Verrucano could have been formed later, normally beginning from the Middle Triassic, as indicated by many authors in the type-locality of the Pisan Mts, above all on the basis of its stratigraphic position.

Igneous events

– Bulgaria

The Upper Palaeozoic igneous, extrusive and intrusive products of Bulgaria need further research, because of the lack of an updated synthesis for interregional correlation. These conditions, therefore, forced us to promote some investigations, which are still in progress, on the Permo-Carboniferous volcanism of the Balkan region.

From the literature (*e.g.* Tchounev & Bonev, 1975; Zukof *et al.*, 1976), the first igneous event occurred during the Variscan orogeny, probably at the end of the Early Carboniferous, through the intrusion of granitoids in the basement. The following event took place throughout the Middle Carboniferous ("Westphalian") and was essentially characterised by calcalkaline andesitic products. A third event was bracketed between the latest Carboniferous and the earliest Permian, again distinguished by the previous deposits. The subsequent episode occurred during the Lower Rotliegend, giving rise locally to the most developed examples. As already mentioned, calcalkaline dacitic to rhyolitic volcanic rocks near Sliven are associated with intrusive and subvolcanic rocks of acidic composition. Later, up to the Triassic, no igneous activity was recorded in Bulgaria.

– Italy

In Italy, the Carboniferous and Permian igneous activity of the areas examined shows some affinities with that of Bulgaria. In the Southern Alps, intrusive bodies were probably initiated at the end of the Carboniferous and continued intermittently up to the early Late Permian (see synthetic data in Cassinis, 1996; Cassinis *et al.*, 1999 and in press). Huge volcanics were widespread only to the west of the Comelico region, probably pertaining to pre-Verrucano-Val Gardena Permian times. Carnia did not experience any manifest plutonic and volcanic activity.

In contrast, the "internal Briançonnais" of the Ligurian Alps includes a first Late Carboniferous episode of calcalkaline rhyolitic ignimbrites which rest unconformably upon a pre-Namurian polymetamorphic basement. Upwards, associated with the finegrained clastics of the Murialdo Fm., there is a second episode formed by andesitic pyroclastics and lavas, generally assigned to latest Carboniferous times (Cortesogno *et al.*, 1982). These volcanics display a subalkaline affinity. Above, during the Early Permian, calcalkaline rhyolitic-to-dacitic ignimbrites and pyroclastics generated the third and last main episode. In the uppermost part, unconformably below the Upper Permian Verrucano, these rhyolitic products show a subalkaline potassic composition.

In the "external Briançonnais", above the aforementioned first Late Carboniferous episode, the coarse- to finegrained Ollano Fm. bearing a late Westphalian megaflora (Block, 1966) includes rhyolitic and andesitic volcanics slightly metamorphosed. The Lower Permian metavolcanics are mainly represented by calcalkaline ignimbrites and are unconformably capped by the Verrucano deposits.

From the above Alpine overview, we are led to remark on the similarity of magmatic events, especially in petrographic and chemical composition, between the Ligurian Alps and Bulgaria.

In Tuscany, the available data on Permo-Carboniferous magmatism are insufficient to provide clear correlations. Apparently, only Permian volcanism can be ascertained so far. This striking difference in the igneous development of this and other Italian and Bulgarian regions is still a matter for speculation, and consequently further research is required.

Tectonic events

– Bulgaria and Italy

Following the Variscan collision, during early to middle Late Carboniferous times, Bulgaria and Italy were locally affected by the onset of an irregular topography, which gave rise to as-yet-uncertain tectonic activity. Some continental basins (*e.g.* in Dobrudgea and western Liguria), partly infilled by magmatic rocks, could be interpreted as fault-bounded subsiding basins that have collapsed in the general geological setting of the Variscan orogeny. This basin-and-swell framework spread progressively, from the latest Late Carboniferous, throughout the above territories and reached its peak of development during the Early Permian. Furthermore, these basins show sedimentary and structural features which are consistent with an extensional regime; they evolved from narrow zones to wider depositional areas. Presumably, transtensional movements determined the inception and in part the subsequent development of this structural scenario. Subsidence, sedimentation and volcanism, in the framework of pronounced tectonism, marked such times.

Subsequently, during the Late Permian, the spread of a more or less ubiquitous red sedimentation (Targovishte Fm., Verrucano Lombardo, Val Gardena Sandstone) and the extinction of any magmatic activity are compatible with a new tectonic framework, probably connected to a marked extensional regime. This led to anorogenic conditions (*e.g.* Bonin, 1988; Cortesogno *et al.*, 1998), as clearly documented by the further presence of bimodal alkaline volcanic products in nearby areas (such as southern France, northern Spain, the Corsican-Sardinian massif), and highlighted by the geological scenarios described in this paper. French authors (Toutin-Morin in Cassinis *et al.*, 1992, 1995) interpret this event as the birth of the Alpine cycle.

The gap to the underlying Lower Permian or older rocks represents a striking tool for subdividing the post-Variscan Late Carboniferous to Permian evolution of southern continental Europe into two tectonosedimentary cycles of the second order (Italian IGCP 203 Group, 1986; and so on). The first cycle is irregularly developed from Moscovian or older ages up to about the Early-Late Permian; the second one encompasses the Late Permian and the Early Triassic, locally ending in later times.

However, other unconformities within these megasequences (*e.g.* in Bulgaria and the northern Apennines) could also allow us to establish other cycles, but, in the authors' opinion, the two above-mentioned cycles are the only ones recognisable on an interregional scale, although they probably acted according to diachronic trends. In other European regions, their boundary is placed at about the Lower-Upper Permian transition, and the consequent discontinuity has been identified with the "Palatine" tectonic phase (*e.g.* Kozur, 1980 b), and with the "post-Saalian" or the "Altmark" phases (*e.g.* Hoffmann *et al.*, 1989). It probably marked the birth of new geodynamic conditions, probably connected with an active reorganisation of plates in Mediterranean and circum-Mediterranean regions.

CONCLUDING REMARKS

In conclusion, the Late Paleozoic evolution of Bulgaria and Italy could be generally interpreted as the result of two main sedimentary cycles, separated by a marked unconformity. The older cycle developed irregularly from a "generic" Middle Carboniferous up to about the onset of the Late Permian; where present, the second cycle began during the Late Permian and continued, at least in some places, up to Triassic times. However, throughout their development, these cycles seem to have been affected by an as-yet-imprecise number of more or less protracted phases of non-deposition and/or erosion. Furthermore, their deposits, which are essentially made up of continental siliciclastic and volcanic products, and also subordinately by marine sediments, often display uncertain ages and prevent clear correlations. A varied and complex geological scenario commonly arose, emphasised once again by the important role played by the Alpine orogenesis in the investigated regions.

In spite of these complications, the unconformity between the two cycles generally appears to have been a very significant event in Permian evolution, which could be related to the opening of Neotethys. Thus, after the Gondwana-Eurasia Variscan collision and the subsequent rearrangement and structuration of the European areas between, a Permo-Triassic extensional regime resolutely developed.

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OVERVIEW OF THE CONTINENTAL PERMIAN DEPOSITS OF BULGARIA AND ROMANIA

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Key words – Permian; stratigraphy; tectonic evolution; correlation; Bulgaria; Romania.

Abstract – Continental Permian deposits show widespread development within the territories of Bulgaria and Romania. The Prebalkan Unit of Bulgaria can be correlated with the Danubian Unit of Romania, while the Balkan and Srednogorie Units may correspond to the Getic Nappe of Romania. This paper presents a short lateral and temporal overview of the lithostratigraphy and sedimentology of the Permian continental sedimentation in the Balkan Mountains, Moesia, Kraishte, the South Carpathians and Apuseni Mountains. Many characteristics were inherited from the Late Paleozoic paleogeography. A Variscan Balkan – Carpathian system coincides with the Variscan orogenic chain and with the adjacent lowlands. In all domains of the Balkan area, the Permian System can be divided into two well-differentiated sedimentary groups (cycles), separated by a clear unconformity. Many similarities with the Romanian territory are established. In both countries, the Lower Permian in particular shows mostly molasse features. Terrigenous, volcanic, volcaniclastic and locally evaporitic sediments accumulated in various depositional systems, such as fluvial, alluvial-plain to palustrine, lacustrine, proluvial, playa, colluvial and sabkha facies.

Parole chiave – Permiano; stratigrafia; evoluzione tettonica; correlazione; Bulgaria; Romania.

Riassunto – I depositi continentali permiani mostrano un esteso sviluppo nei territori della Bulgaria e della Romania. L'unità pre-Balcanica della Bulgaria può essere correlata con l'unità Danubica della Romania, mentre le unità dei Balcani e di Srednogorie possono corrispondere alla falda Getica della Romania. Questo lavoro presenta un breve panorama spaziale e temporale della litostratigrafia e sedimentologia relative alla sedimentazione continentale permiana nei Monti Balcani, in Moesia, in Kraishte, nei Carpazi meridionali e nei Monti Apuseni. Molte caratteristiche furono ereditate dalla paleogeografia tardo-paleozoica. Un sistema balcano-carpatico varisico coincide con la catena orogenica varisica e con le basse terre limitrofe. In tutti i domini dell'area balcanica, il Permiano può essere suddiviso in due ben differenziati gruppi sedimentari (cicli), separati da una chiara discontinuità stratigrafica. Sono precisate molte somiglianze col territorio romeno. In entrambi i paesi, in particolare il Permiano inferiore mostra soprattutto aspetti molassici. Sedimenti terrigeni, vulcanici, vulcanoclastici e localmente evaporitici si accumularono in vari sistemi deposizionali, come facies fluviali, di piana alluvionale, palustri, lacustri, proluviali, di playa, colluviali e di sabkha.

INTRODUCTION

This paper presents a short overview of the Permian continental sedimentation in the Balkan Mountains, Moesia, Kraishte, the South Carpathians and Apuseni Mountains. The continental Permian deposits show widespread development in the study area – the present day territories of Bulgaria and Romania.

OCCURRENCE OF PERMIAN DEPOSITS IN BULGARIA AND ROMANIA

In Bulgaria, the Permian deposits occur mainly in five of the present-day morphotectonic units (from north to south): the Moesia, Prebalkan, Balkan, Sredna Gora and Kraishte units (Fig. 1). In Romania, Permian deposits are recorded in the South Carpathians (with Getic and Danubian units), the Apuseni Mountains (with Bihor Autochthonous and Codru Nappe systems), North Dobrogea, and the Moesian and Scythian Platforms (Fig. 1). Sedimentation took place in several basins – Resita (Getic Nappe), Sirinia and Presacina (Danubian Units) in the South Carpathians, Codru-Bihor Basin in the Apuseni Mountains, Carapelit Basin in North Dobrogea, the Scythian Basin (with Aluat and Lower Danube sub-basins) and the Moesian Basin within the Carpathian foreland.

Surrounded by the Carpathians and the Balkans, the

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Moesian Platform extends over both Romania and Bulgaria (Figs 2, 3). The variations of Permian sedimentation in the Bulgarian part of the Moesian Platform can be distinguished, based on borehole data (Yanev, 1992, 1993a) from the following areas: Vidin-Rasovo (west), Pleven-Tarnovo-Targovishte (centre), Mirovo-Komunari (southeast) and South Dobrogea (northeast).

The Prebalkan Unit of Bulgaria can be correlated with the Danubian Unit of Romania, while the Balkan and Srednogorie Units may correspond to the Getic Nappe of Romania (Figs 2, 3). During the Permian, the Prebalkan Unit represented the former foredeep of the Variscan orogen. In the best exposed, western part of Bulgaria, this unit comprises the localities of Vrashka chuka, Belogradchik, Smolyanovtsi and Vratsa (Tenchov & Yanev, 1963; Yanev & Tenchov, 1978).

The Balkan and Sredna Gora Units coincide with the position of the Variscan orogenic belt, crossing the Balkan Peninsula from WNW to ESE (Figs 2, 3). Within these units occur intramontane basins separated by grabens and half-grabens. They are confined to several diagonally extended, tectonically bounded belts. Within the northern belt (from northwest to southeast) the following localities occur: Stakevtsi, Prevala, Melyane, Draganitsa-Lyu-

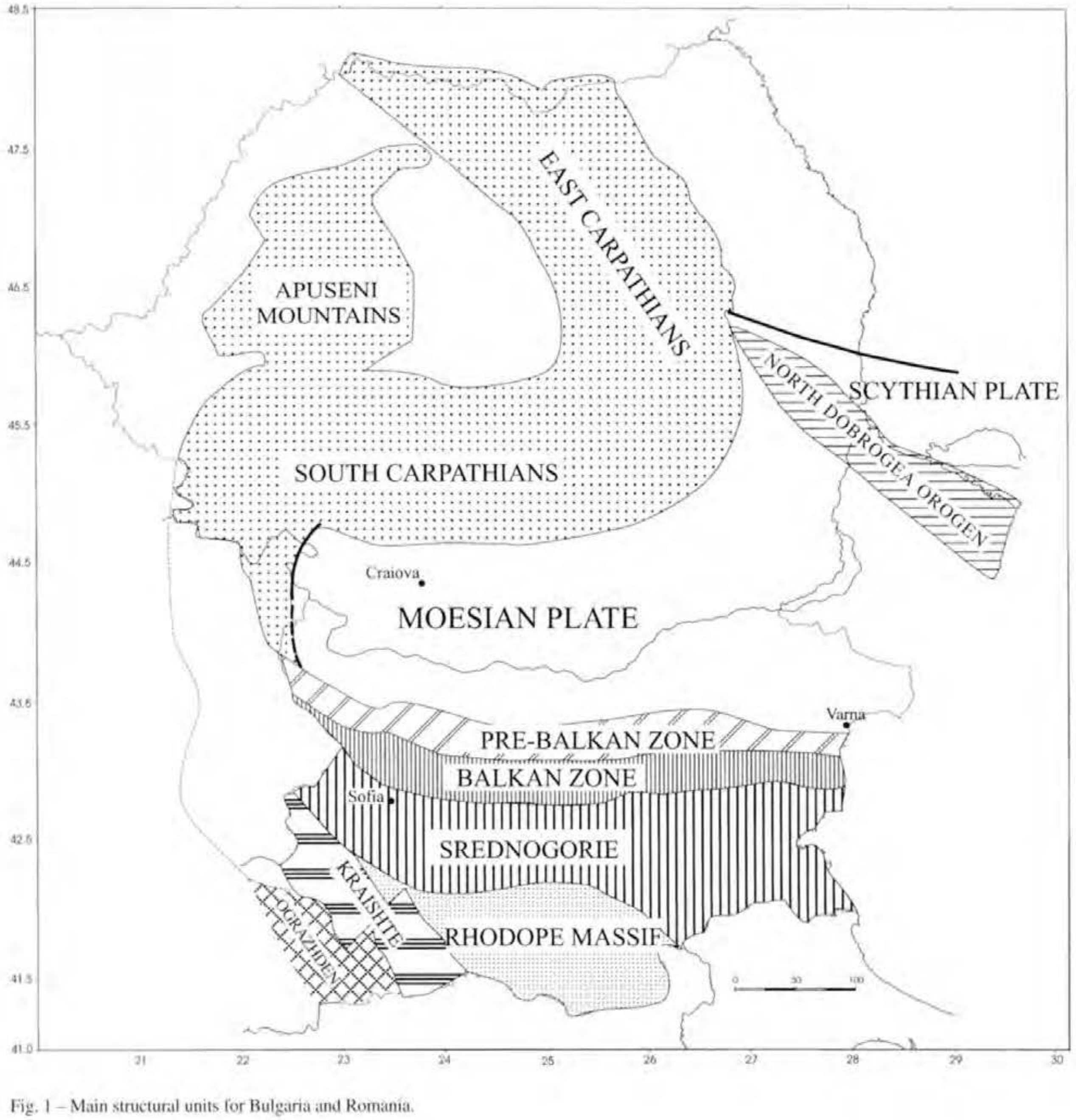

Fig. 1 – Main structural units for Bulgaria and Romania.

tadzhik, Zverino-Ignatitsa, Teteven, Troyan Mountain, and Sliven (Spassov & Zafirov, 1961; Yanev & Tenchov, 1962, 1972, 1978; Chatalov *et al.*, 1962, 1963; Zhukov *et al.*, 1971). The peaks of Midzhur-Kopren, Godech-Buchino Pass, Svoge, and the Sveti Iliya Hills mark the middle zone (Tchumacenko & Shopov, 1965; Yanev, 1981, 1982 a; Chatalov, 1985). The next belt includes localities in Sofioter Stara Planina Mountain, Bunovo area, Lozenska Mountain area, and Chernogorovo (Kulaksazov *et al.*, 1966; Kozhukharov *et al.*, 1980; Yanev, 1982 a).

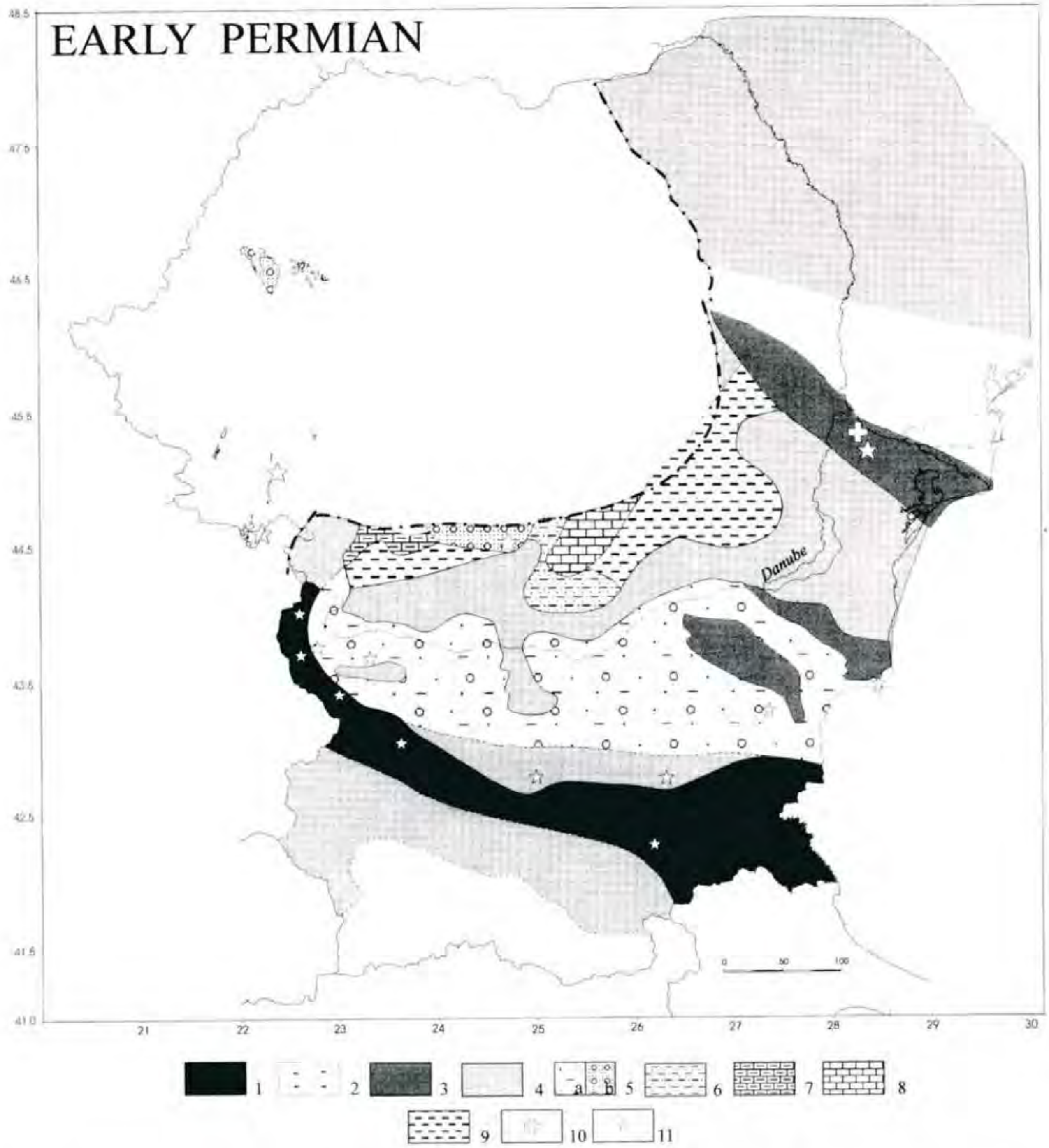

Fig. 2 – Scheme for the main palaeogeographic zones during the Early Permian and their dominating lithology. 1 to 5: Zones of continental sedimentation. 1 – high relief, deposition in intramountain basins of conglomerates, breccia-conglomerates, sandstones and siltstones or non-deposition. 2 - coal bearing deposits in the intramountain basins; 3 - moderate relief, deposition in grabens and half-grabens of conglomerates, sandstones, siltstones, mudstones; 4 - low relief with deposition of more fine grained clastics (mainly sediment by-pass); 5 - low-land with isolated basins: a - with variegated lithology; b - mainly conglomerate and sandstone - bearing; 6 - deltaic, coastal and shallow marine clastics; 7 - shallow marine carbonates and shales; 8 - marine carbonates; 9 - marine shales; 10 - batholiths; 11 - volcanics.

Permian deposits in the Bulgarian Kraishte Unit (in the opinion of S.Y.) may be compared partly with the Romanian Apuseni Mountains. The main difference is the lack of intensive volcanism in the Kraishte area. Present Lower Permian outcrops in the Kraishte are confined only to the Boboshevo-Vukovo area (Yanev, 1982 b), but the earliest distribution of the Upper Permian sediments was larger since their relicts are well exposed between Tran, Noevtsi, Batanovtsi, Boboshevo, Stanke Lisichkovo, Padesh and other villages (Yanev, 1979) (Figs 2, 3).

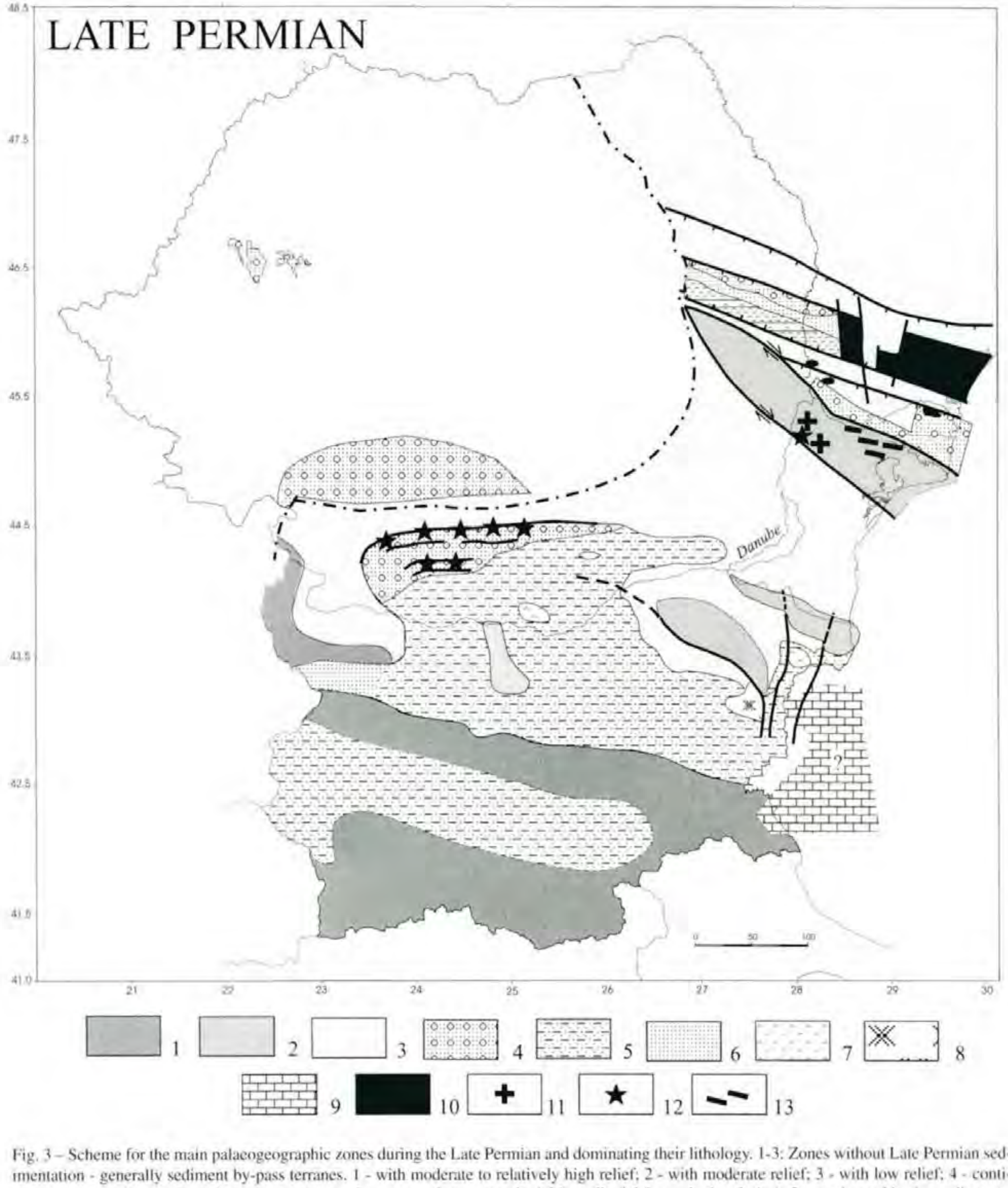

Fig. 3 – Scheme for the main palaeogeographic zones during the Late Permian and dominating their lithology. 1-3: Zones without Late Permian sedimentation - generally sediment by-pass terranes. 1 - with moderate to relatively high relief; 2 - with moderate relief; 3 - with low relief; 4 - continental coarse sedimentation of conglomerates, sandstones, mudstones (proluvial fan, alluvial fan and other facies); 5 - continental basins sedimentation (generally poorly graded red beds: mudstones, sandstones, siltstones, etc.); 6 - deltaic sediments, related with the continental basin delta - mainly sandstones; 7-8 - zones of evaporite sedimentation: 7- sulphate-bearing deposits; 8 - halite-bearing sabkha deposits; 9 - zone of supposed marine carbonate sedimentation; 10 - plateau basalts; 11 - batholiths; 12 - volcanics; 13 - dyke systems.

Permian deposits are lacking in the Rhodope Unit and the Serbo-Macedonian ("Dardan") Massif and so cannot be correlated with any Romanian units.

LITHOSTRATIGRAPHY AND STRATIGRAPHIC CORRELATION OF THE PERMIAN DEPOSITS IN ROMANIA AND BULGARIA

In all domains of the Balkan area, the Permian System can be divided into two well-differentiated sedimentary groups (cycles), separated by a marked unconformity (Yanev, 1981).

From the above general palaeogeographical schemes (Figs 2, 3) and the following Bulgarian stratigraphical successions (Figs 4, 5 and 6), some correlations could be suggested. The NW Balkan prolongation of the Variscan belt is recognised through the western part of the South Carpathians (Banat area). Here, low-scale unconformities occur at the "Stephanian"/Lower Permian ("Autunian") boundary (Secu area), while sedimentary gaps are recorded between the upper "Westphalian" - "Stephanian" sequences (as in the West Balkan domain). The Permian successions are unconformably overlain by Lower Jurassic deposits.

In both the Getic and Danubian Units, only Lower Permian deposits are known. The succession begins with black shaly sediments (fossiliferous, Early Permian in age

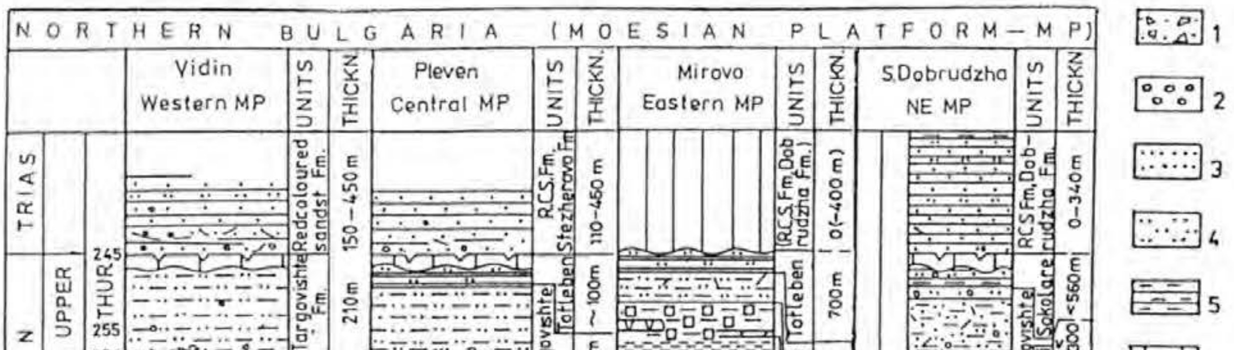

Fig. 4 – Scheme for the lithology, stratigraphic position and thicknesses of selected typical successions of the Upper Paleozoic in the boreholes from the Bulgarian part of the Moesian Plate (see Figs 2-5 in Seghedi *et al.*, this volume). The presented lithostratigraphic units are comparable with some units in the Romanian part of the Moesian Plate.

Symbols (for Figs 4, 5 and 6): 1 - breccia; 2- conglomerate; 3 - sandstone; 4 - siltstone; 5 - shale; 6 - limestone; 7 - dolomite; 8 - anhydrite; 9 - halite; 10 - coal; 11 - volcanics; 12 - volcaniclastics; 13 - carbonate concretions; 14 - anhydrite concretions; 15a - unconformity; 15b - erosional surface; 16 - stratigraphic gaps.

with Autunian type flora), conformably overlain by red beds and volcaniclastic facies, at least late Early Permian (late "Autunian") in age (Raileanu, 1953; Nastaseanu, 1975, 1987; Nastaseanu *et al.*, 1973; Stan, 1987; Stanoiu & Stan, 1986). So far, no paleontological evidence from the red beds of the upper part of the Lower Permian ("Saxonian - Thuringian" age) has been recorded (Antonescu, 1980; Antonescu & Nastaseanu, 1976).

These basins can be correlated with their Bulgarian counterparts. Along the Danube, in the Svinita zone, the Lower Permian Ieliseva Formation can be correlated with the Zelenigrad Formation ("Autunian"; Yanev & Tenchov, 1972) and both the Vranska Formation and the lower two members of the Smolyanovtsi Formation (Variscan orogen zone, Lower Permian; Yanev, 1981). The volcaniclastic sequences of the Sirinia Basin exposed along the Romanian tributaries of the Danube (*e.g.* Staristea Valley) are similar to part of the Vranska Formation near the Vrashka Chuka Hill, the town of Belogradchik and the village of Ozirovo, to the Gyurgich Member of the Smolyanovtsi Formation, and especially to volcanogenic units in the Central Balkan Mountains (Zhukov *et al.*, 1971, 1976; Yanev, 1981, 1982 a).

The Permian sequences of the Resita Basin (Getic Nappe) can be correlated with their Bulgarian counterparts in the following way: the Ciudanovita Formation, mainly the basal Girliste Member (lowest Lower Permian - lower "Autunian", mainly black pelites) is similar to the Levitsa Formation (cropping out around Stakevtsi and Prevala), the Dalgi Del Formation (cropping out near Melyane village), the Lyutadzhik Formation cropping out in the Ozirovo-Lyutadzhik areas) and the Buk Formation (cropping out in the Zverino-Ignatitsa areas). The topmost member of the Ciudanovita Formation, the Lisava Member (at least the upper part of the Lower Permian - upper "Autunian"), in red-bed facies, is rather similar to some sequences of the Milinska Formation (Tchumacenko & Shopov, 1965) and the Koritarska Member of the Smolyanovtsi Formation (Yanev, 1981). For the Upper Carboniferous sequences in Resita Basin (Resita Formation), the similarity points to the Starchovdol Formation (Stakevtsi area), the Melyane Formation (near the village of Melyane), the Ekimska and Draganitsa formations (Draganitsa-Ozirovo-Byala Rechka areas) or the Ochindol Formation (Zverino and Ignatitsa areas).

The Permian of the Codru-Bihor Basin conformably overlies the Upper Carboniferous deposits, or unconformably covers various older basement rocks. The Per-

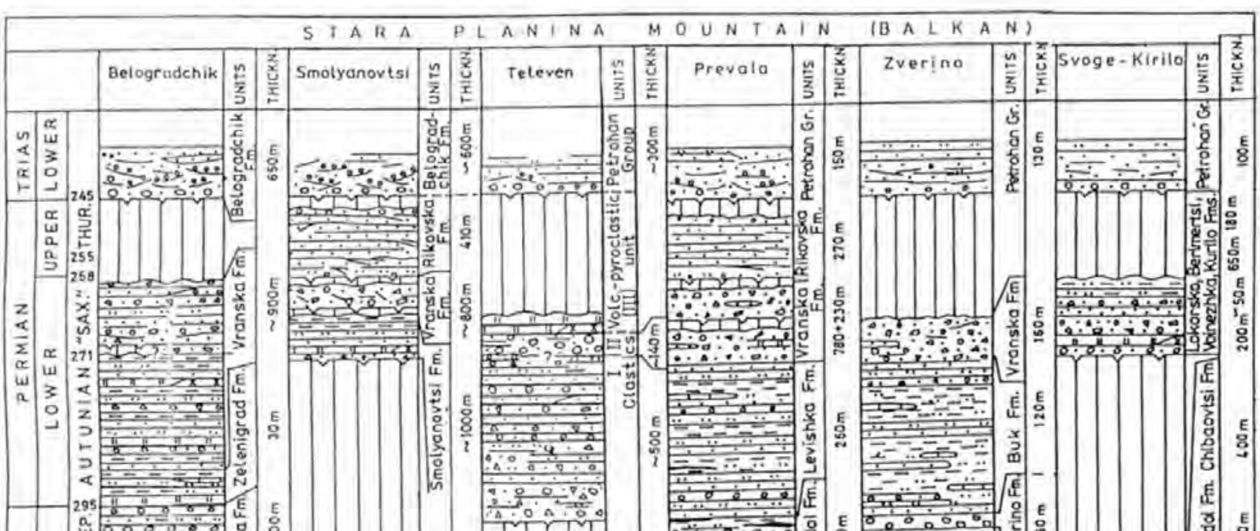

Fig. 5 – Scheme for the lithology, stratigraphic position and ticknesses of selected typical successions of the Upper Paleozoic related to the Variscan Orogen (Stara Planina Mountain System in Bulgaria). The presented lithostratigraphic units are compared with some units mentioned in the text for the Romanian part of the Variscan Orogen (Carpathian Mountain System) (symbols as in Fig. 4).

mian red-beds grade upwards (?) to Lower Triassic (Buntsandstein) quartzitic sandstones (Bleahu, 1963; Bleahu *et al.*, 1985; Bordea & Bordea, 1982; Mantea, 1985). The sequences can be correlated to the Permian from the southern basins in Bulgaria (the area south of the Balkan Mountains, including Kraishte and present-day Sredna Gora Units). The Lower Permian in this domain also conformably overlies upper "Stephanian" sediments or unconformably covers older basement rocks. The Upper Permian red-bed clastic and silty-shaly deposits are separated from the Lower Permian sediments by an erosional surface, or they lie unconformably on various older metamorphic, igneous or sedimentary rocks. In the Kraishte area a narrow unconformity between the Upper Permian and Triassic sediments is recorded (Yanev, 1964).

Facial similarities between the red beds of the Aries Valley (close to Arieseni) and their Bulgarian counterparts are the following: the upper part of the Gabra Formation - upper "Stephanian" to Lower Permian in age (transition between the "Autunian" and "Rotliegend" facies; Kozhukharov *et al.*, 1980), the Tarnavska Formation (Lower Permian) and the Ravulya Formation (Upper Permian) in Lozenska Mountain, the Boboshevo Formation (Lower Permian), in the Vukovo area (Yanev, 1982 b), the Skrino Formation (Zagorchev, 1980), or the Noevtsi, Kiselichka and Nepraznentsi formations (Upper Permian or (?) Upper "Rotliegendes" in Yanev, 1979).

The age and lithology of the Permian deposits from the northern part of the Moesian Platform (Romania) generally corresponds to those on the southern, Bulgarian part (Yanev, 1992, 1993 a). This is caused by the mirrored positions of both zones – in the foreland of the Variscan chain, relatively close to the Balkan part southwards and to the Carpathian part, northwards. The distribution of the coarse, proximal facies in the northern and southern areas of the "Platform" suggests tectonically controlled deposition, related to E-W to NW-SE trending extensional faults. A second control on Permian sedimentation was the active subaerial volcanism, which according to Bulgarian authors was developed only in the Early Permian (lower part of the Rotliegend facies). In the Romanian part of the Moesian Platform, bimodal volcanism continued during the Triassic, as indicated by boreholes.

For Bulgaria during Late Paleozoic times, two main cycles of continental sedimentation can be again envisaged in the whole eastern part of the Balkan Peninsula. The first group (generally spanning Late Carbonifer-

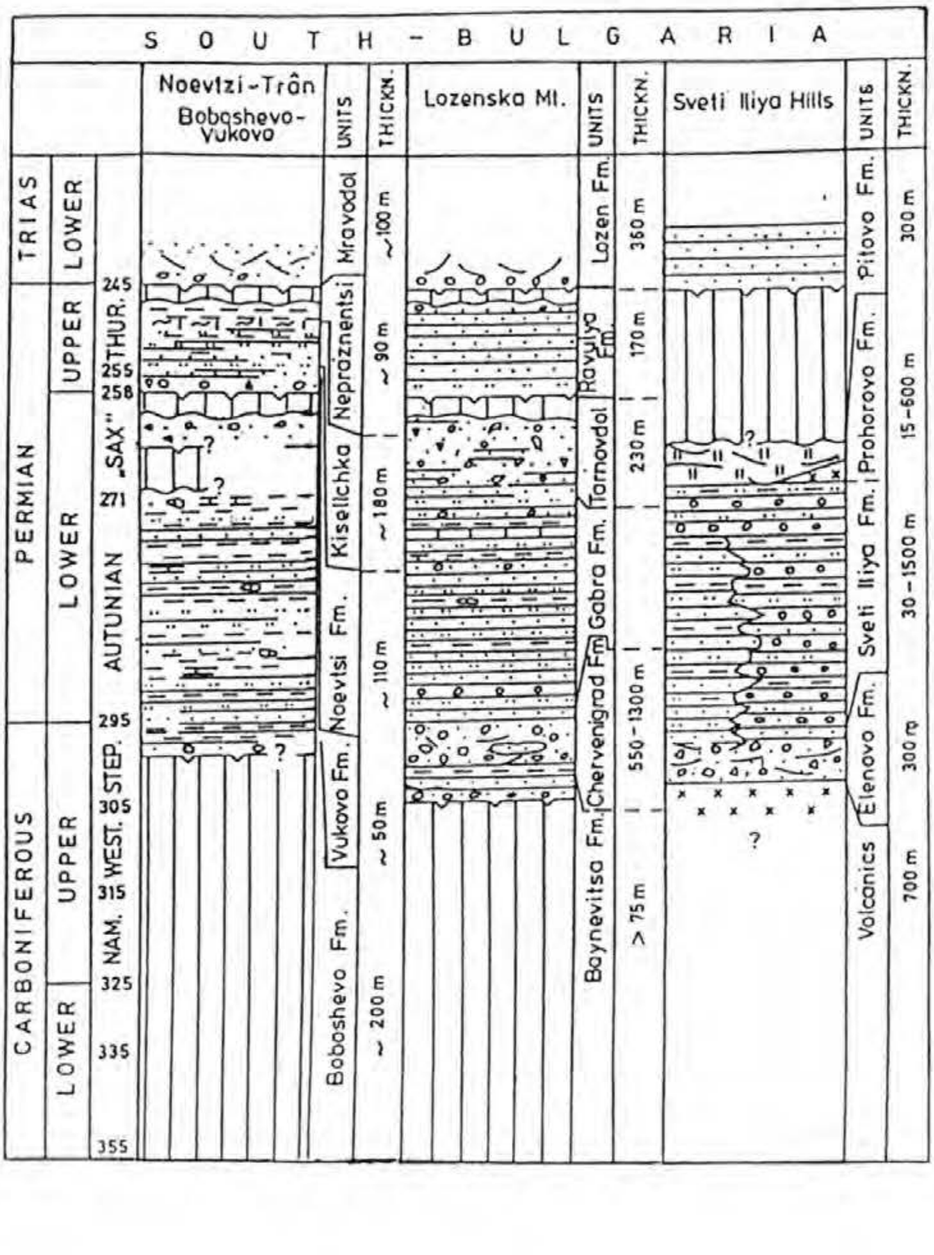

Fig. 6 – Scheme for the lithology, stratigraphic position and thicknesses of selected typical successions of the Upper Paleozoic in South Bulgaria (for symbols see Fig. 4)

ous, mainly late Stephanian to Early Permian times) consists of lacustrine, fluvial and proluvial fan deposits, as well as of volcanic rocks, both infilling intermontane and foremontane grabens or semigrabens (Yanev, 1969, 1988, 1989). The basins were separated by highs of Lower Paleozoic, metamorphic and igneous basement rocks. The boundary faults generally have WNW-ESE trends and often coincide with long-lived tectonic structures reactivated as late as the Alpine orogeny.

The second group (Late Permian, from the Tatarian (?) to the P/T boundary after data from Schirmer, 1960 and Yanev, 1993 a) is represented by deltaic and continental clastics and, only in the southeastern part of the Moesian domain (in the Provadia depression), by halite evaporites, as well as by sulphate evaporites in a zone north of the town of Varna (Yanev, 1993 a). These deposits form a widespread blanket, covering both basins of the first group and the surrounding highs of western Bulgaria (Moesian Platform). The continuity of tectonic control is documented by strong thickness variations - from a few metres (or total absence in many localities of the former Variscan Orogen, such as over the NE-Bulgarian Uplift) to more than 1200 m in the depocentres and in areas with evaporite sedimentation (former foremontane depression).

SEDIMENTOLOGY, FACIES AND SOME PALEOGEOGRAPHICAL ASPECTS OF THE PERMIAN DEPOSITS OF BULGARIA AND ROMANIA

Permian deposits in both Romania and Bulgaria generally show molasse features, mainly and more typically for the Lower Permian in the Balkan, the Prebalkan, the Sredna Gora, the Kraishte, the South Carpathians, the Apuseni Mountains and the Carapelit Basin. In the Moesian and Scythian Platforms, as well as in the Kraishte area, the molassic character, especially of the Upper Permian sediments, is not so obvious, since both continental-basins and partly transitional facies occur. The Permian sequences from the present-day Alpine fold-belts show clear Variscan molasse features. This evidence is related to sedimentation in relatively narrow, deeper intramontane basins and half-grabens within a folded terrain with steep relief. Those sequences from the Moesian and Scythian Platforms are fault-related, since deposition occurred in shallower, sometimes larger grabens and half-grabens inside a hilly terrain.

Terrigenous, volcanic, volcaniclastic, and locally evaporitic sediments accumulated in various continental environments: from fluvial, proluvial, playa, colluvial and alluvial-plain to palustrine, lacustrine, continental-basin and sabkha conditions (Yanev, 1970, 1989).

For Bulgaria, the chain of the Variscan Orogen extended NW-SE across Bulgaria, bordered by lowlands both to the north and the south. In intramontane valleys within the orogen, as well as to its borders were deposited clastic, shaly and coal-bearing sediments in river-beds, terrace, lacustrine, palustrine and other facies (Yanev, 1969, 1989). They follow from proluvial cones and playa sediments during the late Early Permian (two clastic successions separated by erosional surfaces, and as facies corresponding to the early Rotliegend and late Rotliegend). During the Late Permian, two larger continental basins formed to the north and the south, controlled by a lower-altitude main watershed (Yanev, 1981).

For Romania, the molassic character is well defined in intramontane basins in the South Carpathians (Resita, Sirinia and Presacina basins) and in the Apuseni Mountains. The molassic sedimentation was controlled by alluvial, fluvial, lacustrine, and swampy (with no significant coal seams as a result) depositional environments, to which volcanic and volcaniclastic material was added. For the South Carpathians, the ratio of clastic vs. volcaniclastic sedimentation is high in the Resita Basin and low in the Sirinia and Presacina basins, where volcanic and volcaniclastic rocks predominate.

In the Apuseni Mountains, terrigenous and volcaniclastic sedimentation occurs as well, the dominance of one sedimentological type over the other depending on area. The Permian deposits conformably overlie Upper Carboniferous deposits, the terrigenous facies often presenting typical red-bed features. The terrigenous vs. volcaniclastic sedimentation continued until the Triassic, the Permian/Triassic boundary being difficult to establish within the red-bed facies.

TECTONIC EVOLUTION OF THE PRE-LATE PALEOZOIC TERRANES DURING THE PERMIAN

The various regional, lithological, paleoclimatic, paleomagnetic, paleobiogeographical and other data show a peri-Gondwana provenance for the basement of the Upper Paleozoic successions in Bulgaria (Yanev, 1997 and cited references). Three Lower Paleozoic terranes (from north to south): Moesian, Balkan and Thracian are distinguished (Yanev, 1990, 1993 b, 1997). At the start of the Late Carboniferous, the *en echelon* movement from Gondwana to Paleo-Europa brought the Moesian and Balkan (+ Thracian?) terranes into collision. The building of the Variscan Orogen was related to the collisional accretion between both Moesian and Balkan terranes. The development of the Late Carboniferous and Permian molasse sedimentation took place in late-orogenic and post-orogenic conditions. During the Late Permian, the formerly variable relief decreased in energy and deposition was controlled particularly by transtensional movements. As a whole, the Perm-

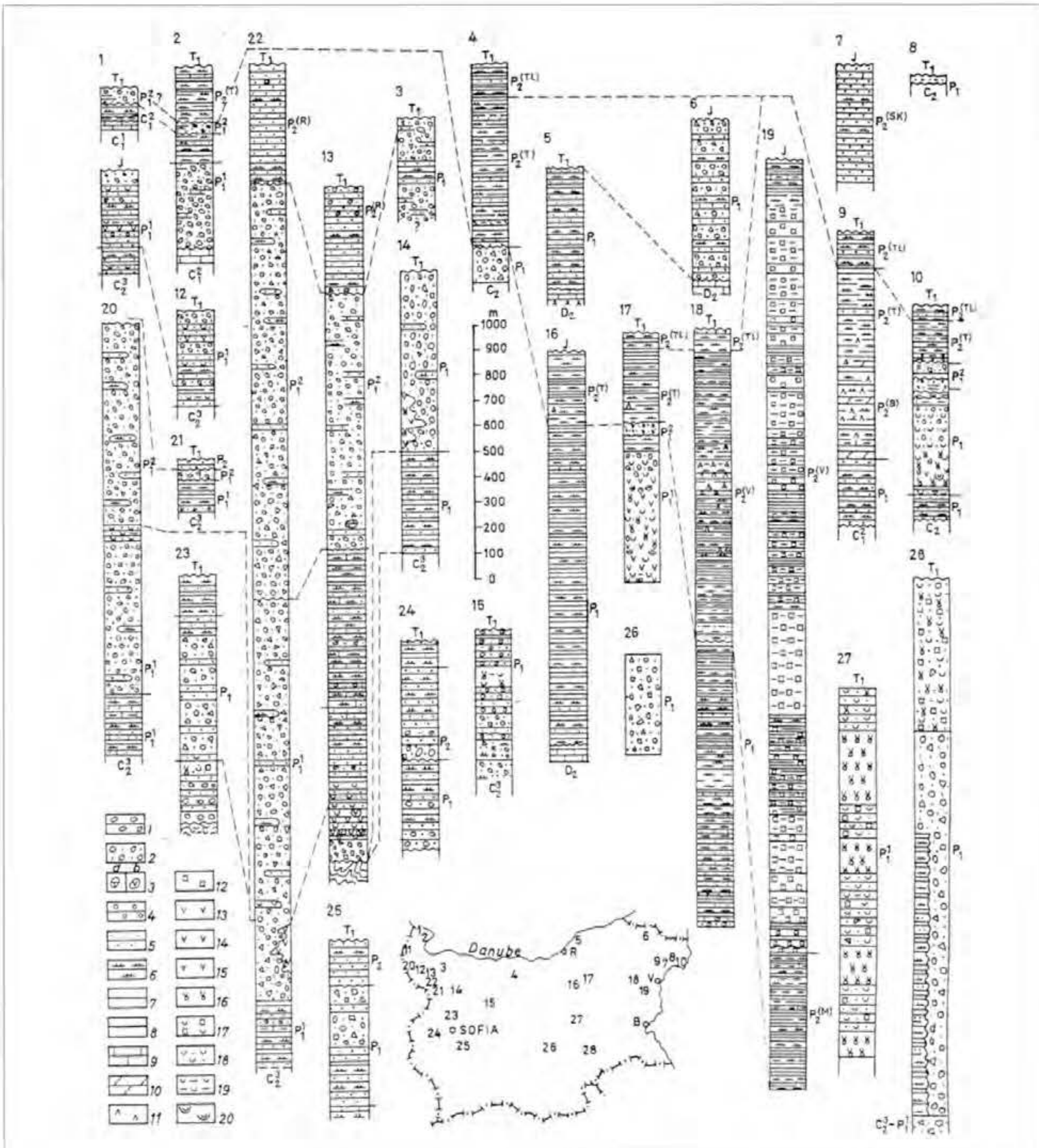

Fig. 7 – Schematic logs of some typical Permian successions in Bulgaria (based on outcrops and borehole data) presented with their comparable real thicknesses.

Lithological symbols: 1 - conglomerate; 2 - breccia; 3 - concretions: a - limey, b - anhydritic; 4 - gravelite; 5 - sandstone; 6 - siltstone; 7 - argillite; 8 - coal; 9 - argillaceous limestone; 10 - dolomite; 11 - anhydrite; 12 - halite (salt); 13 - andesite; 14 - trachyte; 15 - latite; 16 - dacite; 17 - xenotuffs; 18 - tuffs; 19 - tuffites; 20 - ignimbrites.

Stratigraphic symbols: D_3 - Upper Devonian; C_1 - Lower Carboniferous; C_1^1 - Lower Carboniferous, Tourniasian; C_1^2 - Lower Carboniferous, Visean; C_2 - Upper Carboniferous; C_2^3 - Upper Stephanian; P_1 - Lower Permian (Rotliegend); P_1^1 - Lower Rotliegend; P_1^2 - Upper Rotliegend; $P_2^{(TG)}$ - Upper Permian, Targovishte Fm.; $P_2^{(TL)}$ - Upper Permian, Totleben Fm.; $P_2^{(R)}$ - Upper Permian, Rikovska Fm.; $P_2^{(V)}$ - Upper Permian, Vetrino Fm.; $P_2^{(M)}$ - Upper Permian, Mirovo Fm.; $P_2^{(SK)}$ - Upper Permian, Sokolevo Fm.; T_1 - Lower Triassic; J - Jurassic.

Sketch: distribution of sections. Localities: 1 - Gomotartsi; 2 - Vidin; 3 - Rasovo; 4 - Totleben ; 5 - Chereshovo; 6 - Severtsi; 7 - Sokolovo; 8 - Gurkovo; 9 - Bezvoditsa; 10 - Kaliakra; 11 - Kiryaevo; 12 - Belogradchik; 13 - Smolyanovtsi; 14 - Ozirovo; 15 - Teteven; 16 - Dolna Zlatitsa; 17 - Vasil Levski; 18 - Vetrino; 19 - Mirovo-Hrabrovo; 20 -Stakevtsi; 21 - 22 - Prevala; 23 - Kurilo; 24 - Kraishte (Boboshevo-Noevtsi); 25 - Lozen Mt.; 26 - Chernogorovo; 27 - Sliven; 28 - Sakar Mt.

ian basins evolved from narrow zones (the Variscan Orogen and locally the Moesian and Kraishte lowlands) into wider depositional areas, due to lower relief and the increase of accumulation areas at the expense of source areas.

CONCLUSIONS

As parts of the Carpathian-Balkan chain, Romanian and Bulgarian territories recorded the influence of Variscan and Alpine orogenies in the same way, with almost the same evolution, stratigraphy and depositional features. This fact is demonstrated by the possibility of correlation between the South Carpathians and the Balkans. In the Moesian Platform, shared by Romania and Bulgaria, the structure and stratigraphy could well be correlated, this prospect remaining valid for the Permian deposits presented here. At the same time, the comparison between Permian deposits from Apuseni Mountains and the Kraishte area is not so certain. Those from North Dobrogea and Bulgarian zones without Late Paleozoic sedimentation (Thracian and Dardan massifs) are difficult to correlate.

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THE PERMIAN SYSTEM IN ROMANIA

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Key words – Permian; sedimentology; paleontology; magmatism; tectonics; Romania.

Abstract – Permian deposits from major tectonic units in Romania (South Carpathians, East Carpathians, Apuseni Mountains, North Dobrogea, Moesian and Scythian Platforms) are mainly developed in continental facies, with molassic characteristics. Red beds are the dominant facies, but basal black and grey shaly deposits are present in the South Carpathians and Apuseni Mountains. In all areas, sedimentation took place mostly in alluvial fan, fluvial and lacustrine systems. Shallow-marine carbonate platform conditions were restricted to the northern part of the Moesian Platform in the Early Permian. Evaporites are associated with the red beds only within the platforms.

The sedimentary record of most basins includes volcano-sedimentary sequences. The Permian volcanism was bimodal, with an alkaline signature in the Apuseni Mountains, North Dobrogea and Scythian Platform. The basalt-rhyolite bimodal association typically occurs in the South Carpathians, Apuseni Mountains and the Moesian Platform, while the basalt-trachyte association is found in North Dobrogea and the Scythian Platform.

The South Carpathians (Getic Nappe and Danubian units) yielded by far the most fossiliferous Permian deposits in Romania, the fossil remains being represented by flora (compressed macroflora, microflora), and fresh water fauna (ganoid fishes, ostracods, bivalves). The Apuseni Mountains include deposits yielding flora (silicified woods, microflora), while the North Dobrogea has yielded no fossils to date. The Moesian and Scythian Platforms include faunal remains, but palynological evidence was also found in the latter.

An extensional tectonic setting, related to Late? Permian rifting, is suggested by both field evidence and the geochemistry of the volcanic rocks from the South Carpathians, Apuseni Mountains, Moesian and Scythian Platforms. Typical rift basins are those from the platforms. For North Dobrogea, magmatic associations illustrate the transition from a compressional, post-collisional setting in the Early Permian to a transtensional setting in the Late Permian, related to the tectonic collapse of the Hercynian crust.

Parole chiave – Permiano; sedimentologia; paleontologia; magmatismo; tettonica; Romania.

Riassunto – I depositi permiani delle maggiori unità tettoniche della Romania (Carpazi meridionali, Carpazi orientali, Monti Apuseni, Dobrogea settentrionale, Piattaforme Moesia e Scitica) sono rappresentati essenzialmente da facies continentali con caratteristiche di molasse. I *red beds* corrispondono alla facies dominante, ma peliti (*shales*) di color nero affiorano altresì, in posizione sottostante, nei Carpazi meridionali e nei Monti Apuseni. In tutte le sopra citate aree, la sedimentazione si espresse principalmente in sistemi di coni alluvionali, fluviali e lacustri. Condizioni di piattaforma carbonatica di mare basso furono limitate alla parte settentrionale della Piattaforma Moesia, nel corso del Permiano inferiore. Evaporiti si trovano associate a *red beds* solo all'interno delle piattaforme. Il record sedimentario della maggior parte dei bacini include sequenze vulcano-sedimentarie. Il vulcanismo permiano fu bimodale, con un marchio di natura alcalina nei Monti Apuseni, nella Dobrogea settentrionale e nella Piattaforma Scitica. L'associazione bimodale basalto/riolite ricorre tipicamente nei Carpazi meridionali, nei Monti Apuseni e nella Piattaforma Moesia, mentre l'associazione basalto-trachite è presente in Dobrogea meridionale e nella Piattaforma Scitica.

I Carpazi meridionali (Falda Getica e unità Danubiche) comprendono di gran lunga i maggiori depositi fossiliferi della Romania, con resti fossili rappresentati da flore (macroflore limitate, microflora) e faune (pesci ganoidi, ostracodi, bivalvi). I Monti Apuseni includono depositi a flore (legni silicizzati, microflora), mentre la Dobrugea settentrionale non ha ancora dato fossili utili per eventuali datazioni. Le Piattaforme Moesia e Scitica contengono resti faunistici; tuttavia, nella seconda di esse è stata anche riscontrata la presenza di elementi palinologici.

Un assetto tettonico estensionale riferito al *rifting* permiano è suggerito da evidenze di campagna e dalla geochimica delle rocce vulcaniche provenienti dai Carpazi meridionali, dai Monti Apuseni, e dalle Piattaforme Moesia e Scitica. Bacini tipici di *rift* sono quelli di piattaforma. Per quanto riguarda la Dobrogea settentrionale, le associazioni magmatiche segnano la transizione da un assetto compressionale e post-collisionale, durante il Permiano inferiore, ad un assetto transtensivo, riferito al collasso tettonico della crosta ercinica, nel corso del Permiano superiore.

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INTRODUCTION

In Romania, Permian deposits are exposed in the Alpine belts (East and South Carpathians, Apuseni Mountains and North Dobrogea orogen), but they are also known from boreholes in the Moesian and Scythian Platforms (Fig. 1). In the East Carpathians, scarce terrigenous sediments overlain by Lower Triassic sandstones are ascribed to the Permian within the Bucovinian, Sub-Bucovinian and Infra-Bucovinian Nappes. In the South Carpathians the Permian deposits are recorded in both the Getic and Danubian Nappe systems. The Permian occurrences are confined to the central part of the Apuseni Mountains (Codru and Biharia Nappe systems and Bihor "autochthonous" unit of the Codru Moma, Bihor and Padurea Craiului Mountains).

The Permian sedimentation took place in several basins in the South Carpathians – Resita and Pui (Getic Nappe), Sirinia and Presacina (Danubian units) (Codarcea, 1940; Raileanu, 1953; Stilla & Luta, 1968). Small patches of thin Permian red beds, mostly fanglomerates, initially described as "Verrucano" facies, are scattered locally in the area of the Danubian Window and in the Godeanu outlier of the Getic Nappe (Gherasi, 1937; Pavelescu, 1953).

The Resita Basin exposes the sedimentary cover of the Getic Nappe in the westernmost part of the South Carpathians (Banat region). It is a north-south elongated basin, faulted and folded longitudinally, its deposits belonging to a Variscan (Westphalian A?-B-Lower Permian - "Autunian") and an Alpine cycle (Hettangian-Albian). The Pui area is located within the Hateg Depression (central South Carpathians), which is dominated by Mesozoic and Tertiary deposits.

The Sirinia Basin represents the cover of the Upper Danubian units and lies in the southwesternmost Carpathians (Banat region), mainly within the Almaj Mountains. It is a north-south oriented basin, located eastwards of and parallel to the Resita Basin. The Presacina Basin, oriented north-south, lies eastwards of the Sirinia Basin. Both Danubian basins include products of two main cycles, a Variscan and an Alpine cycle.

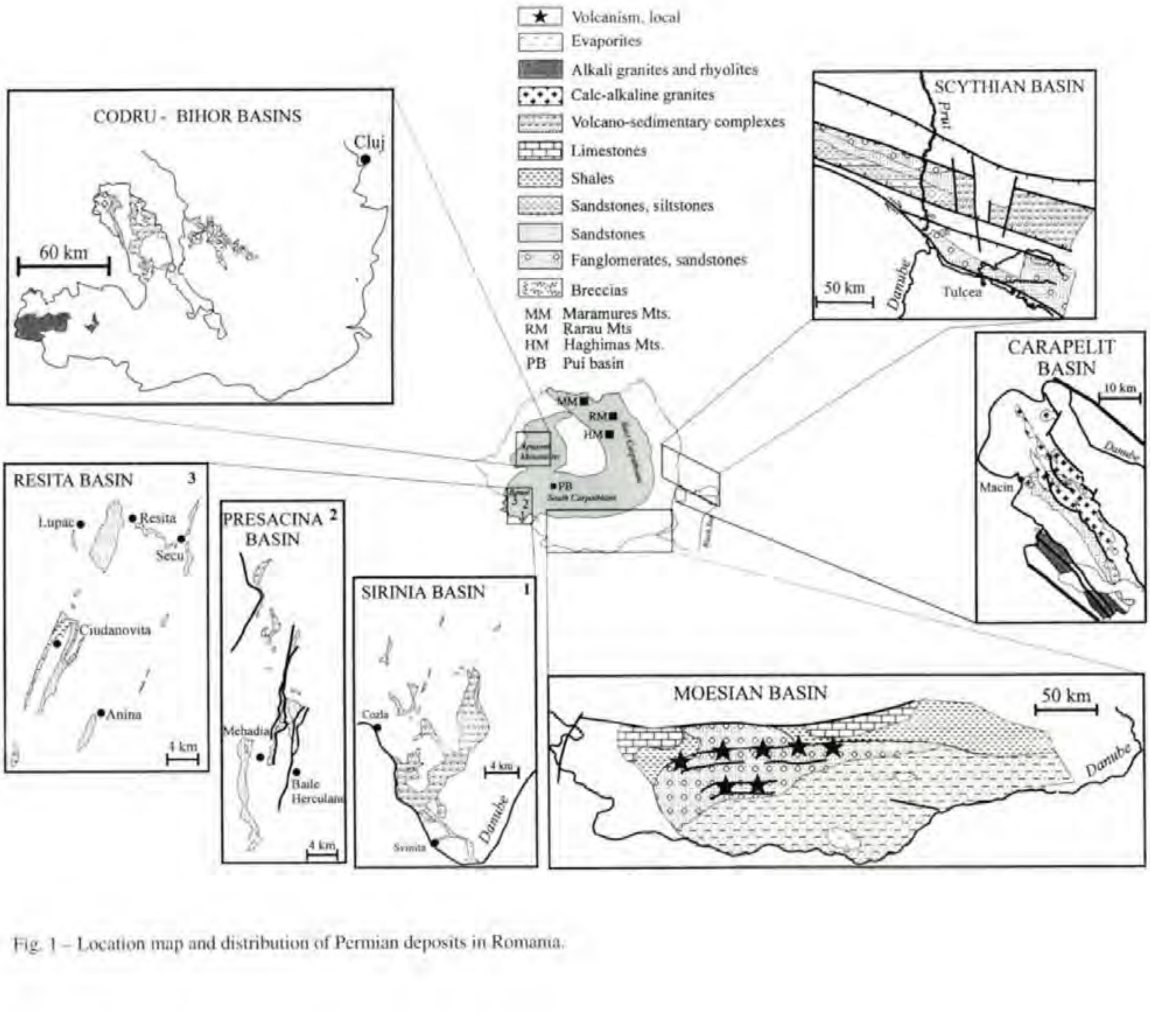

Fig. 1 – Location map and distribution of Permian deposits in Romania.

The Permian of the Apuseni Mountains is known in the Codru Nappe System (Dieva, Moma, Finis and Codru Nappes) of the Codru Moma Mountains, as well as in the Biharia Nappe System (Garda and Arieseni Nappes) and the Bihor Unit (Bihor "Autochthonous") of the Bihor and Padurea Craiului Mountains (Bleahu, 1963; Istocescu *et al.*, 1970; Patrulius, 1972; Bordea & Bordea, 1982; Bleahu *et al.*, 1985; Dimitrescu, 1988). Recent studies (Bordea & Bordea, 1993) revealed the presence of the Lower Permian bioturbated sandstones in the Highis Mountains (south-western part of the Apuseni Mountains).

In North Dobrogea, Permo-Carboniferous continental sedimentation took place in largely E-W oriented, narrow piggyback basins related to back-arc thrusting (Seghedi & Oaie, 1995 a). The NW-SE elongation of the outcrop area of Carapelit Formation (Fig. 1) is the result of subsequent deformation, and preservation of the Upper Paleozoic sediments in the core of a Kimmerian syncline.

An E-W oriented rift basin controlled the Permian sedimentation in the northern part of the Moesian Platform. This was located south of an elongated basement high (Craiova - Bals - Optasi rise) (Paraschiv, 1979), which probably represented a rift shoulder. Other part of the Permian basin occurred in the southern part of the Romanian Moesian Platform and continued southwards into the present-day Bulgarian territory.

The Scythian Basin, bordered and fragmented by major faults, is oriented WNW-ESE and includes the Aluat-Sarata and Lower Danube sub-basins (Neaga & Moroz, 1987), separated by tectonic ridges or push-ups.

STRATIGRAPHY

Coarse-grained sediments (Haghimas Breccias) in the Bucovinian Nappe of the East Carpathians, consisting mainly of unsorted clasts of metamorphic rocks, were ascribed to the Permian in Rarau and Haghimas Mountains, based on palynological associations (Muresan, 1970). In the Bucovinian and Sub-Bucovinian Nappes from the Maramures Mountains, the Haghimas Breccia directly overlies the metamorphic basement and is unconformably overlain by red Permian siliciclastics; based on field relations, an Upper Carboniferous and possibly Lower Permian age was ascribed to these rocks, overlain in turn by Lower Triassic sediments (Sandulescu *et al.*, 1989). A Permian age was assigned to the red sandstones and conglomerates from the Infrabucovinian Nappes in the same area, by lithological correlation with the Rozis series from Ukraine (Sandulescu, 1985).

The Variscan molasse deposits of the Resita Basin are subdivided into the Upper Carboniferous ("Westphalian A?-B" – "Stephanian") Resita Formation and the Lower Permian ("Autunian") Ciudanovita Formation (Bucur, 1991); the latter consists of lower black deposits (the Gîrliste Member) overlain by red beds (the Lisava Member) (Fig. 2). The Ciudanovita Formation lies conformably (west-

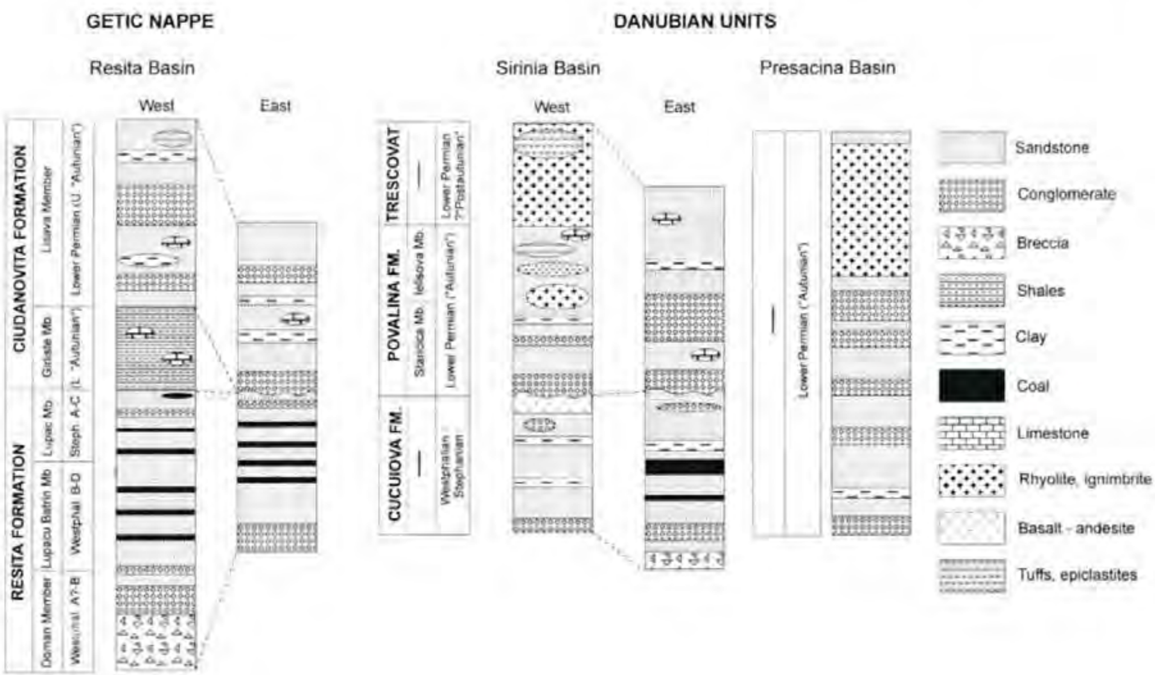

Fig. 2 – Variscan molasse deposits of the South Carpathians. Data compiled from Raileanu (1953), Nastaseanu *et al.* (1981), Stanoiu & Stan (1986) and Stanoiu *et al.* (1996).

wards) or unconformably (eastwards) on the Upper Carboniferous deposits of the Resita Formation, and it is unconformably overlain by the Lower Jurassic deposits of the Steierdorf Formation (the first unit of the Alpine cycle). The typical molasse characteristics suggest that the continental deposits of the Ciudanovita Formation accumulated in an intramontane depression. During Late Carboniferous and Early Permian times, the lateral shift of the depositional centre of the basin explains the numerous heteropic, partially juxtaposed Upper Paleozoic sequences in the Resita Basin (Stanoiu *et al.*, 1996). The Girliste Member is the lowest sequence of the Lower Permian, dominated by lacustrine deposits. Its thickness (150-300m) decreases from the west (Lupac, Ciudanovita and Jitin areas) to the east, disappearing in the Secu area. This member consists of black pelites with sandstone and freshwater limestone interlayers; rare microconglomerates or thin coal layers also occur. The Lisava Member is represented by red beds (red, green and grey sandstones, clays, conglomerates), with freshwater limestone interbeds; rare volcanic tuff and tuffite interlayers occur westwards in Lupac area. Its thickness varies between 1000-1500m. The paleontological data recorded within the middle part of the succession indicate an upper Lower Permian ("Autunian") age, but there are no markers for younger ages ("Saxonian"-"Thuringian"?) recorded for the uppermost sequences of the Lisava Member (Antonescu, 1980).

Below the Lower Jurassic deposits in the Pui area, scarce outcrops of grey sandstones recovered close to the Cioclovina Cave yielded palynological assemblages ascribed to the Lower Permian (Stilla & Luta, 1968; Stilla, 1980).

The Variscan cycle of the Sirinia Basin (Fig. 2) is represented by "Westphalian-Stephanian" clastics overlain by

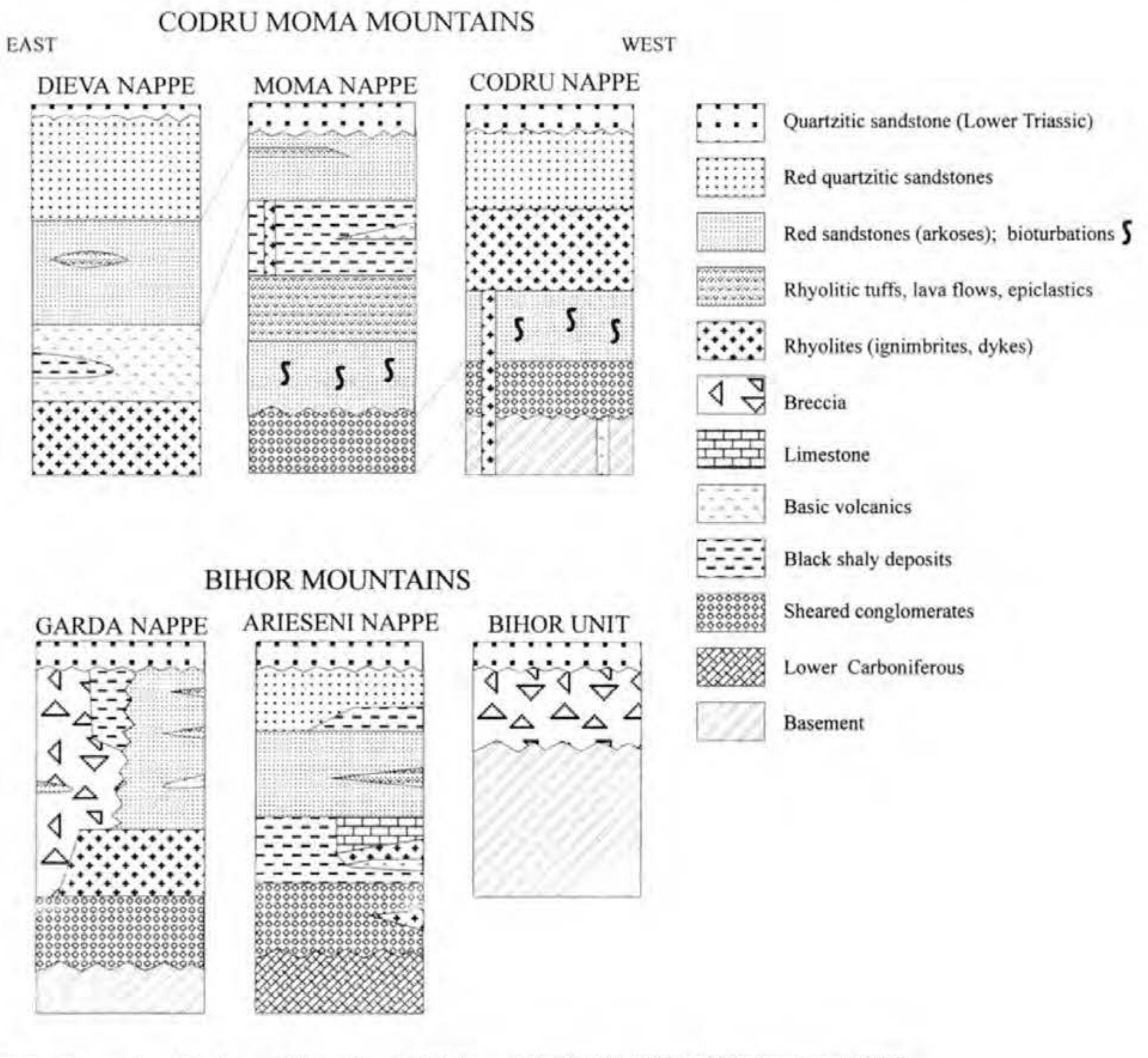

Fig. 3 – Permian logs of the Apuseni Mountains, after Bleahu *et al.* (1979, 1981, 1985) and Dimitrescu *et al.* (1977).

the Lower Permian sediments of the Povalina Formation; this Permian sequence unconformably rests on the lower-middle "Stephanian" deposits, upper "Stephanian" being absent. The Povalina Formation includes black and grey shales (Staricica Member) overlain by red beds (Ielisova Member) (Stanoiu & Stan, 1986). The Ielisova Member conformably overlies the Staricica Member in the central areas of the basin, and unconformably rests on the basement towards the marginal parts of the basin, mainly eastwards (Stanoiu *et al.*, 1996). The red beds are dominated by thick volcano sedimentary sequences, with local lacustrine limestones (Raileanu, 1953). Paleosol layers with caliche concretions often occur in the red bed sequence. The volcano-sedimentary dominance is very strong within the western part of the Sirinia Basin, while to the east the sedimentation is more terrigenous. The black fine clastics of the Staricica Member are thin and fossiliferous.

Within the Presacina Basin, the Lower Permian deposits show both terrigenous (red beds) and volcaniclastic deposits (Codarcea, 1940; Nastaseanu *et al.*, 1973; Nastaseanu, 1975, 1987) (Fig. 2).

Volcaniclastic, volcanic and terrigenous Permian sequences are exposed in various units of the central Apuseni Mountains (Fig. 3). The discovery of Lower Permian ("Autunian") rocks is based on geometric and facies criteria; a possible "Saxonian" or even younger Permian age is indicated by silicified wood remains. The basal phyllitic sequence of the Permian corresponds to the black bituminous pelites and is overlain by red beds with tuffaceous interbeds. To the east, in the Bihor Mountains, the Lower Permian consists of sheared conglomerates; the Upper Permian sequence starts either with ignimbritic rhyolites, or with breccias and fanglomerates sourced from the nearby metamorphic basement. The fanglomerates interfinger with coarse sandstones or fine shale sequences, as well as with ignimbritic rhyolites and occasionally with basalts.

The stratigraphy of the Permo-Carboniferous continental deposits (Carapelit Formation) of North Dobrogea includes a lower member of grey alluvial fan - alluvial plain sediments (loosely ascribed to the Carboniferous), unconformably overlying older metamorphic basement or Silurian-Lower Devonian sediments; the succession continues upsequence with red beds (0-900 m thick), overlain in turn by an upper volcanosedimentary member (Oaie, 1986; Seghedi & Oaie, 1986, 1995 b; Seghedi *et al.*, 1987) (Fig. 4).

In both the Moesian and Scythian Platforms, the Permian overlies older Paleozoic sequences (Fig. 5); the Permo-Triassic boundary is often difficult to trace, since both Permian and Triassic sediments show Germanic facies development. Evaporites occur in red beds devoid of carbonate sediments, suggesting a desert or coastal sabkha environment. In the Moesian Platform, red continental deposits prevail, but recent micropaleontological evidence suggests the presence of shallow-marine, Lower Permian carbonate facies in some parts of the platform (Pana, 1997) (Figs 1, 5). The red beds are interbedded with products of bimodal volcanism, with rhyolitic and basaltic rocks prevailing in the Moesian Platform, and the basalt-trachyte association well represented in the eastern part of the Scythian Platform.

PALEONTOLOGICAL DATA

Paleontological data were recorded for the Resita and Pui basins of the Getic Nappe, the Sirinia Basin of the Danubian units, the Codru-Biharia Nappes of the Apuseni Mountains, and the Moesian and Scythian Platforms. No paleontologi-

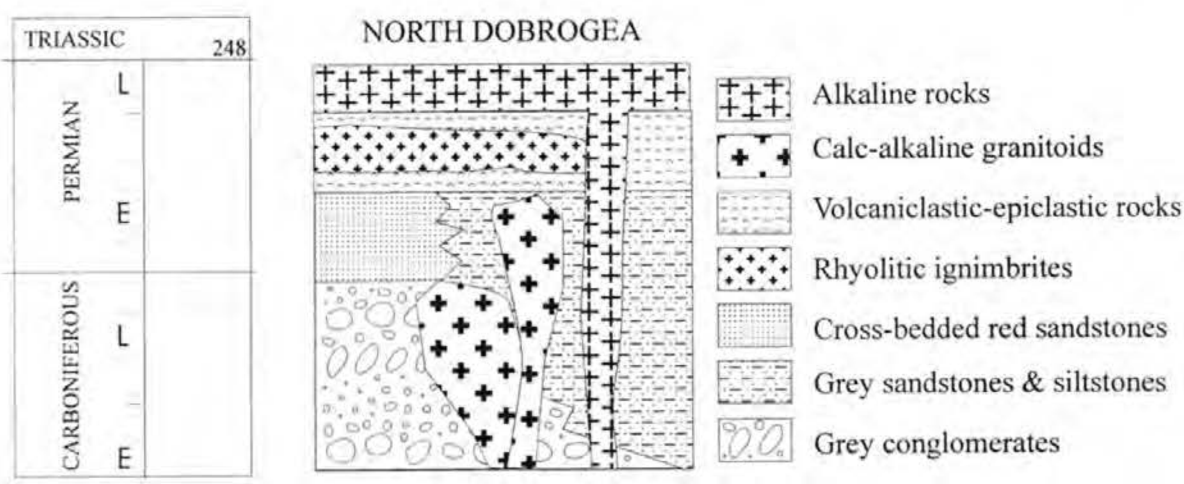

Fig. 4 - Log of the Carapelit Formation, North Dobrogea.

cal evidence has been recorded so far for the Presacina Danubian Basin of the South Carpathians, the Bihor Autochthon or the Carapelit Basin in North Dobrogea.

The first paleobotanical data from the Resita Basin were recorded by Stur (1870) and Telegd (1890), citing various Permian and Upper Carboniferous megaflora taxa. These taxa were later cited by Schreter (1910), Bitoianu (1973, 1974, 1987, 1988), Antonescu & Nastaseanu (1976) and Dragastan *et al.* (1997). Popa (1999) undertook a taxonomic revision and has established for the first time the majority of the Permian megaflora taxa of the Ciudanovita Formation: *Calamites* sp., *Annularia* cf. *stellata*, *Annularia* cf. *sphenophylloides*, *Asterophyllites longifolius*, *Sphenophyllum oblongifolium*, *Autunia conferta*, *Autunia naumannii*, *Arnhardtia scheibei*, *Lodevia suberosa*, *Gracilopteris bergeronii*, *Rhachyphyllum schenkii*, *Neuropteris* cf. *cordata*, *Neuropteris* sp., *Odontopteris* sp., *Pecopteris polymorpha*, *?Linopteris* sp., *Cyclopteris* sp., *Pecopteris* cf. *polymorpha*, *Taeniopteris* sp., *Alethopteris zeilleri*, *Cordaites principalis*, *Walchia piniformis*, *Ernestiodendron filiciformis*, *?Otovicia* sp., *Carpolithes* sp. *A* and *Carpolithes* sp. *B*. For the Gîrliste Member, the paleoflora are diverse and well preserved, represented by pteridophytes, pteridosperms and conifers, recording some Late Stephanian taxa as well as Early Permian taxa. For the Lisava Member, the diversity of the flora decreases substantially, with only Walchiaceae conifers occurring.

The palynological associations of the Permian deposits of the Resita Basin (Beju, 1970; Antonescu, 1980; Antonescu & Nastaseanu, 1976), record an early assemblage with *Florinites*, within the lowermost part of the Gîrliste Member (black sediments), and an upper assemblage with *Potonieisporites*, in the upper part of the Gîrliste Member and within the Lisava Member (red beds).

The first assemblage, with *Florinites*, is dominated by *Florinites* div. sp. (*F. circularis*, *F.* cf. *junior*, *F.* sp., 60-90% of the assemblage), followed by *Potonieisporites novicus* and *P. bharadwaji* (8-15%) or *Reticulatisporites facetus*, *Platysaccus papillionis*, *Alisporites* div. sp. (*A. aequus*, *A. saarensis*, *A.* sp.), *Pityosporites* sp., with less than 1-2% (Beju, 1970, in the Gîrliste area). Antonescu & Nastaseanu (1976) cited the same assemblage from Vidra Valley (Gîrliste Member), correlating the zone with the zone A1 of Doubinger (1974). They also cited S. Luta (in Antonescu & Nastaseanu, 1976) who described a Lower Permian assemblage with *Urospora kosankei*, *Raistrikia microhorrida*, *Complexisporites chaloneri*, *Cordaitina rotata*, *C. uralensis*, *Florinites similis*, *F. volans*, *P. novicus*, *Vittatina simplex* and *Guthorlisporites verus* from the Secu-Lupac area.

The second assemblage, dominated by *Potonieisporites* div. sp. (*P. novicus*, *P. bharadwaji*, 30-70%), is also represented by *Florinites* div. sp. (*F. circularis*, *F.* cf. *junior*, *F.* sp., 15-30%), *Platisaccus papillionis*, *Alisporites* div. sp. (*A. aequus*, *A. saarensis*, *A.* sp., 3-10%), *Pityosporites* sp. (3-8%) and *Striatoabietites* sp. (Beju, 1970, in the Gîrliste zone). To this assemblage was added *Crucisaccites* sp. and *Vesicaspora wilsonii* by Antonescu & Nastaseanu (1976, from Vidra Valley), these authors correlating the assemblage with the zone A2 of Doubinger, and later, (Antonescu, 1980) *Leiotriletes* sp., *Verrucosisporites* sp., cf. *Concetricisporites* sp., *Halletheca reticulata* and cf. *Schopfipollenites* sp. in samples from the Jitin-Ciudanovita Veche Valleys. These assemblages were described from the basal or terminal sequences of the Gîrliste Member. Furthermore, from the median sequence of the Lisava Member (within the Permian red beds) Antonescu & Nastaseanu (1976) identified an assemblage with cf. *Punc-*

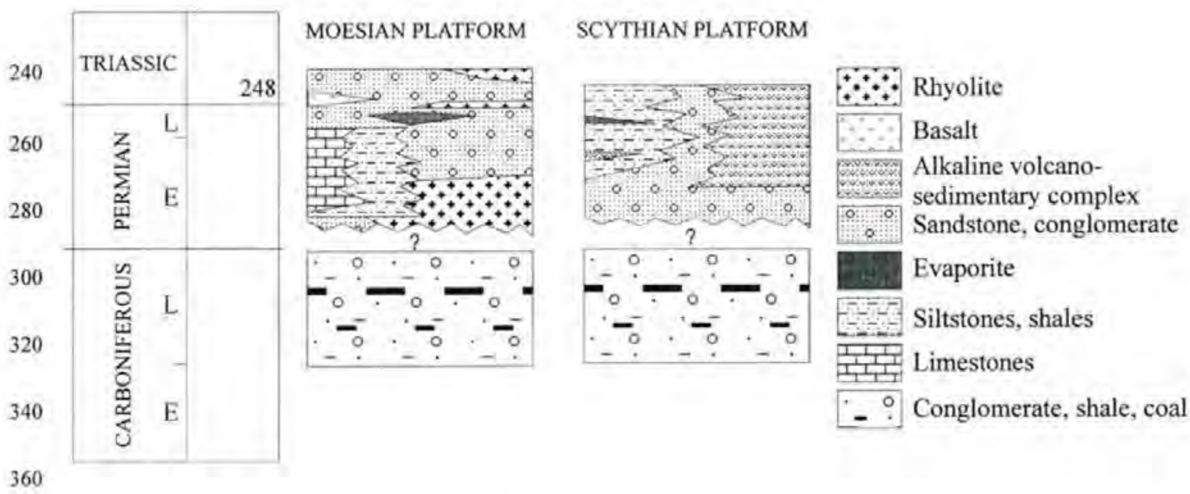

Fig. 5 – Paleozoic deposits of the Moesian and Scythian Platforms, compiled from Paraschiv (1979), Pana (1997) and Neaga & Moroz (1987).

tatisporites sp., *Granulatisporites* sp., *Lophotriletes* sp., cf. *Reticulisporites* sp., cf. *Neoraistrikia* sp., *Knoxisporites* sp., *?Speciosporites* sp., *Guthorlisporites* sp., *Potonieisporites* div. sp. (*P. novicus*, *P. bharadwaji*, *P.* cf. *microdens*, *P. neglectus*), *Florinites* sp., *Crucisaccites* sp., *Vittatina costabilis*, *Protohaploxypinus*, *Striatopodocarpites* sp., *Hamiapollenites* sp., *Ginkgocycadophytus* sp., cf. *Gardenasporites* sp., *Platysaccus* sp., *Alisporites* cf. *aequus* and *Vesicaspora wilsonii*. This is the uppermost palynological assemblage described within the red beds facies and it is considered Lower Permian ("Autunian") in age (zone A2 *sensu* Doubinger, 1974). In a paper discussing the age of the red beds of the Resita Basin, Antonescu (1980) attributed at least an Early Autunian age to the microflora from the basal sequences, but not excluding younger ages (Late Autunian?) for the upper sequences of the red beds, which lack paleontological evidence.

The fauna of the Resita Basin includes fish remains as *Palaeoniscus duvernoy* (Eufrosin, 1957) and the bivalves *Carbonicola carbonaria* and *Anthracomya* cf. *thuringensis*.

At Cioclovina, in the Pui area, Stilla & Luta (1968) mentioned and figured *Azonomonoletes vulgaris*, *Leiotriletes gulaferus*, *Punctatisporites* sp., *P. obliquus*, *P. spathulatus*, *Azonotriletes* cf. *tuberculatus*, *A. microrugosus*, *A.* cf. *rezistens*, *Laevigatisporites* sp., *Pemphygaletes minor*, *Reticulatosporites facetus*, *Punctatisporites marattioides*, *Cyclogranisporites leopoldi*, *Cycadopites* sp., *Vitrisporites* cf. *signatus*, *Ephedripites primus* and *Pityosporites* sp. This assemblage was considered Early Permian in age.

The megaflora of the Povalina Formation in the Sirinia Basin was cited by Telegd (1890) and Raileanu (1953). As their material is not yet revised, their lists include old designations (*Hymenophyllites semiallatus*, *Odontopteris obtusiloba*, etc.); however, *Walchia piniformis* and *Cordaites principalis* were mentioned.

Antonescu (1980) recognised from Sirinia well no. 22766/586 (Berzasca area), within the Povalina Formation, a palynological assemblage with *Leiotriletes* sp., *Dictyophyllidites* sp., *Calamospora breviradiata*, *C. mutabilis*, *Granulatisporites* sp. 1, 2, *Nigrisporites nigritellus*, *Punctatisporites obliquus*, *P.* sp., *Cyclogranisporites* cf. *palaeophytus*, *C.* sp., *Lophotriletes scotii*, *Knoxisporites* cf. *glomus*, *Crassispora* cf. *kosankei*, *Laevigatisporites* cf. *vulgaris*, *L. desmoinesensis*, *Spinosporites exiguus*, *Florinites* spp., cf. *Potonieisporites* sp., *Vittatina* sp., *Vesicaspora wilsonii* and *Ginkgocycadophytus* sp. The assemblage is dominated by *Laevigatisporites* and *Punctatisporites obliquus*, and it is considered Late Stephanian - Early Permian in age.

Faunal remains from the Sirinia Basin consist of the fresh-water bivalves *Carbonicola carbonaria*, *Anthracomya thuringensis* and the ostracod *Estheria* sp. (Raileanu, 1953).

The megaflora recorded for the Codru-Biharia Nappe System are very scarce, represented only by rare rachises, which are impossible to identify. Silicified woods, collected from the red bed sequences, were described by Arabu (1941) as *Dadoxylon* sp. Matyasi (1998) recorded silicified wood fragments identified by Popa (thanks to J. Galtier from Montpellier, France) as *Dadoxylon* of type III *sensu* Doubinger & Marguerier, 1975; they may represent a marker for a possible Late Permian age for the above-mentioned red-beds.

The palynological assemblage identified in the Codru Nappe (at Scarisoara) is represented by *Calamospora microrugosa*, *Turrisporites pyramidalis*, *Verrucosporites* sp., *Florinites* sp., *Cycadopites* sp., *Vittatina* sp., *Azonotriletes* cf. *nodosus*, *Zonotriletes* cf. *anubilus*, *Azonoletes similis*, *Stenozites compactus* and *S.* cf. *bulbiferus* (Visarion & Dimitrescu, 1971). This assemblage is related to a Late Carboniferous - Early Permian time interval.

The presence of ichnogenous *Planolites* is suggested to support an Early Permian age of the bioturbated sandstones from the Apuseni Mountains, as this ichnogenus is confined to the Lower Permian (Rotliegendes) from the Thuringerwald, in Germany, and widespread within coeval formations in continental facies from the Sudets (Poland) and West Carpathians (Slovakia) (Brustur, 1986, 1997). Moreover, identification of the resting trace of a primitiv aquatic amphibian of *Diplocaulus* type (*Hermundurichnus patruliusi*) within the bioturbated sandstones from the Finis Nappe (Padurea Craiului Mountains) enables their correlation with the Lower Permian from Thuringerwald (Brustur, 1997).

Within the central-eastern part of the Moesian Platform, paleontological data from wells Ileana, Hirlesti, Peretu, Peris and Amara indicate Early Permian marine environments (Pana, 1997). The author described Permian species of several foraminifer groups: Textulariina, Miliolina, Lagenina, Stafellidae (nine species), Ozowainellidae (six species), Schubertalidae (three species), Neoschwageridae, Earlandinidae, Nodosinellidae, Corallinellidae, Dagmaritinae, Louizettitinae. *Medlicottia* sp. was also cited. Among the conodonts, *Gnathodus defectus* and *Spathognathodus whitei* were identified in the Lower Permian, and *Anchirognathus typicalis*, *Neospathodus divergens* and *N. profundis* in the Upper Permian (Pana, 1997). As for ostracods, Pana found *Healdia* aff. *axensis*, *Coronakirbia fimbriata* for the Lower Permian and *Permiana oblonga* for the Upper Permian, among many other ostracod species (*Sishaella*, *Bythocypris*, *Tomiella*, *Iniella*, *Kirkybia*, etc.).

Paleontological and palynological studies of the red beds from the Scythian Platform identified a phyllopode association with *Pseudoestheria*, as well as spores indicating a Permian age (Kaptan & Safarov, 1965, 1966). Geochronological data show ages of 290-248 Ma for the associated volcanic rocks (Neaga & Moroz, 1987). However, as they are K-Ar ages, they represent cooling ages.

SEDIMENTOLOGY

Sedimentological studies on the Permian deposits of the South Carpathians and Apuseni Mountains are still in progress. The red-bed sequences from the Resita Basin (Lisava Member) and the Bihor Mountains show frequent cross-bedding, and their depositional features generally indicate alluvial to fluvial environments. The Lower Permian sequence of the Resita Basin coarsens upwards (Stanoiu *et al.*, 1996).

In North Dobrogea, the lower terrigenous members in-

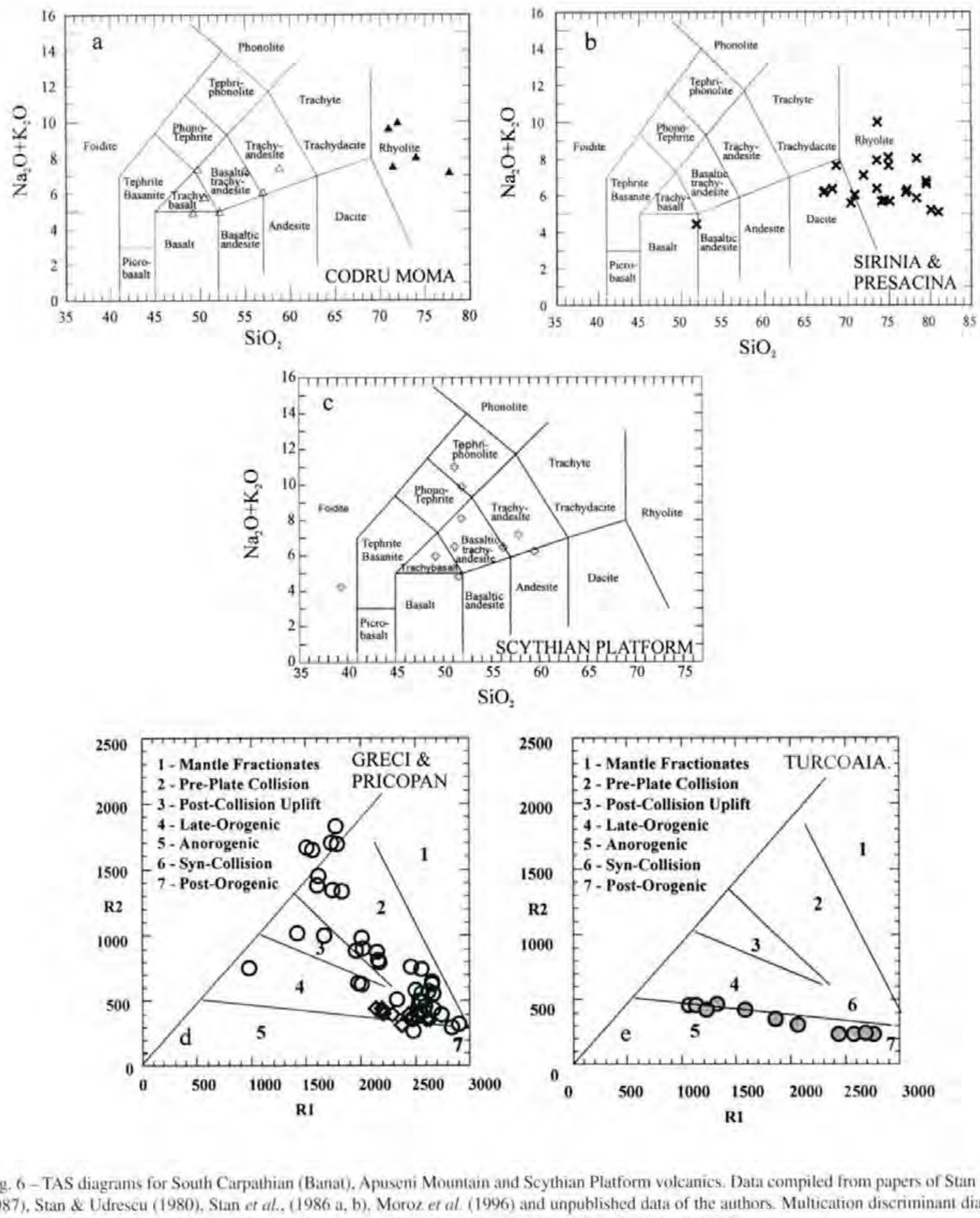

Fig. 6 – TAS diagrams for South Carpathian (Banat), Apuseni Mountain and Scythian Platform volcanics. Data compiled from papers of Stan (1984, 1987), Stan & Udrescu (1980), Stan *et al.*, (1986 a, b), Moroz *et al.* (1996) and unpublished data of the authors. Multication discriminant diagrams of Greci and Turcoaia massifs from North Dobrogea (modified from Tatu, 1999 and Tatu & Seghedi, 1999).

clude alluvial-fans to alluvial plain clastic wedges, with fanglomerates dominating the coarse members and sandstone-siltstone cycles in the flood-plain deposits (Seghedi & Oaie, 1986). Red beds make up a thick, upward-coarsening sequence, deposited by a sandy braided river with fluctuating discharge, showing upward progradation of coarse, longitudinal bar deposits over sand dunes with planar cross-beds (Oaie, 1986). Sandstone petrography suggests that the onset of red-bed deposition was related to a major climatic change, switching from a warm and humid climate which prevailed during alluvial fan sedimentation, to an arid, dry climate, controlling red-bed accumulation. The thick volcaniclastic successions from the upper part of the Carapelit Formation consist of superimposed cycles of pyroclastic deposits and coarse rhyolitic epiclastic sequences. Volcaniclastic rocks are dominated by large volumes of ignimbritic rhyolites (up to 1000 m in thickness), interbedded with airfall tuffs and base surge deposits, displaying the geometry of superimposed flow units (Seghedi *et al.*, 1987). Vertical facies associations suggest that the style of deposition was controlled by intermittent volcanic eruptions. Sedimentological, petrographical and mineralogical evidence reveals that accumulation of the Carapelit Formation was controlled by two major factors: a tectonically active source area, supplying metamorphic rocks, granites and earlier Paleozoic sediments and an active volcanic source, delivering large amounts of feldspars (mostly plagioclase feldspars) and volcanic lithoclasts (Seghedi & Oaie, 1986, 1994; Oaie, 1986; Seghedi *et al.*, 1987).

In the eastern part of the Aluat-Sarata half-graben from the Scythian Platform, several fining-upwards cycles are superimposed in the 1685 m-thick column of Permian volcanosedimentary sequences pierced by borehole 1 Furmanovka (Moroz, 1984; Neaga & Moroz, 1987). In the western part of the half-graben, the distribution of the coarse fanglomerates suggests sedimentation controlled by activity along the northern boundary fault (Fig. 1). Fanglomerates interfinger and grade upwards to a finegrained sequence of siltstones and mudstones, with thin, discontinuous layers of gypsum and anhydrite.

MAGMATISM

Permian magmatism was characterised by bimodal volcanism in the South Carpathians and Apuseni Mountains, as well as in both the Moesian and Scythian Platforms; volcanosedimentary sequences are typically developed in most areas, with the exception of the Resita Basin, where minor volcanism occurred. Granite intrusions took place only in the southern part of the Apuseni Mountains (Highis massif) (Tatu, 1998). In North Dobrogea a volcano-plutonic association with calcalkaline geochemistry is well represented, and the transition to alkaline magmatism probably occurred in the Late Permian, reflecting a change in geotectonic setting from compression to transtension.

Acid, rhyolitic volcanic and volcaniclastic deposits prevail in the Permian of the Danubian units (Banat). Rhyolitic and dacitic volcanics and volcaniclastic-epiclastic successions are variously interbedded with continental red beds, while thick bodies of ignimbritic rhyolites occur at the top of the Danubian sequences (Stanoiu & Stan, 1986; Stan *et al.*, 1986 a, b). The thickness of the volcanic sequences increases westwards. The volcanism was bimodal (Fig. 6 a), as basic dykes intrude the red beds in some areas (Pop, 1997), and basic flows are exposed below red beds from the right bank of the Danube (Serbia) (Grubic *et al.*, 1997). Trachytic rocks occur southward (A. Grubic, oral comm., 1997). An intraplate setting is suggested for these rocks, based on geochemical data.

In the Codru Moma Mountains (western Apuseni), the volumes of both ignimbritic rhyolites and basic rocks decrease eastwards (Dimitrescu *et al.*, 1977; Bleahu *et al.*, 1979, 1981, 1985). Thin dykes and elongate rhyolitic bodies cutting the older basement (Bleahu *et al.*, 1984) suggest that volcanic centres, aligned in a NNW-SSE direction, were located along the present western border of the massif. Several basic dykes, elongated in the same direction, may represent the feeder channels for the basic flows interbedded in the middle part of the Permian sequence. In the Bihor Mountains, the ignimbrite eruptions took place in the upper part of the Permian sequence (Dimitrescu *et al.*, 1977). These acid rocks, mostly rhyolites, occur as pyroclastic flows and lava flows, but also as tuffs (Dimitrescu *et al.*, 1973; Dimitrescu, 1975; Stan, 1983, 1984, 1987). Basic volcanic rocks include basalts and basaltic andesites as pillow lavas, minor pyroclastic sequences and dykes (Bleahu *et al.*, 1979, 1981,1985; Stan, 1987). Chemical analyses of the volcanic rocks reveal both the bimodality and the alkaline features of the basic rocks; the latter plot in the trachybasalt-trachyandesite field of the TAS diagram (Fig. 6 b). The bimodal character of the volcanism is explained by having two distinct magma sources (Stan, 1987).

In North Dobrogea, the volcano-plutonic association with a calcalkaline geochemistry is related to crustal convergence at the end of the Hercynian Orogeny. For the volcanosedimentary member of the Carapelit Formation, overall volcanological features indicate that the calcalkaline volcanism was subaerial, with calderas and plinian eruptions, and that volcanic products accumulated in both subaerial and subaqueous environments (fluvial and lacustrine) (Seghedi *et al.*, 1987).

Field relationships indicate that two major phases of granite emplacement occurred in North Dobrogea, one preceding and the other post-dating the deposition of the Carapelit Formation or at least its lower and middle members (Rotman, 1917; Mirauta & Mirauta, 1962). However, the age of the granites is not well constrained, since no reliable geochronological data exist. Younger generation intrusives thought to have been emplaced during the Early Permian represent a highly differentiated I-type calcalkaline suite, ranging from dioritic and gabbroic facies to leucogranites, but dominated by biotite-hornblende granodiorites and tonalites (Seghedi *et al.*, 1994 a). The suite of the Greci Massif was emplaced in lower members of the Carapelit Formation as high level, high temperature intrusives, with both cross-cutting and partly conformable contacts, producing large contact-metamorphic aureoles and local garnet-pyroxene skarns. Rocks are rich in xenoliths of hornfels inferred to have been country rocks, as well as in various types of cognate xenoliths. The discriminant multicationic diagram (Fig. 6 d) suggests a mantle source for the basic end members and a deep or mid-crustal conditions for the genesis of granitic magma, as well as a syncollisional geotectonic setting (Tatu & Seghedi, 1999).

In the Late Permian, products of alkaline, intraplate volcanism were emplaced along the crustal faults bordering North Dobrogea. They form both subvolcanic bodies (granites, syenites, rhyolites of the Turcoaia – Cîrjelari lineament) in the south and basalt-trachyte dyke swarms in the north (Fig. 7) (Seghedi *et al.*, 1994 b; Tatu & Teleman, 1997). For the Turcoaia massif, a crustal magma source is indicated by the $^{87}Sr/^{86}Sr$ values (Pop *et al.*, 1985), suggesting that crustal anatexis occurred in a continental, intraplate tectonic setting (Fig. 6 e) (Tatu, 1999). The geochemistry of Permian magmatic rocks suggests that this transition from calcalkaline to alkaline magmatism reflected a change in tectonic setting from compressional to (trans)tensional.

In the Moesian Platform, volcanic products belonging to a bimodal basalt-rhyolite association, interbedded at various levels of the Permo-Triassic sequence (Savu & Paraschiv, 1985) are obviously related to several episodes of extension and rifting of the Moesian Platform.

Products of a typical continental within-plate bimodal volcanism of the alkali basalt-trachyte association accumulated in the two major half-grabens from the Scythian Platform (Aluat and Sarata – Tuzla). Alkaline basalt flows, trachytes, rhyolites and various pyroclastic and epiclastic products make up volcanic-volcaniclastic sequences that are up to 300 m thick, interbedded with thick, synrifting continental clastics. The Permian volcanism was subaerial and effusive, displaying a subalkaline geochemical signature (Moroz, 1984; Neaga & Moroz, 1987). Trachy-

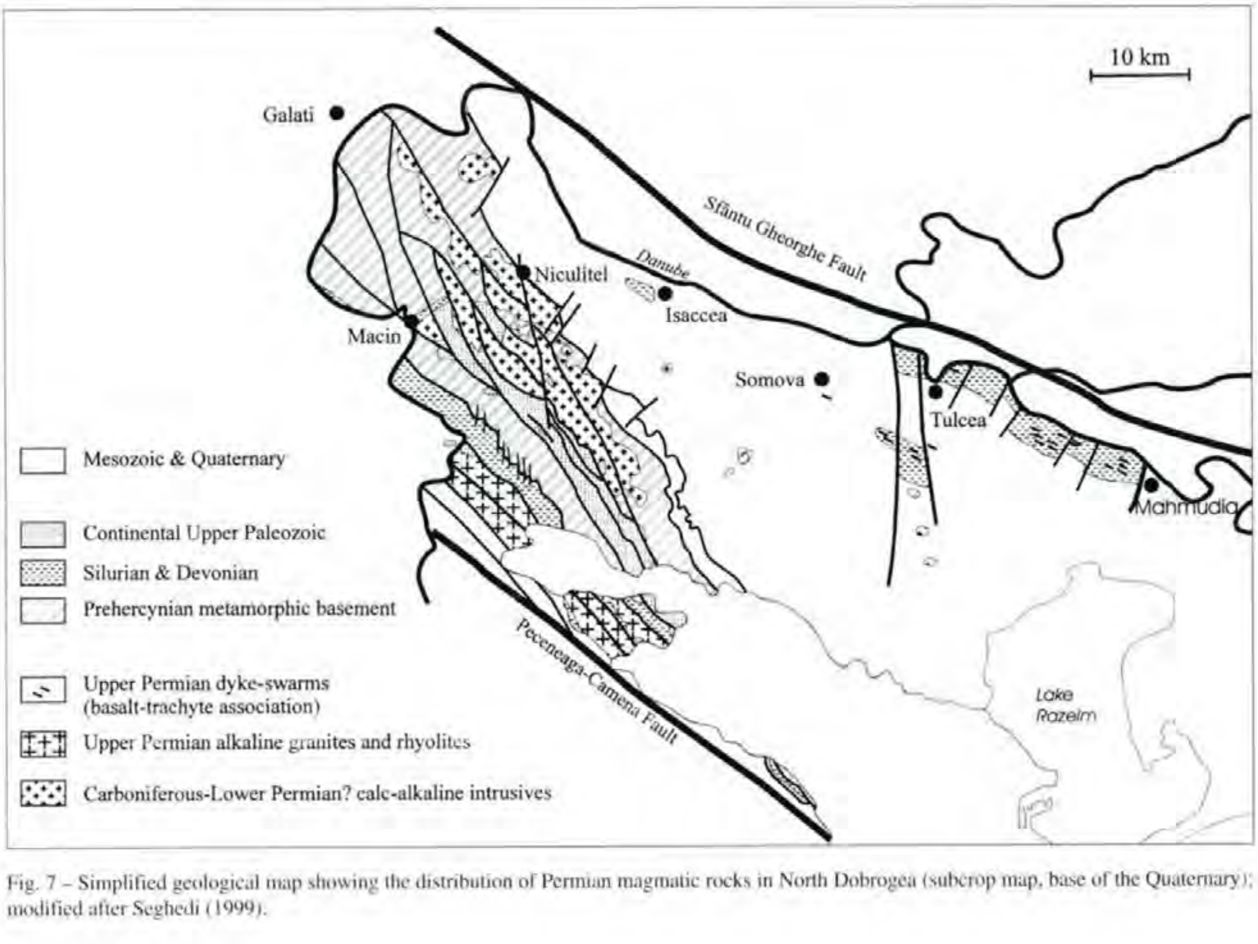

Fig. 7 – Simplified geological map showing the distribution of Permian magmatic rocks in North Dobrogea (subcrop map, base of the Quaternary); modified after Seghedi (1999).

basalts and trachyandesites prevail, most rocks showing a shoshonitic affinity (Fig. 6 c). Geochemical data suggest a great inhomogeneity of samples compared with the Codru-Moma volcanics. However, the intraplate setting is quite clear from the geological evidence.

CONCLUSIONS

The Permian deposits of Romania are mainly developed in continental facies, with molassic characteristics. The sedimentary record of most basins includes volcanosedimentary sequences. Red beds are the dominant facies, but the lower black shaly member of the Permian is present in the South Carpathians and Apuseni Mountains. Evaporitic sediments are associated with the red beds in both the Moesian and Scythian Platforms. In all areas, sedimentation took place mostly in alluvial fan, fluvial and lacustrine systems. Only in the northern part of the Moesian Platform do shallow-marine limestones occur. The South Carpathians (Getic Nappe and Danubian units) yield by far the most fossiliferous Permian deposits in Romania, the fossil remains being represented by both flora (compressed macroflora, microflora) and fauna (ganoid fishes, ostracods, bivalves). The Apuseni Mountains include deposits yielding flora (silicified woods, microflora), while North Dobrogea has no fossils recorded so far. The Moesian and Scythian Platforms include faunal remains.

The Permian volcanism was bimodal. The basalt-rhyolite association typically occurs in the South Carpathians, the Apuseni Mountains and the Moesian Platform, while the basalt-trachyte association is found in North Dobrogea and the Scythian Platform.

An extensional tectonic setting, related to Late? Permian rifting, is suggested by both field evidence and geochemical features of the magmatic rocks from the South Carpathians, Apuseni Mountains, Moesian and Scythian Platforms. For North Dobrogea, both magmatic associations and their geochemical features illustrate a transition from a calcalkaline post-collisional setting to a transtensional setting related to the collapse of the overthickened Hercynian crust, which was displaced along major wrench faults.

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UPPER PERMIAN TYPE SECTIONS OF THE EAST EUROPEAN PLATFORM AND THEIR CORRELATION

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Key words – Ufimian; Kazanian; Tatarian; stage; boundaries; fauna; flora; paleomagnetic zones.

Abstract – The authors describe zonal subdivision of the Upper Permian (Ufimian, Kazanian, Tatarian) on the East European Platform, and suggest stratigraphic levels for these stages for global correlation.

Parole chiave – Ufimiano; Kazaniano; Tatariano; piano; limiti; fauna; flora; zone paleomagnetiche.

Riassunto – Gli autori descrivono la suddivisione zonale del Permiano superiore (Ufimiano, Kazaniano, Tatariano) inerente alla Piattaforma Est-Europea, e suggeriscono livelli stratigrafici relativi a questi Piani utili per correlazioni globali.

INTRODUCTION

Upper Permian sediments occur on the East European Platform from the Middle Urals in the south to Novaya Zemlya in the north. Ufimian type sections are located in the Kama region near the town of Perm, Kazanian type sections in the Volga and Kama rivers region, and Tatarian type sections in the basins of the Volga, Vyatka, Sukhona and North Dvina rivers (Fig. 1).

CHARACTERISTICS OF THE BASIN

During Late Permian times, this vast basin occupied an area of over 2 million km² and ranged from 8°N to 32°N (Fig. 2). Numerous hiatuses/discontinuities of the Volga-Urals Permian sections are nowadays widely discussed. Some hiatuses are associated with the dispersion of sedimentation. The hiati are evenly distributed over all sedimentary basins, including marine ones, and reflect the alternation between continuous and discontinuous sedimentation. There are hiatuses caused by sedimentary cyclic recurrence. At the rhythmic boundaries, coarse basal sediments overlap calcareous formations, sometimes disrupting their structure and bedding. Such hiatuses do not distort the whole picture, provided there are more complete sections nearby.

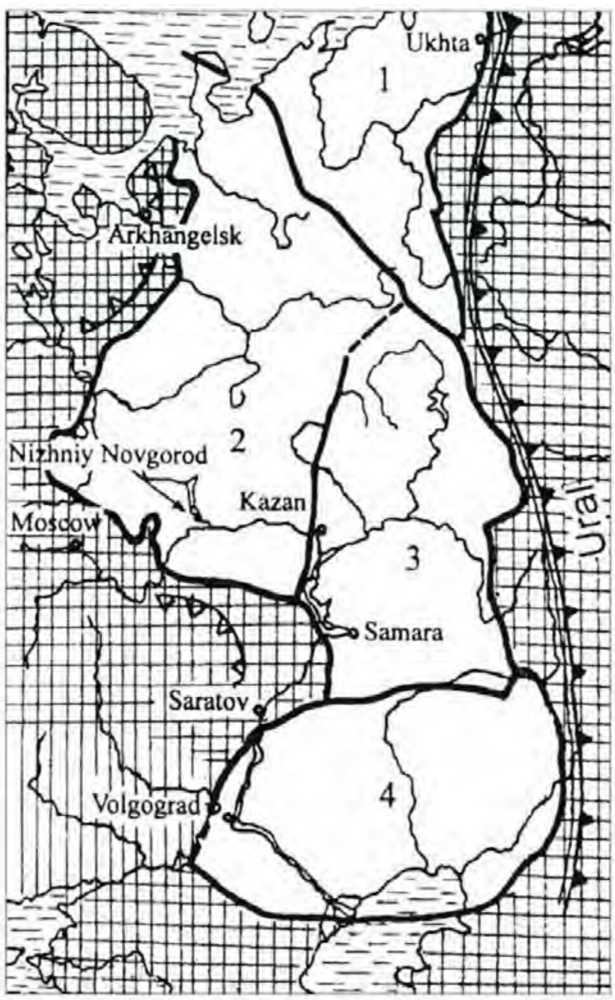

Fig. 1 – Late Permian sedimentary basins in European Russia.
1. Pechora Basin; 2. Dvina Basin; 3. Volga-Urals Basin; 4. Caspian Basin.

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There can be marginal disconformities associated with the oscillation of the sedimentary basin's margins. Thus, the Permian in the east of the basin overlies the Kungurian, but in the west the Artinskian, Sakmarian and Asselian. The Kazanian in the east conformably overlies the Ufimian Sheshminian red beds, but in the west the Sakmarian and Asselian. As for the profile running from Kazan to Nizhny, basal layers of the Tatarian Urzhum horizon near Kazan overlie the Upper Kazanian 'transitional' series, farther to the west the Podluzhnik series, then the Shikhany series, and so on up to the Lower Kazanian. In some sections, the Induan overlies the Severodvinian horizon. This unconformity is clearly recognised by the lack of some Permian paleomagnetic zones. It can be concluded that hiatuses that substantially distort stratigraphic sequences are developed at the margins, and are associated in the east with water erosion, and in the west with movement of the basin's margin caused by the development cycles. More complete sections can be found in central areas of the sedimentary basin.

UPPER PERMIAN FAUNA

We stick to a conventional two-part division of the Permian that globally indicates mostly marine sedimentation in Early Permian times, and transitional sedimentation in Late Permian times, while the Triassic is known to have mostly continental sediments. The Upper Permian is represented by marine, transitional and continental facies, and characterised by **normal marine** radiolarians, foraminifers, bryozoans, corals, bivalves, gastropods, cephalopods, nautiloids, ostracods, ichthyloids and conodonts, **freshwater** bivalves, ostracods, ichthyloids, stromatoliths and charophytes, and **continental** terrestrial vertebrates, macroflora and miospores. At the Permian Symposium in 1998, we had the opportunity to present detailed paleontological characteristics of the Ufimian, Kazanian and Tatarian (Proceedings..., 1999). In this report, we would like to examine several stratigraphic levels that can be used for global correlation.

In the Permian Cis-Urals region, the Upper Permian sediments (Solikamian horizon of the Ufimian) occur on gypsiferous carbonates of the Kungurian Irensky horizon, and contain stunted bivalves (Silantiev, 1998, p. 20). As a parastratotype of the Lower/Upper Permian boundary, we would like to propose a section on the Kazhim river

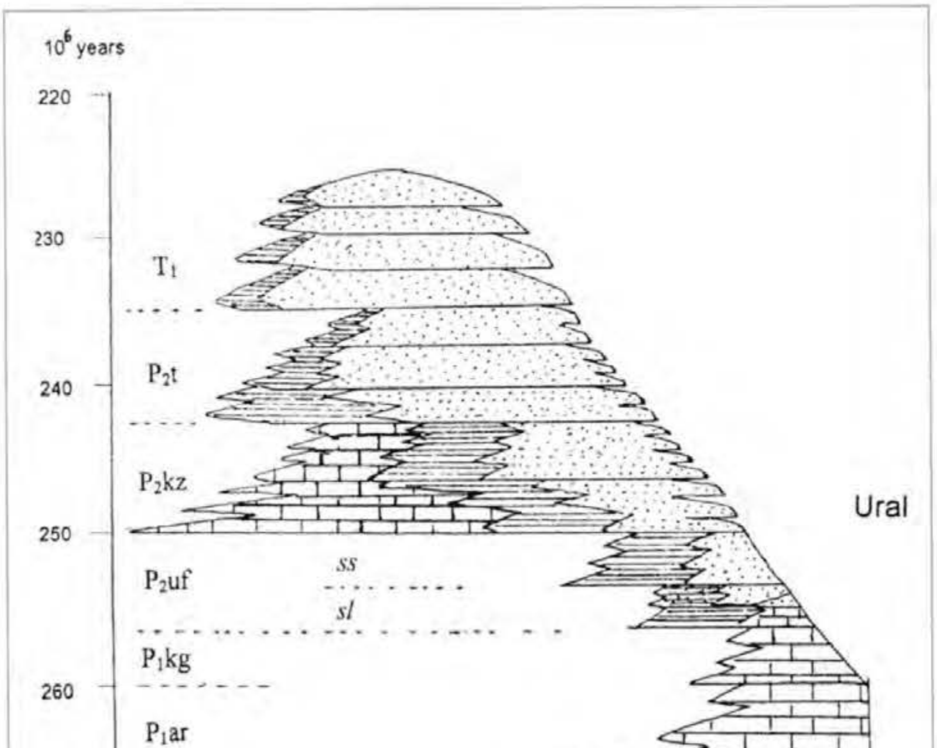

Fig. 2 – Permian sedimentary basin of the Volga-Ural region (Burov *et al.*, 1998).

Abbreviations: P_1 Lower Permian; as Asselian; s Sakmarian; ar Artinskian; kg Kungurian. P_2 Upper Permian; uf Ufimian (*sl: sensu lato; ss: sensu stricto*); kz Kazanian; t Tatarian. T_1 Lower Triassic.

since it has been described in detail using marine foraminifers, bryozoans, brachiopods, bivalves and gastropods (Biota…, 1998, Fig. 25). A gradual transition from deep-marine to shallow-marine facies is reflected by the change in faunal taxa at the boundary between the Kozhimian and Kozhimrudnitskian series. The boundary is observed throughout the basin, and can serve as a basis for inter-regional correlation. Moreover, Grunt (Biota…, 1998) has reported that brachiopods *Kochiproductus, Yakovlevia, Spirifirella* of the Solikamian Kozhimorudnitskian horizon have also been found at the Jisu-Hongora section of Inner Mongolia. Numerous *Monodixodina* associated, according to the viewpoint of Leven (pers. comm.), with *Armenia* and highly developed *Misselina* and *Parafusulina* which are peculiar to the Kubergandinian have also been found at this Mongolian site. This provides an adequate basis for the correlation of the Solikamian horizon of the East European scale with the Kubergandinian of the Tethyan scale. In our monograph (Biota…, 1998), we provide a detailed description of the Lower/Upper Permian boundary. The Ufimian Solikamian and Sheshminian in the Cis-Urals near the town of Perm are mostly characterized by shallow-marine sediments with marine and non-marine bivalves (Silantiev, 1998, p. 28). Marine bivalve sediments also contain foraminifers. The first foraminifer level is characterised by *Pseudonodosaria, Rectoglandulina* and *Langella*, and corresponds to the Solikamian horizon, while the second level corresponds to the Sheshminian horizon which is, according to Kotlyar (pers. comm.), characterised by a first appearance of *Dentalina* and *Lingulina*, with new species of *Nodosaria, Geinitzina* and *Frondicularia* corresponding to the Guadalupian Roadian of North America.

The Kazanian of the Volga-Urals basin is represented by marine, transitional and continental facies with diverse fauna and flora. Foraminifers from the region were studied by G. Pronina (1998, p. 163). They belong to 126 species of 43 genera . The third foraminifer level corresponds to the Lower Kazanian, the fourth to the Upper Kazanian, and the fifth to the Lower Tatarian. The Lower Kazanian contains *Frondicularia*, known from the Upper Ghizhiginsky horizon of the Omolon Massif and Polish Zechstein (Pronina, 1998). The Upper Kazanian also contains taxa known from other Boreal regions such as Omolon, Taimyr, Novaya Zemlya and Kolguev.

More than 35 brachiopod species of 20 genera of the Paleozoic taxa are known from the Kazanian stratotypes of the Volga region (Fig. 3).

Most characteristic of this complex are spiriferids represented by such boreal genera as *Licharevia, Blasispirifer, Odontospirifer* and *Tumarinia* (Gubareva, 1998, p. 30). In topotype sections along the Vyatka River, there are also *Permospirifer* and *Kaninospirifer*. Besides a single find of *Compressoproductus sp.*, the stratotype contains abundant productoids *Cancrinella, Terrakea, Globiella* and *Aulosteges. Stenoscisma, Rhynchopora, Cleiothyridina, Pinegathyris, Baitugania, Crurithyris* and *Spiriferellina* are also found in abundance. These brachiopods allow the correlation with Zechstein of Germany, Novaya Zemlya (New Land), northeastern Russia, Mongolia, China, and also Australia and New Zealand, within a single boreal area

BRACHIOPODS FROM THE PERMIAN SOLIKAMIAN HORIZON OF THE KAMA REGION

- *Lingula: L. orientalis* Gol.
- *Cancrnella: C. cancrini* (Vern)
- *Cleiothiridina: C. pectinifera* (Sow.)
- *Phrycodothyris: Ph. rostrata* (Kut.)

BRACHIOPODS FROM THE KAZANIAN STRATOTYPICAL AREA

- *Lingulina: L. orientalis* Gol., *L. credneri* Geinitz, *L. lawrskii* Netch.
- *Crania: C. orientalis* Netsch.
- *Orbiculoidea: O. konincki* (Gein.)
- *Streptorhnchus: S. pelargonatus* Schlot.
- *Cancrinella: C. canarini* (Vern.)
- *Terrakia: T. hemisphaeroidalis* (Netsch.)
- *Globiella: G. hemisphaerium* (Kut.)
- *Aulosteges: A. wangwnheimi* (Vern.), *A. horrescens horrescens* (Vern.), *A. horrescens sokensis* Grig., *A. fragilis* (Netsch.)
- *Compressoproductus: C.* sp.
- *Stenoscisma: S. superus* (Vern.), *S. globulina* (Phill.)
- *Rhynchopora: R. geinitzina* (Vern.), *R. globulina* (Phill.)
- *Cleiothyridina: C. pectinifera* (Sow.)
- *Pinegathyris: P. royssiana royssiana* (Keys.), *P. stuckenbergi* (Netsch.)
- *Baitugania: B. netschaevi* Grunt
- *Licharewia: L. rugulata* (Kut.), *L. stuckenbergi* (Netsch.), *L. schrencki* (Keys.)
- *Tumarinia: T. latiareata* (Netsch.)
- *Blasispirifer: B. multiplicoststus* (Netsch.)
- *Odontospirifer: O. subcristatus* (Netsch.), *O. parvula* (Netsch.)
- *Crurithyris: C. nucella* (Netsch.)
- *Spiriferellina: S. netschajewi* (E. Ivan.)
- *Beecheria: B. netschajewi* Grig., *B. angusta* (Netsch.),
- *B. elliptica* (Netsch.), *B. itiatubense* (Derby), *B. nikitini* (Netsch.)

Fig. 3 – Upper Permian brachiopods from the stratotypical area (Gubareva, 1998).

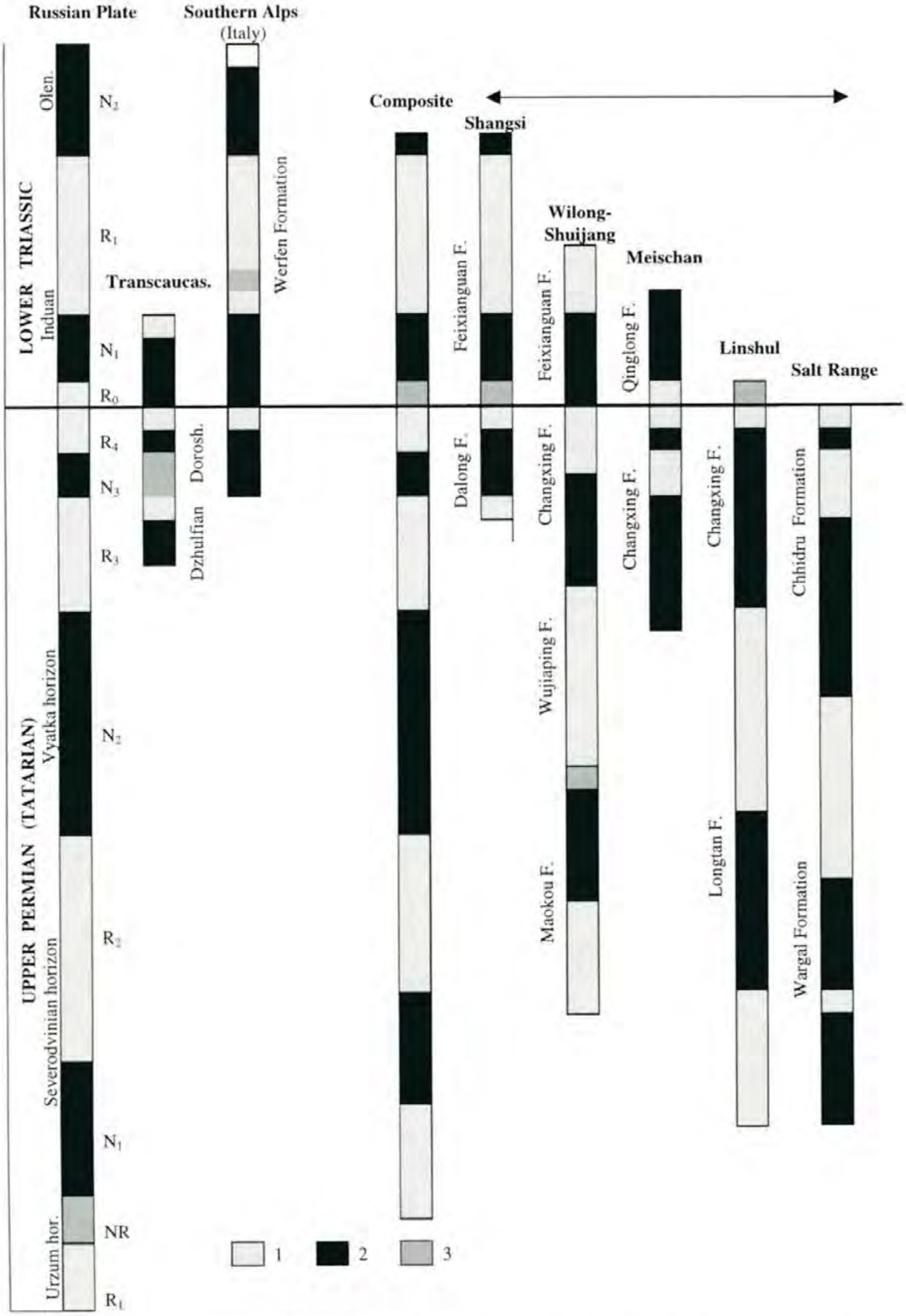

Fig. 4 – Scheme of paleomagnetic correlation of the Upper Permian Stratotype (the Volga-Kama region) with Tethyan sections. 1. reversed polarity magnetic zones; 2. normal polarity magnetic zones; 3. changing polarity zones.

(Gubareva, 1998). The complex can easily be recognised in the East European region where the Samaro-Tataro-Bashkir subprovince with the *Licharewia-Tumarinia-Blasispirifer* zone and the Kirov-Arkhangelsk subprovince with the *Licharewia-Tumarinia-Blasispirifer* + *Permospirifer* zones were outlined. At the same time, *Aulosteges* and *Compressoproductus* are known from the Tethyan sections of the northern Caucasus, and *Stenoscisma* and *Spiriferellina* from the Midian of the Eastern Tethys, allowing the correlation between various regions of the world.

The Kazanian basin, as a northward-opening intracontinental gulf, offered favourable living conditions for the development of bryozoans. The Kazanian association is characterised by 70 species of 40 genera, both cosmopolitan and endemic. In the Volga-Urals region, Lisitsin & Morozova (1998, p. 91) have studied 20 species of 16 genera. Eight of them are also found in the Kozhimian of the Pechora basin, nine in the Starostinian series of Spitsbergen, eleven in the Chandalazsky horizon of South Primorye, and ten genera in the Wordian Herster formation of northwest USA.

Nowadays, multi-element species of *Sweetina*, *Merrilina* and *Stepanovites* have been found throughout the Kazanian section. *Sweetina tritica* Wardlaw et Collinson in the Lower Kazanian allows correlation with the Upper Roadian and Lower Wordian. The Upper Kazanian of the East European Platform, according to *Stepanovites meyeni* and *Merrilina divergens*, corresponds to the Upper Wordian and Lower Capitanian.

An outstanding feature of the Late Permian Volga-Urals basin is a facies change from marine to continental, permitting co-occurrence of conodonts and macroflora within one section, although in different layers (Esaulova, 1998, pp. 206-208). Macroflora make possible the detailed zonal division and correlation of the whole Angarian region up to South Primorye (Esaulova, 1996, p. 470). It is noteworthy that the Chandalazsky horizon of the South Primorye has some elements of the

EASTERN EUROPE						TETHYS
				Tetrapod zones		
LOWER TRIASSIC			RYBINSKIAN		Thoosuches jakovlevi	LOWER TRIASSIC
LOWER TRIASSIC			VOKHMIAN		Tupilakosaurus wetlugensis	DORASHAMIAN
UPPER PERMIAN	TATARIAN	U	VYATKIAN	Scutosaurus	Archosaurus rossicus	DORASHAMIAN
UPPER PERMIAN	TATARIAN	U	VYATKIAN	Scutosaurus	Scutosaurus karpinskii	DZHULFIAN
UPPER PERMIAN	TATARIAN	U	SEVERO-DVINIAN	Scutosaurus	Proelginia permiana	DZHULFIAN
UPPER PERMIAN	TATARIAN	U	SEVERO-DVINIAN	Scutosaurus	Deltavjatia viatkensis	DZHULFIAN
UPPER PERMIAN	TATARIAN	L	URZHUMIAN	Titanophoneus	Ulemosaurus svijagensis	MIDIAN
UPPER PERMIAN	TATARIAN	L	URZHUMIAN	Titanophoneus	Estemmenosuchus uralensis	MIDIAN
UPPER PERMIAN	KAZANIAN	U	POVOLZHIAN	Titanophoneus	Estemmenosuchus uralensis	MURGHABIAN
UPPER PERMIAN	KAZANIAN	L	SOKIAN	Titanophoneus	Parabradysaurus silantjevi	KUBERGANDINIAN
UPPER PERMIAN	UFIMIAN		SHESHMINIAN	Titanophoneus	Parabradysaurus silantjevi	KUBERGANDINIAN
UPPER PERMIAN	UFIMIAN		SHESHMINIAN		Clamorosaurus noctumus	BOLORIAN

Fig. 5 – Correlation of the Upper Permian deposits of the Boreal regions using vertebrates (Golubev, 1998).

Kazanian phylladodermic floristic complex as well as the Tethyan goniatites *Timorites markevichi* from the Lower Capitanian.

The Tatarian is represented by shallow-marine and freshwater facies characterised by diverse and rapidly-evolving fauna and flora, allowing the detailed division using bivalves, ostracods, fish, macroflora and miospores. All the biostratigraphical zones are correlated with paleomagnetic ones (Gusev *et al.*, 1993). The lithology of the Tatarian allows a detailed study of the structure of geomagnetic field. The paleomagnetic data from the Boreal Volga-Urals basin and the Salt Range of Pakistan permit their correlation (Burov *et al.*, 1998, p. 263) (Fig. 4).

Golubev (1998, p. 57) indicated tetrapod zones in the Upper Permian continental facies of the East European Platform. In Late Permian times, Eurasia and Gondwana were separated by Tethys. Vast regressions resulting in land bridges allowed the correlation between the East European and Tethyan vertebrate zones (Fig. 5).

CONCLUSION

Thus, we believe that the Lower/Upper Permian boundary, the Kazanian, and the Kiama/Illawarra boundary of the Tatarian can all serve as a basis for the global correlation of the Upper Permian according to the East European scale.

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TRANSITIONAL PERMIAN-TRIASSIC DEPOSITS IN EUROPEAN RUSSIA, AND NON-MARINE CORRELATIONS

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Key words – Permian; Triassic; stratigraphic correlation; fossil flora; palynology; magnetostratigraphy.

Abstract – A recent discovery of a relatively complete transitional Permian-Triassic (Tatarian-Vetlugian) sequence in the Vologda region, European Russia, bears on the problem of non-marine PTB correlation. It shows a zone of reversed polarity at the base of the Vetlugian.
The plant megafossil assemblage of the transitional interval is dominated by Tatarian survivors, with a few conifer species with affinities to the Zechstein flora.
The palynological assemblage is of mixed Upper Permian-lowest Triassic aspect.
The megaspore assemblage contains *Otynisporites eotriassicus*, a zonal index species of the lowermost Buntsandstein, occurring also in the Tesero Oolite, Southern Alps, with conodonts *Hindeodus praeparvus* (Kozur, 1998), as well as in the Upper Guodikeng Formation of the Junggar Basin with *Dicynodon* and *Lystrosaurus* (Liu, 1994).
These occurrences are considered to mark a stratigraphic level corresponding to the conodont zone *Clarkina meishanensis* of the Meishan section, south China.

Parole chiave – Permiano; Triassico; correlazioni stratigrafiche; flora fossile; palinologia; magnetostratigrafia.

Riassunto – Il recente rinvenimento di una successione relativamente completa in corrispondenza della transizione tra il Permiano e il Trias (Tatariano-Vetlugiano) nella regione di Vologda (Russia europea) porta al problema di una correlazione relativa al limite P/T in ambiente continentale. Essa mostra una zona di polarità inversa alla base del Vetlugiano. L'associazione fossilifera a piante dell'intervallo di transizione è dominato da organismi superstiti del Tatariano, con poche specie di conifere affini alla flora dello Zechstein. L'associazione palinologica mostra un aspetto misto tra quella del Permiano superiore e quella inerente alla parte più bassa del Trias. L'associazione a megaspore contiene *Otynisporites eotriassicus*, un indice zonale della parte più bassa del Buntsandstein, che è anologamente presente sia nell'Oolite di Tesero, delle Alpi Meridionali, dove sono stati recentemente rinvenuti conodonti del tipo *Hindeodus praeparvus* (Kozur, 1998), e sia nella porzione superiore della Formazione di Guodiken, del Bacino di Junggar, che include *Dicynodon* e *Lystrosaurus* (Liu, 1994). Questi eventi sono considerati come indicatori di un livello stratigrafico corrispondente alla zona a conodonti *Clarkina meishanensis* della sezione di Meishan, in Cina meridionale.

INTRODUCTION

In European Russia the Tatarian deposits are unconformably overlain by the Lower Triassic Vetlugian Series. Paleomagnetic correlation indicates a hiatus at the boundary encompassing most of the Changhsingian-Dorashamian stages (Lozovsky & Esaulova, 1998). Equivalents of the upper Zechstein seemed likewise lacking in the transboundary Tatarian to Vetlugian sections. A find of *Lystrosaurus (Paralystrosaurus) georgi* (Kalan.) in the lower Astashikhinsk Member of the Vetlugian Series was taken as evidence of the lowermost Triassic age (Lozovsky, 1998). Our new data indicate that relatively complete sequences, supposedly continuous over the PTB, occur in the central part of the Permian–Triassic basin at about the Volga–Severnaya Dvina watershed (Krassilov *et al.*, 1999). This conclusion is based on the results of paleobotanical and magnetostratigraphic studies, supplemented by a few faunistic finds, in the Nedubrovo Section, Vologda region.

TRANSBOUNDARY SECTION AT NEDUBROVO

The Nedubrovo Section is exposed in a series of large outcrops on the left bank of the Kichmenga River (left tributary

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of the Yug River), northeast of Vologda City between the Glebovo and Nedubrovo riverside villages. Here the uppermost Tatarian (Vyatkian) variegate marls and clays with small carbonate nodules are overlain, at a sharp contact, by: (1) the basal Vetlugian cross-bedded sands with gravel and pebbles, up to 8 m thick; (2) reddish-brown micaceous clays and siltstones about 3 m thick; (3) alternating thinly-bedded grey (in the lower part), greenish-grey and purple siltstones and silty clays, 2.5 m thick, with abundant plant debris on the bedding planes and with ostracods *Gerdalia* sp., conchostracans, aquatic beetles and a few remains of terrestrial insects (under study at present); (4) red clay containing thin interbeds of bluish siltstones, with crack-fill wedges penetrating the underlying deposits; (5) cross-bedded polymictic sand beds with gravel, more than 2.5 m thick, starting the second sedimentary cycle, with a few vertebra of amphibians *Tupilakosaurus* sp. and the *Procolophonidae* gen. sp. indet. (defined by M. A. Shishkin). These vertebrate fossils first appear in the latest Tatarian and are widespread in the Early Triassic.

MAGNETOSTRATIGRAPHY

Magnetostratigraphic studies of the Nedubrovo Section have shown a high magnetic susceptibility (mean χ 159.5 $\cdot 10^{-5}$) that is more typical of the Vetlugian deposits than of the Tatarian (Burov *et al.*, 1998). However, the polarity is reversed, while all the hitherto-studied lower Vetlugian sections fall in the direct polarity zone NPT (Lozovsky & Esaulova, 1998). We therefore designate the basal Vetlugian of Nedubrovo as a new reversed polarity zone R_0T, supposedly correlatable with the upper basalts of the Tchernyshov Ridge in the Timano-Petchorsk region.

A reversed polarity zone probably corresponding to R_0T at Nedubrovo was also found at the base of the Nyamunsk Formation in Lithuania, the stratigraphic equivalent of the lowermost Vetlugian, as well as of the basal Buntsandstein of Poland (Kisnerius & Saidakowsky, 1972; Katinas, 1997).

FOSSIL FLORA

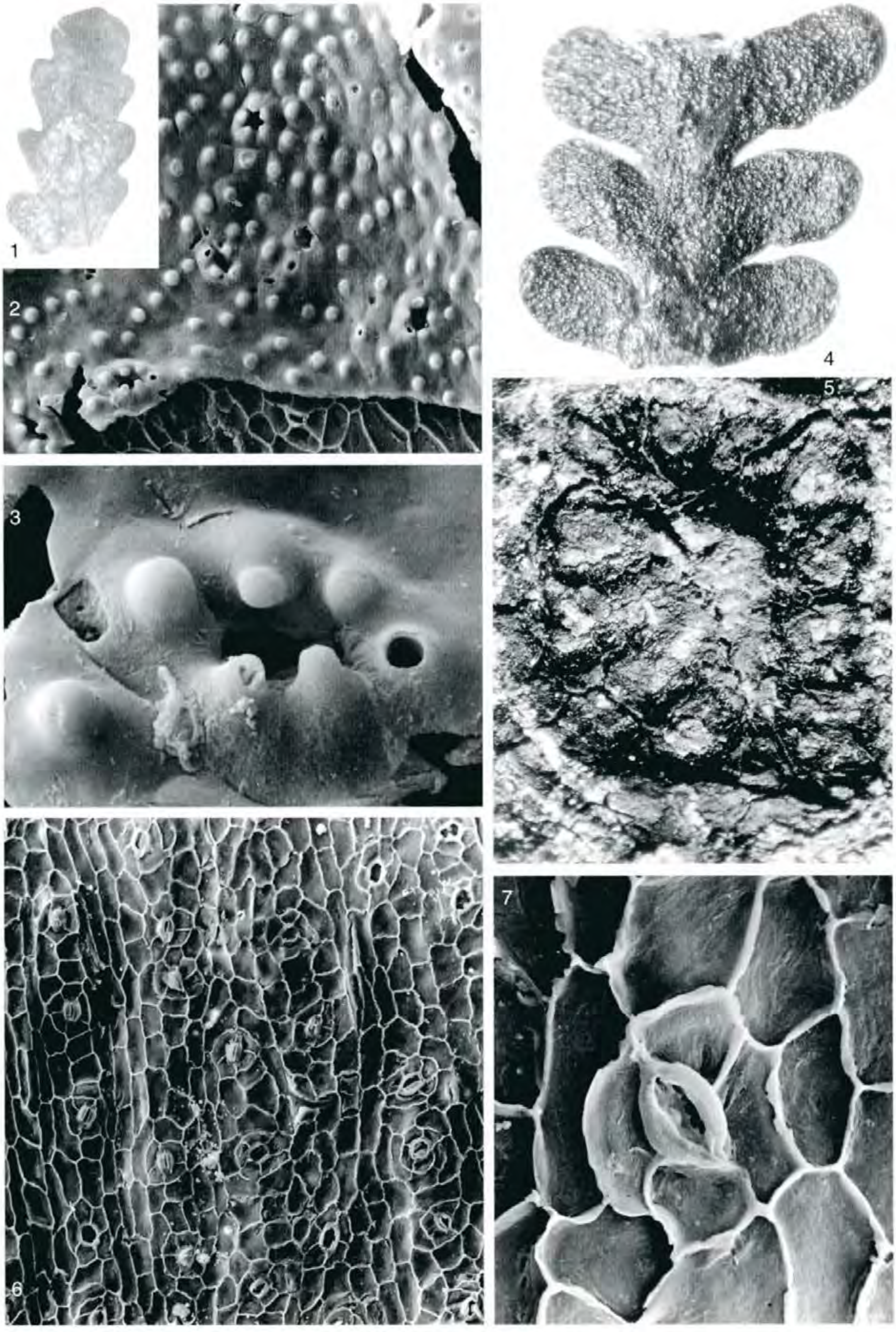
The plant mega- and mesofossils occur abundantly in the grey laminated siltstones and clays (bed 3). These accumulations of plant debris on bedding planes locally appear as a thin, lenticular coal. The plant remains are fragmentary but with well-preserved cuticles providing epidermal characteristics that are crucial for classification of the Permian and Triassic gymnosperms. A classification by S. Meyen (in Gomankov & Meyen, 1986; Meyen, 1992) is followed here for the sake of comparison with the Tatarian flora, although some generic assignments are in need of revision.

The plant megafossils constitute an essentially peltasperm–conifer assemblage with a few fern remains. The assemblage is dominated by peltasperms *Tatarina conspicua* S. Meyen, *T. lobata* S. Meyen, *Phylladoderma (Aequistomia) annulata* Meyen, *Raphidopteris antiqua* S. Meyen, *Peltaspermopsis buevichiae* (Gomankov et S. Meyen) Gomankov, and *Salpingocarpus variabilis* S. Meyen (Plate I). These species, with the single exception of *Tatarina lobata*, are known from the uppermost Tatarian (Vyatkian) localities (we accept the Vyatkian age for a controversial Aristovo locality with *Phylladoderma annulata* and *Peltaspermopsis buevichiae*). The cuticles of a

Fig. 1 – Sketch map of the Volga-Severnaya Dvina watershed region showing the geographical position of the Nedubrovo Section (black circle).

Plate I – Peltasperms from the fossil plant bed of Nedubrovo.

1. *Tatarina lobata* S. Meyen, leaf fragment, x15. 2, 3. *Tatarina (Tatarinopsis)* cuticle, x230, and stoma with hollow papillae, 1200. 4. *Rhaphidopteris antiqua* S. Meyen, pinna with decurrent pinnules, x10. 5. *Peltaspermopsis buevichiae* (Gomankov et S. Meyen) Gomankov, pelta with ovules, x20. 6, 7. *Tatarina conspicua* S. Meyen, cuticle, x110, and stoma, x700.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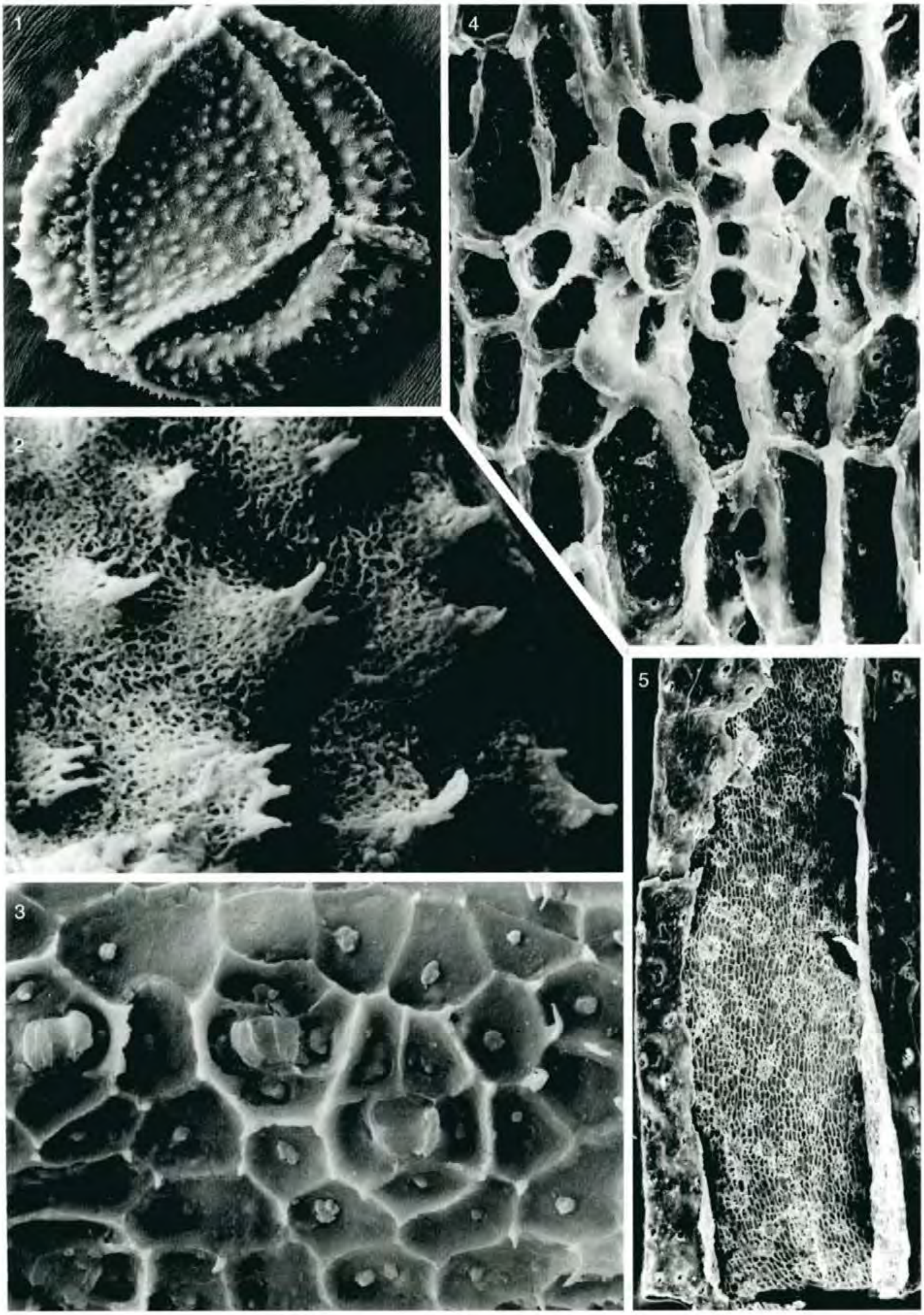
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dominant Tatarian species *T. conspicua* are also fairly common in the Nedubrovo locality. *T. lobata* was originally described from the Korvunchan Formation of the Tungusska Basin (Meyen & Gomankov, 1980). In Nedubrovo, the peltasperm leaves and reproductive organs are somewhat smaller than in typical Tatarian material and the cuticles often show anomalous cell patterns.

The conifers are represented by scattered leaves, the taxonomic assignments of which are based solely on epidermal characteristics (Plate II). Alongside a typically Tatarian *Quadrocladus dvinensis* S. Meyen there are *Ullmannia.* cf. *bronnii* Goepert and *Quadrocladus* cf. *solmsii* (Gothan et Nagalhard) Schweitzer, both comparable with the Zechstein conifers.

Thus the Nedubrovo megafossil flora is still essentially Permian, with a number of species surviving from the Tatarian. However, a few Zechstein and Korvunchan forms indicate a younger age than the uppermost Tatarian. It bears a general similarity to the late Changhsingian flora of Tieqiao Section, Laibin County, south China, dominated by the Permian peltasperms, gigantopterids and conifers, with conifer assemblages of Zechstein aspect (Jin *et al.*, 1998 and our unpublished data).

Prominent in the plant mesofossil assemblage of bed (3) is *Otynisporites eotriassicus* Fugl. (Plate II), the index species of a megaspore zone comprising the basal Suboolitic Member of Buntsandstein immediately above the Zechstein (Fuglewicz, 1977).

PALYNOLOGY

The spore-pollen assemblages were obtained from the beds (2-4), with insignificant variation from bed to bed (Plate III). They are dominated by *Klausipollenites schaubergeri* (Potonié et Klaus) Jansonius and *Cycadopites* sp., summarily including more than 50% of the palynomorphs. Non-taeniate pollen is also represented by the subordinate *Klausipollenites decipiens* Jansonius, *Alisporites nuthallensis* (Clarke) Balme, *A. grauvogelii* Klaus, *Falcisporites zapfei* Potonié et Klaus, and *Platysaccus queenslandii* de Jersey. The taeniate pollen grains are assigned to *Protohaploxypinus* cf. *pantii* (Jansonius) Orlowska-Zwolinska, *Lueckisporites virkkiae* Potonié et Klaus, *Lunatisporites noviaulensis* (Leschik) Foster, and *L. transversmundatus* (Jansonius) Fisher, ranging from 0.5% to 2% each, *Striatoabieites richteri* (Klaus) Hart, up to 3%, and *L. pellucidus* (Goubin) Helby, locally up to 12-15%. Occasional grains belong to *Ephedripites permasensis*, *E.* sp., *Striomonosaccites* sp. and *Triadispora* cf. *crassa* Klaus.

Spores are less diverse, with a few numerically prominent forms, such as *Apiculatisporis*, up to 30%, and *Limatulasporites fossulatus* (Balme) Foster, up to 15%. *Punctatisporites triassicus* Schulz, *Polycingulatisporites densatus* (de Jersey) Playford et Dettmann, *Leptolepidites jonkeri* (Jansonius) Yarosh. et Golubeva, *Proprisporites pocockii* Jansonius, *Densoisporites playfordii* (Balme) Dettmann and *Pechorosporites disertus* Yarosh. et Golubeva amount to 1-2% each.

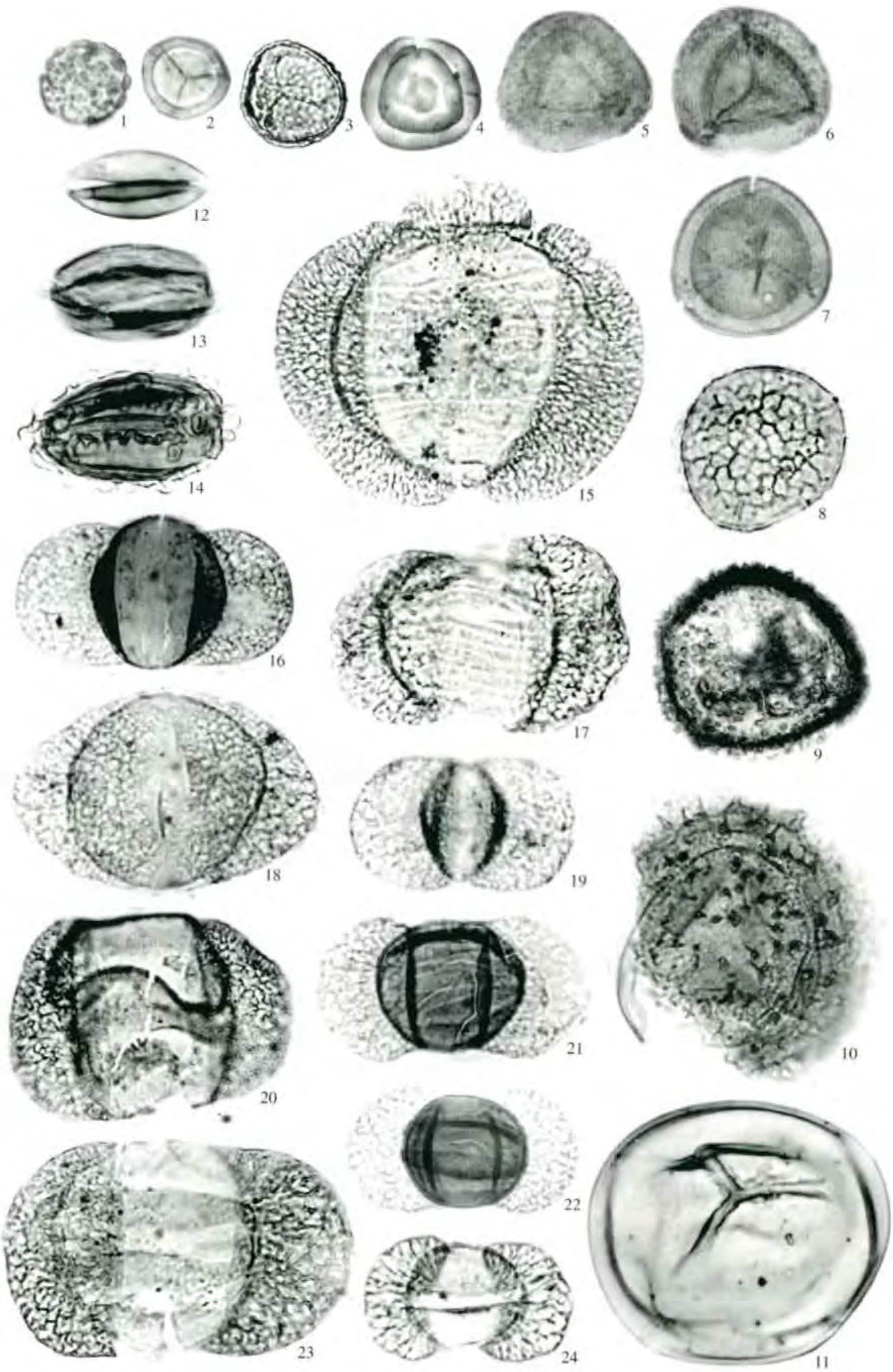
Common in the assemblage are *Tympanicysta stoschiana* Balme as well as the planktonic prasinophytes *Pterospermella*, *Pilasporites* and *Inaperturopollenites nebulosus* Balme.

The association of the prevailingly Permian *K. schaubergeri* (with a few reliably-dated Early Triassic records in the early Griensbachian of Arctic Canada and the Induan Panchet Formation with *Lystrosaurus:* Fisher, 1979; Tiwari & Tripathi, 1992), *Lueckisporites virkkiae*, *Falcisporites zapfei* and *Alisporites nuthallensis* with the Early Triassic *Proprisporites pocockii*, *Leptolepidites jonkeri*, *Polycingulatisporites densatus*, *Densoisporites playfordii*, *Pechorosporites disertus*, *Lunatisporites pellucidus*, *L. transversmundatus*, *Ephedripites permasensis* and abundant *Cycadopites*, indicate a transitional uppermost Permian to lowermost Triassic age for the Nedubrovo palynological assemblage. It is closely comparable to palynofloras from the lowermost Buntsandstein of Poland (Orlowska-Zwolinska, 1984), *Otoceras* beds of western Canada (Jansonius, 1962), Arctic Canada (Fisher, 1979; Utting, 1994) and the *Protohaploxypinus* zone of eastern Greenland (Balme, 1979).

CORRELATION

A correlation of the major continental sequences is shown in Fig. 2. In a relatively complete Permian-Triassic sequence of the Junggar Basin, northern China, the fossiliferous transitional deposits are exposed in two limbs of the Dalongkou Anticline (Yang *et al.*, 1986; Cheng *et al.*, 1989). A graphical correlation made by the senior author has shown that the megaspore zone *Otynisporites eotriassicus* of the Upper Guodikeng Formation (Liu, 1994) comprises the interval of joint occurrences of *Dicynodon* and *Lystrosaurus* and extends upsection with *Lystrosaurus* alone. Thus the FAD of *Lystrosaurus* coincides with that of *Otynisporites eotriassicus*. Paleomagnetic zonation is not yet completed for Junggar Basin. However, the Upper

Plate II – Megaspore and cuticles from the fossil plant bed of Nedubrovo: 1, 2. *Otynisporites eotriassicus* Fugl., proximal aspect, x280, and distal appendages, x2000, 3. *Quadrocladus dvinensis* S. Meyen, group of stomata, x 480, 4, 5. *Ullmannia* cf. *bronnii* Goepp., stoma with a ring of subsidiary cells anomalously intruded by an ordinary cell, x800, and whole leaf cuticle showing the arrangement of stomata, x53.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Guodikeng, as well as the overlying basal Jiucaiyuan Formation, show a reversed polarity (Cheng *et al.*, 1989).

Of a certain importance for the non-marine to marine PTB correlations is the occurrence of *Otynisporites eotriassicus* in the marginal marine Tesero Oolite near the base of the Werfen Formation, Southern Alps, at about the PTB position as defined by Broglio Loriga & Cassinis (1992). According to Kozur (1989, 1998), the megaspores were found in the Tesero section about 1.8-2.2 m above the boundary with the underlying Bellerophon Formation. They associate with a palynological *Lundbladispora obsoleta-Lunatisporites noviaulensis* assemblage similar to that of the lower Buntsandstein, with a mass occurrence of *Tympanicysta stoschiana*, as well as with conodonts *Hindeodus praeparvus* Kozur and *Isarcicella? prisca* Kozur. These species indicate the conodont zone *Clarkina (Neogondolella) meishanensis-Hindeodus praeparvus* (Kozur, 1998), the base of which correlates with the first appearance of *Otoceras* (zone *O. concavum/latilobatum*). This level corresponds to the PTB as defined by Orchard & Tozer (1997) and Orchard & Krystyn (1998). In the Meishan Section of south China, the *Clarkina meishanensis* zone (Mei, 1996) falls in the interval of reversed polarity (Zhu & Liu, 1999). It should be noted that a previous report of direct polarity in the lower part of the *O. concavum* zone of Arctic Canada was not confirmed by the recent studies (Ogg & Steiner, 1991).

This level is also marked by the appearance of *Lystrosaurus* in continental facies (Lozovsky & Esaulova, 1998) and, in terms of event stratigraphy, by the onset of a widespread transgression, trap basalt eruptions, a peak of *Tympanicysta* and the prominent isotopic excursions (Fig.

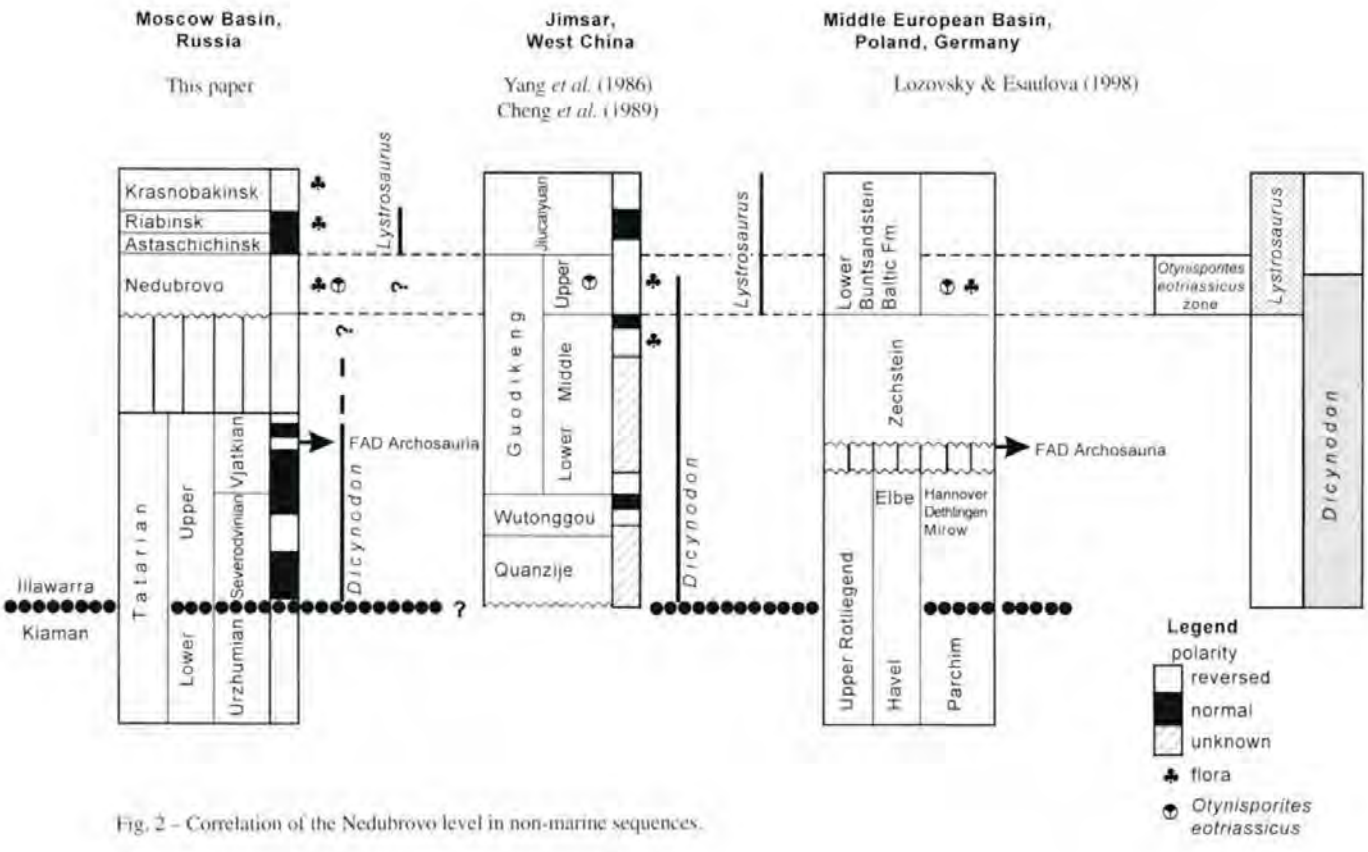

Fig. 2 - Correlation of the Nedubrovo level in non-marine sequences.

Plate III - Palynological assemblage of Nedubrovo, x625.

1. *Leptolepidites jonkeri* (Jansonius) Yarosh. et Golubeva, 2. *Limatulasporites fossulatus* (Balme) Helby et Foster, 3. *Apiculatisporis* sp., 4. *Polycingulatisporites densatus* (de Jersey) Playford et Dettmann, 5. *Densoisporites* sp., 6. *Densoisporites* sp., 7. *Densoisporites playfordii* (Balme) Dettmann, 8. *Proprisporites pocockii* Jansonius, 9. *Pechorosporites* sp., 10. *Pechorosporites disertus* Yarosh. et Golubeva, 11. *Punctatisporites triassicus* Schulz, 12. *Cycadopites* sp., 13. *Ephedripites* sp., 14. *Ephedripites permasensis* Yarosh., 15. *Striomonosaccites* sp., 16. *Falcisporites zaptei* Potonié et Klaus, 17. *Striatoabieites richteri* (Klaus) Hart., 18. *Klausipollenites schaubergeri* (Potonié et Klaus) Jansonius, 19. *Platysaccus* sp., 20. *Scutasporites* sp., 21. *Protohaploxypinus* sp., 22. *Lunatisporites noviaulensis* (Leschik) Foster, 23. *Lunatisporites pellucidus* (Goubin) Helby, 24. *Lueckisporites virkkiae* Potonié et Klaus.

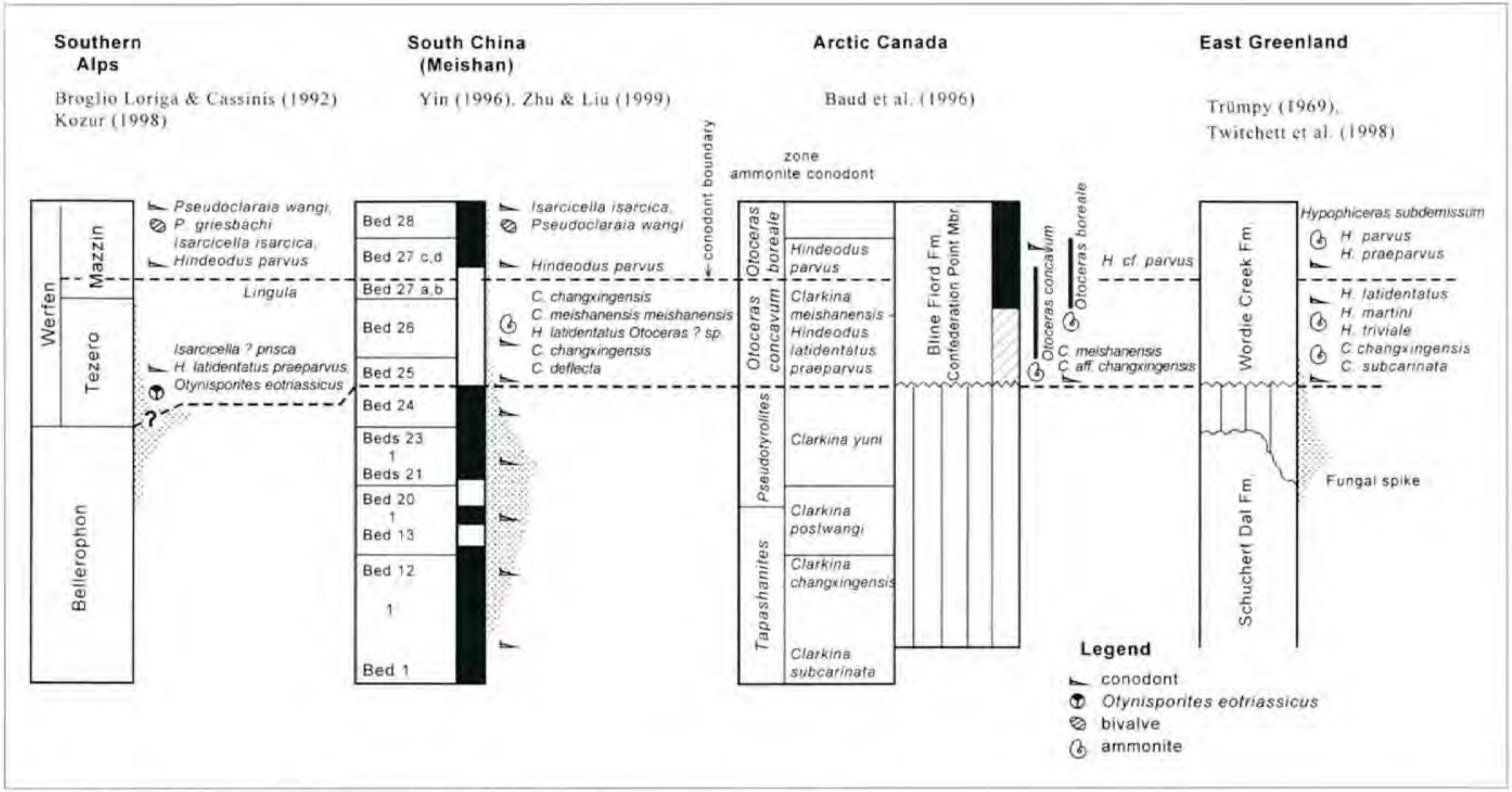

Fig. 3 – Correlation of the Nedubrovo (*Otynisporites eotriassicus*) level in marine sequences.

3). However, in the widely accepted conodont zonation, the PTB is drawn above this level, at the base of the next conodont zone *Hindeodus parvus*. Whatever the final decision on the PT GSSP, it has to be taken into consideration that at the level of the earliest *Otoceras*, *Lystrosaurus* and *Otynisporites* records, both marine invertebrate assemblages and terrestrial flora still retained the Late Permian aspect, with subordinate Triassic newcomers.

CONCLUSIONS

The Nedubrovo Section on the Kichmenga River, Vologda region, represents a relatively complete transboundary PT sequence, with the upper Tatarian overlain by the basal Vetlugian which contains a plant megafossil assemblage of Permian aspect, with most species having survived from the Late Tatarian, megaspores of *Otynisporites* zone (basal Buntsandstein of Poland), and a rich palynological assemblage of a mixed Zechstein-Lower Griesbachian character. The relatively diverse planktonic Prasinophyceae probably indicate a marine influence at a high stand of the end-Permian boreal transgression. A reversed polarity zone is established for these deposits. The Nedubrovo sequence thus appears older than the basal Vetlugian elsewhere in European Russia. It conceivably represents a stratigraphic interval missing in the less complete transboundary sections.

On the basis of the evidence, the Nedubrovo sequence is correlated with the upper Guodikeng Formation of the Junggar Basin in China, both showing a reversed polarity. It is stratigraphically equivalent to or somewhat older than the lowermost Buntsandstein of Western Europe. Probable marine correlates of Nedubrovo are the lowermost part of the *Otoceras* zone as well as the Tesero Oolite and the Transitional beds 1 and 2 below the *Hindeodus parvus* FAD in the Meishan Section of south China. This stratigraphic level is traceable by the joint occurrences of *Otynisporites*, the earliest *Lystrosauridae* and, in marginal marine deposits, conodonts of the *Clarkina (Neogondolella) meishanensis-Hindeodus praeparvus* zone. It is also marked by the onset of a widespread transgression, trap eruptions in Siberia and prominent isotopic anomalies. There may have been a certain time lag between these events and biotic change, since the biota was still of a prevailingly Permian character.

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THE CONTINENTAL PERMIAN OF NORTHEAST EUROPE

ELENA O. MALYSHEVA[1]

Key words – Facies; depositional environments; paleogeography; correlation; sequence stratigraphy; Pechora Basin.

Abstract – A sequence stratigraphy approach applied to the Upper Permian of northeastern Europe appeared to be very helpful in subdivision and correlation of complex marine to non-marine successions and identification of the intervals favourable for paleogeographical reconstructions. The biostratigraphically proven Kungurian-Ufimian stage boundary is interpreted as a sequence boundary marked by basinward facies shift and a change in parasequence stacking patterns. Each following sequence is characterised by further progradation of low-stand deltaic-nearshore facies to the north and northwest. By the end of Ufimian times, continental deposition dominated. Alluvial and lacustrine depositional environments, paleosols and calcretes have been interpreted within the continental complexes. Mature calcrete horizons have been used to mark sequence boundaries within the continental part of the section. The confidence in correlations and sequence identification decreases up the section.

Parole chiave – Facies; ambienti deposizionali; paleogeografia; correlazione; stratigrafia sequenziale; Bacino della Pechora.

Riassunto – Un approccio della stratigrafia sequenziale nell'ambito del Permiano superiore dell'Europa nord-orientale si dimostra assai utile nel suddividere e correlare complesse successioni marine e non-marine, nonché nell'identificare intervalli favorevoli per le ricostruzioni paleogeografiche. Il biostratigraficamente dimostrato limite Kunguriano-Ufimiano è interpretato come un limite di sequenza demarcato da uno spostamento di facies verso il bacino e da un cambio nei modelli di parasequenze accumulatesi. Ciascuna successiva sequenza è caratterizzata da un'ulteriore progradazione verso nord e nord-ovest di facies deltizie prossime alla spiaggia, in situazione di *lowstand*. Entro la fine dell'Ufimiano dominò una deposizione continentale. Ambienti deposizionali alluviali e lacustri, paleosuoli e calcrete sono stati interpretati all'interno di complessi continentali. Orizzonti di calcrete mature sono stati usati per tracciare limiti di sequenza entro la parte continentale della sezione. La sicurezza ad operare delle correlazioni e ad identificare delle sequenze decrescono verso l'alto.

INTRODUCTION

Continental deposits of Late Permian age are widespread in the Pechora sedimentary basin, located in northeastern Europe between the Urals in the east and the Timan Ridge in the west (Fig.1). Late Permian sedimentation was associated with the Urals orogeny and occurred in various depositional environments (from marine to non-marine), controlled by two major structural (Pechora Synclise and Pre-Ural Foredeep) and two climatic (northeastern humid and southwestern semi-arid) zones. F.I. Entzova, G.A. Ivanov, V.I. Chalyshev, L.L. Khaytzer, J.N.Lubina, A.V. Makedonov, I.S.Muravyev, N.I. Nikonov, N.S.Oknova, A.P.Rotay, G.M.Yaroslavtzev and others contributed to the facies study of the region. However paleogeographical reconstructions for epochs seem to be very uncertain because of complicated facies combinations caused by steady regression of the sea basin interrupted by short transgressions. Thus, the objectives of the current study were to produce paleogeographical reconstructions of the Pechora Basin for particular time "slices" critical to the evolution of sedimentation in the Late Permian.

The study was based on core and log data from more than 100 wells and eight outcrop sections, and incorporated paleontological evidence (Grunt *et al.* (eds), 1998; Entzova *et al.*, 1969; Kanev & Koloda, 1997; Koloda & Kanev, 1994; Koloda *et al.*, 1992; Konovalova, 1991; Chuvashov, 1997; Chuvashov *et al.*, 1990) and analytical data on the chemistry and mineralogy of the rocks.

CONTINENTAL FACIES

The recognition of facies within the most representative sections of the Upper Permian both in cores and outcrops revealed a variety of depositional environments including

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offshore shelf, lower and upper shoreface, tidal flat, coastal and alluvial plains, and lacustrine (Malysheva, 1997). Coals, alluvial and lacustrine deposits, paleosols and calcretes have been recognised in the Permian continental succession. Many publications have been devoted to the problems of coal-bearing formations in the Pechora Basin (Makedonov, 1965; Dedeev, 1990). Coals are referred to as paralic and limnic types that correspond to coastal and alluvial plain swamps.

Alluvial channel bodies are recognised by significant lateral heterogeneity, lenticular forms, thinning upward sections, sharp to erosional contacts with subjacent layers, planar to trough cross-bedding and some other criteria including the shape of gamma and electric logs. Channel and flood plain deposits have been distinguished within alluvial units. Flood plain accumulation dominated and was commonly accompanied by pedogenic processes. The number and thickness of channels increases up the Permian section and eastward. The thickness of the channel fill within the synclise does not exceed 20 m, while within the Pre-Ural foredeep it reaches 100 m. Moreover, coarse sandy-conglomeratic beds are most frequent within the foredeep, while medium sandstones with thin conglomeratic interlayers are typical of the synclise. Morphology and composition of the channel beds also differ between the northern humid and southern semi-arid zones.

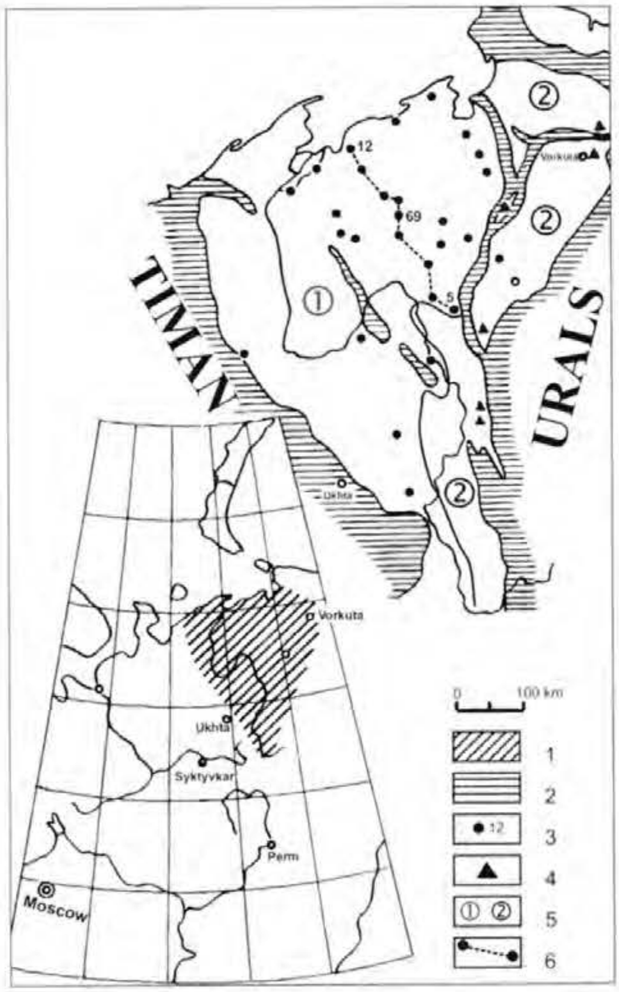


Fig. 1 – Location map.

1. considered area; 2. absence of the Upper Permian; 3. location of the boreholes; 4. location of the outcrops; 5. major structural zones: *1.* Pechora Synclise, 2. pre-Ural foredeep; 6. location of facies cross-section.

Lacustrine deposits are very common but they are regarded as proved only in association with subtidal grey shales with horizontal to lenticular bedding and clayey carbonates with non-marine bivalves and ostracods. Representatives of such genera as *Palaeumutella, Antraconauta, Synjaella,* and *Abiella* preferred a muddy substratum in the subtidal zone of the lacustrine parts of the basins (Koloda & Kanev, 1994).

Fossil soils were primarily proved and comprehensively studied in one of the sections of the Pre-Ural foredeep (Chalyshev, 1974). The major features of the identified mature soil profiles could be summarised as destratification, rootlets and horizonation (formation of soil horizons) with obvious alternations in colour, mechanical and chemical compositions, and content of organic matter from parent rocks upwards in the soil profile. Our studies of the area of the Pechora Synclise revealed abundant intervals in flood plain sections overprinted by pedogenic processes and few paleosols with more or less expressed horizonation. In general, they are represented by mottled, green or red clayey destratified rocks with peds, root relics or dessication fractures, and to a varying degree differ from the parent rocks in clay mineralogy and chemical composition. Immature fossil soils dominate.

Calcretes, in most cases interpreted as calcic paleosols, are widespread in the southern semi-arid zone of the Pechora Basin and mark periods of aridisation, even in the northern areas. Calcretes have been recognised using the criteria suggested by V.P. Wright (1989), M. Esteban & C.F. Klappa (1991) and others. Scattered nodules, globular to massive and brecciated calcretes have been distinguished. Typical microstructural features include rounded peloids and grains, circumgranular fractures and crystallaria, sitting in a micritic groundmass. The most mature profiles contain those calcretes, and are characterised by a gradual transition from weakly calcareous forms with rare carbonate nodules to impermeable massive or brecciated carbonate horizons. The number and thickness of individual calcrete horizons increase up the profile.

EVOLUTION OF SEDIMENTATION

A sequence stratigraphy approach (Weimer & Posamentier, 1994; Van Wagoner *et al.*, 1990) was applied to the

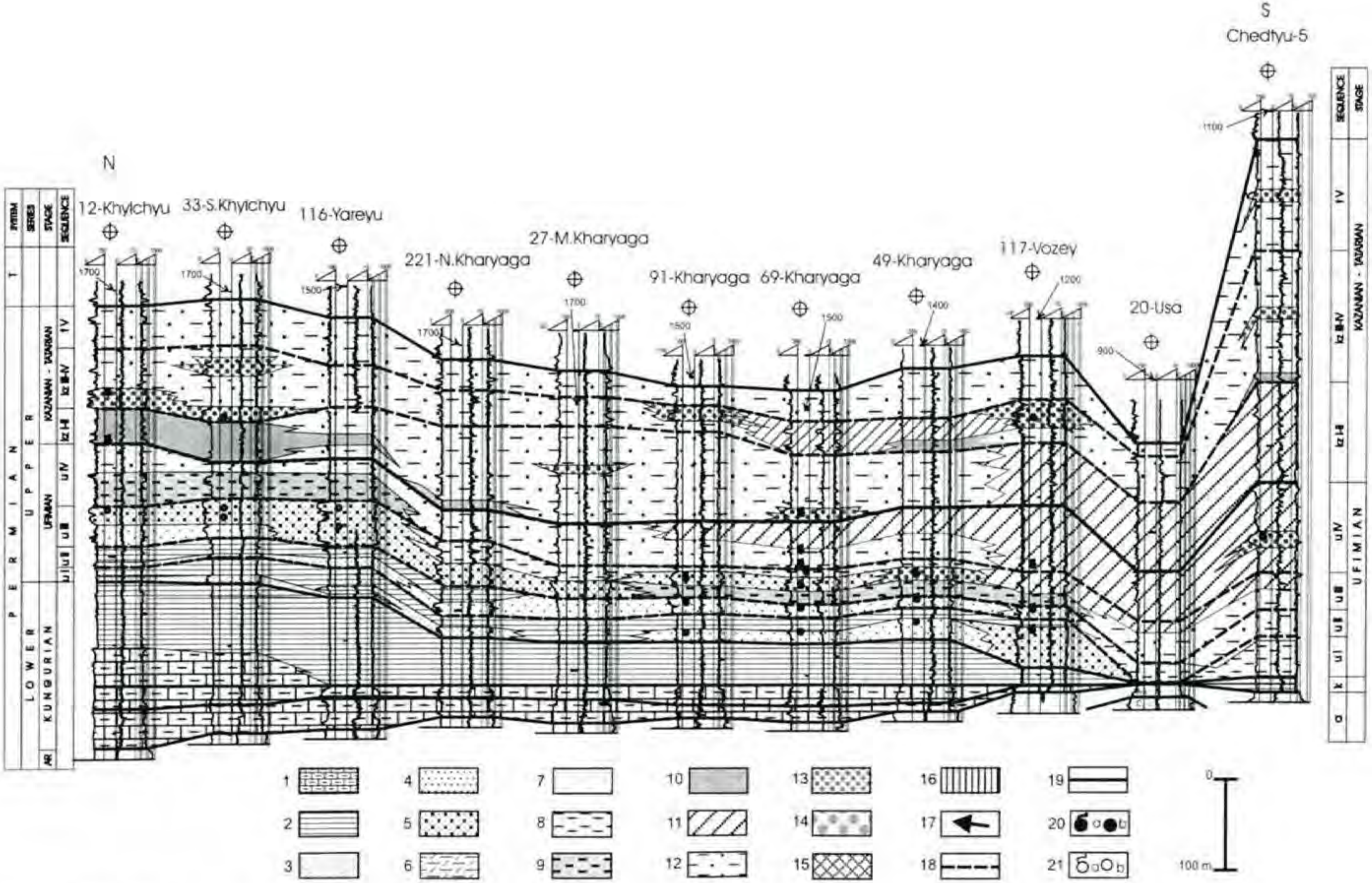


Fig. 2 - Facies cross-section.

1. offshore shelf with mixed carbonate and clastic deposits; 2. offshore shelf dominated by clastics; 3. distal bars (finegrained sandstones); 4. lower shoreface; 5. upper shoreface; 6. partly isolatated shelf or lagoon with carbonates; 7. tidal flat (intertidal clastics and carbonates predominate); 8. coastal plain with red beds; 9. coastal plain with peat accumulation; 10. alluvial-lacustrine plain with peat accumulation; 11. continental environments (in general) dominated by red beds; 12. continental environments (in general), dominated by grey beds; 13. alluvial and distributary channels; 14. alluvial fans (conglomerates and gravelites); 15. absence of Upper Permian deposits; 16. absence of Lower Ufimian deposits; 17. influx of coarse clastics; 18. sequence boundaries; 19. stage boundaries; 20. oil productive reservoir: *a.* in the well, *b.* in the field; 21. gas and condensate productive reservoir.

correlation of the sections and identification of the intervals favourable for paleogeographical reconstruction. So-called reference sections, both lithologically and paleontologically representative, were used as the basis for sedimentological interpretation and identification of possible sequence boundaries (Malysheva *et al.*, in press). The latter was based on such criteria as evidence of erosional truncation, basinward shifts in facies and changes in parasequence stacking patterns. Vertical and lateral associations of depositional environments within the available biostratigraphic framework provided identification of sequences in the Upper Permian succession. One of the cross-sections displayed in Figure 2 shows facies relationships and possible sequence boundaries, sometimes coinciding with stage boundaries that were traced along the Kolva swell. Lowstand and transgressive systems tracts are considered to be the most favourable for paleogeographical maps.

The boundary between the Kungurian and the Ufimian in some sections, the Kozhim River being the best (Grunt *et al.* (eds), 1998), is defined by several groups of fossils: bivalves, brachiopods, bryozoans, as well as microflora. The latter are of greatest importance in the age definitions of the Upper Permian. In the northeast of the Pechora Syncline this boundary is marked by a basinward shift in facies, producing a change in the stacking pattern of parasequences and is regarded as the sequence boundary. Progradation of a bar-deltaic complex forms a well-documented lowstand systems tract, dominated by lower and upper shoreface sandstones (Fig. 3-A). An intertidal flat with oolitic shoals, coquinas and small tidal channels and coastal plain with peat accumulation, channels, first non-marine bivalves and interbeds with lingulas and marine bivalves developed in the Pre-Ural foredeep landward of bar-deltaic zone. An offshore marine shelf draining basinwards is characterised by dark bioturbated shales, siltstones and interbedded, very finegrained sandstones, sometimes with abundant marine fossils.

In the southwestern areas of the Pechora Basin the Kungurian-Ufimian boundary is also marked by a basinward shift in facies, but the lowstand systems tract of the first Ufimian sequence is dominated here by sediments of a semi-restricted shelf or lagoon (with mixed clastic and carbonate accumulation, poor in fossils), intertidal flats and coastal plains with red beds. Alluvial fans, composed of conglomerates and gravelites, have been reported by Muravyev (1972) from the south of the pre-Ural foredeep.

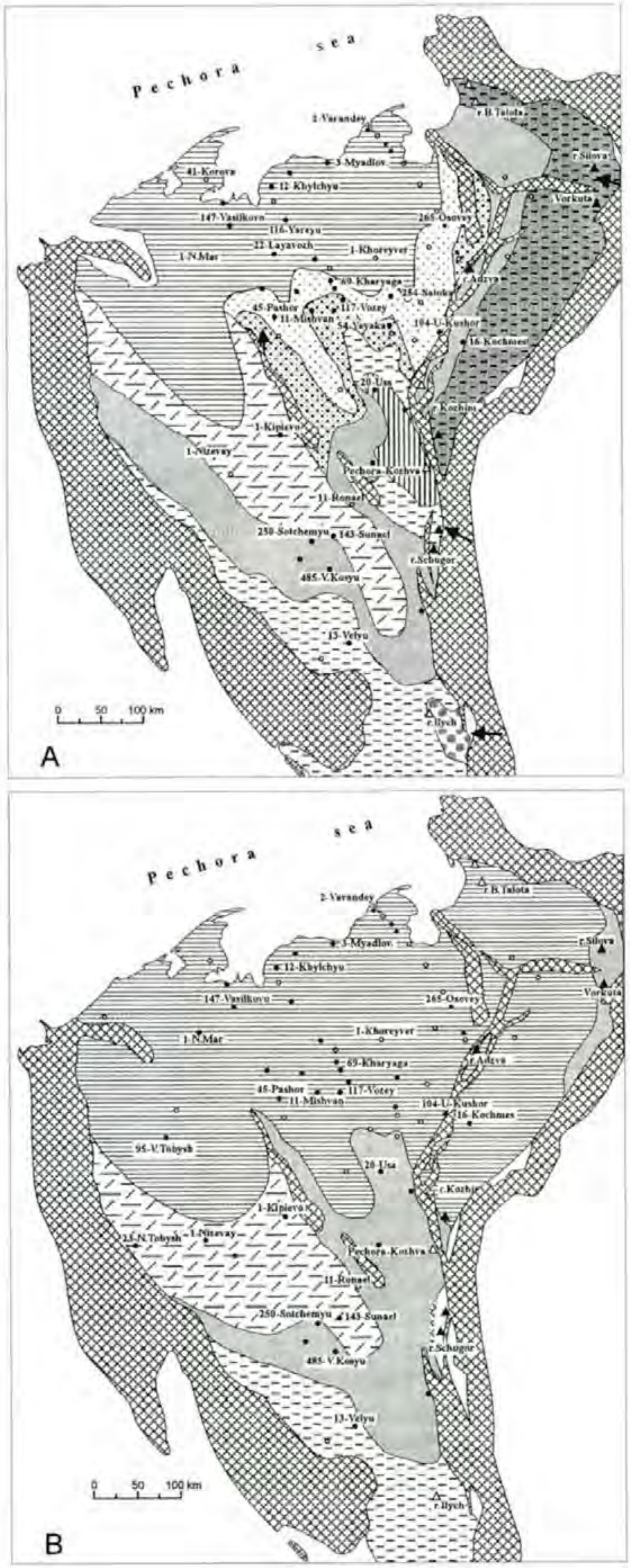
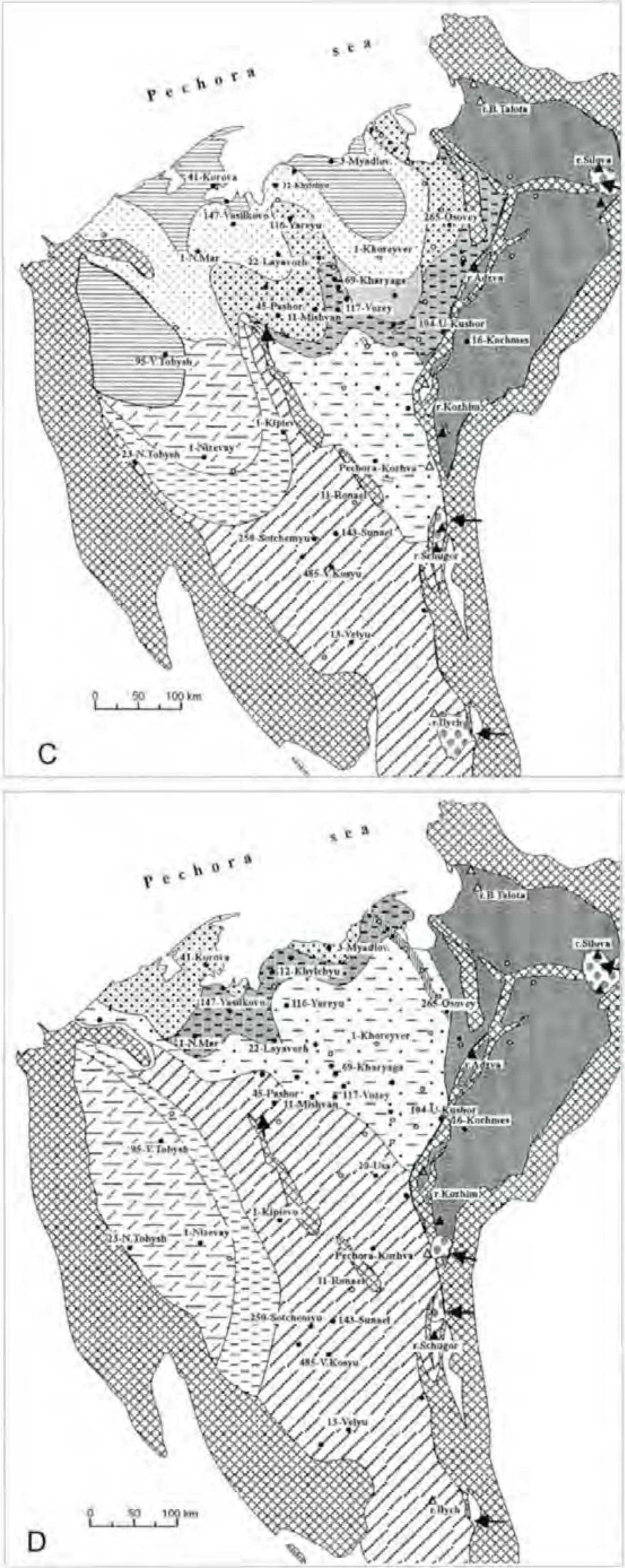


Fig. 3 – Paleogeographical maps:

A. lowstand system tract of the I Ufimian sequence; B. transgressive system tract of the I Ufimian sequence; C. lowstand system tract of the III Ufimian sequence; D. lowstand to transgressive system tract of the I Kazanian sequence. (See Fig. 2 for legend).



Facies and thickness distributions as well as the petrography of sandstones and conglomerates, strongly suggest, that the paleo-Urals mountains were the major source of polymictic clastics. However it is assumed that in the Upper Permian the area of the Timan Ridge represented the lowland plain which could also serve as a transition zone for the Uralian material. A series of islands and peninsulas might have existed on the area of the Pechora-Kozhva swell, restricting and providing specific sedimentation to the southwestern part of the Pechora Basin.

The maximum flooding surface of the first Ufimian sequence is well expressed over the entire area of the Pechora Basin. Maximum transgression occurred in the north. The map of transgressive systems tract is dominated by offshore marine facies in the northwest and semi-restricted facies in the southeast (Fig. 3-B).

Each following Ufimian sequence marks a basinward facies shift, with progradation of lowstand deltaic-nearshore facies to the north and northwest. The third Ufimian sequence is characterised by considerable sand influx which resulted in the development of a widespread bar-deltaic complex over the area of the Pechora Synclise (Fig. 3-C). A coastal plain (lowstand) with peat accumulation shifted westwards from the pre-Ural foredeep, and a deltaic complex with distributaries and mouth bars (highstand) is documented on the Kharyaga field. By the end of Ufimian times, continental deposition dominated over most of the considered area and was strongly controlled by climatic variations: coals became very common in humid climates (northern areas), while red beds with calcretes developed in the semi-arid areas (south). Very often the latter beds mark sequence boundaries. The most mature and thick calcrete horizons are confined to the tops of the Ufimian and the Lower (?) Kazanian.

The boundary between Ufimian and Kazanian stages is defined by spores and pollen, floral complexes, bivalves and ostracods. However, it appeared to be less well grounded than the Kungurian-Ufimian boundary. Non-marine facies dominate (Fig. 3-D). Marine (shoreface) environments and coastal plains with peat accumulation and some interbeds with marine fossils were preserved only in the northwestern part of the area. In the west, semi-isolated environments rich in marls accumulated. Correlation of sequences and identification of system tracts become more complicated in the higher part of the section because of continental facies with alluvial channels. Only some sequence boundaries, marked by calcrete horizons in the southern and some northern areas, could be more or less correlated throughout the basin. Transgressive and highstand systems tracts become most expressed. Transgressive deposits in continental environments are marked by the development (even within the red beds) of grey lacustrine facies with abundant non-marine bivalves and ostracods.

The boundary between the Kazanian and the Tatarian is the least proved over the most part of the considered area. Only in the outcrops and well sections with sufficient coring can they be distinguished by microflora, bivalves and ostracods. The composition of the deposits is similar to the Kazanian.

CONCLUSIONS

During the Upper Permian, continental environments completely replaced marine deposition in northeastern Europe. Vertical and lateral variations of facies patterns caused by steady regression of the sea basin, interrupted by short transgressions, suggest that facies mapping can be most valuable for relatively short time intervals. The constructed maps, illustrating the evolution of sedimentation, should be regarded as the first approach. Cyclic regressive development of the Pechora Basin in the Upper Permian together with climatic variations provide a good opportunity for correlation of non-marine and marine deposits and can be applied to chronostratigraphical analysis using sequence stratigraphy as a tool.

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2. NON-EUROPEAN TERRITORIES AND GLOBAL MATTERS

INVERTEBRATE FAUNAS AND PRELIMINARY PALYNOLOGY, CARBONIFEROUS-PERMIAN BOUNDARY STRATOTYPE, AIDARALASH CREEK, KAZAKHSTAN

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Key words – Ammonoids; Carboniferous; Conodonts; Fusulinaceans; Palynology; Permian; Stratotype; Ural Mountains.

Abstract – The global stratotype section and point (GSSP) for the base of the Permian System and therefore the Permian-Carboniferous boundary and the base of the Asselian Stage has been formally established in the southern Ural Mountains along Aidaralash Creek, Kazakhstan. The definition is based on the First Appearance Datum (FAD) of the conodont *Streptognathodus isolatus*. A previously suggested ammonoid boundary occurs 26.8 meters above, at the base of bed 20. A "traditional" fusulinacean boundary occurs 6.3 m above the conodont boundary. The ammonoid, fusulinacean, and conodont faunas from the GSSP are well studied and provide a basis for correlation with other marine sections within the Urals and internationally. However, the Permian System includes extensive non-marine strata that are not easily correlated with their marine counterparts. Correlation of non-marine strata to marine stratotype requires study of fossil biota that bridge the two. Recent examination of the palynology of the Carboniferous-Permian boundary strata (24.2 m below to 26 m above) from Aidaralash confirms the occurrence of a miospore assemblage similar to the *L. monstruosus-V. costabilis* Assemblage Zone of Utting in stratigraphic association with abundant conodonts, ammonoids and fusulinaceans.

Parole chiave – Ammoniti; Carbonifero; Conodonti; Fusulinidi; palinologia; Permiano; stratotipo; Monti Urali.

Riassunto – Il punto e la sezione dello stratotipo globale (GSSP) per la base del Permiano, e pertanto il limite Permiano-Carbonifero e la base dell'Asseliano sono stati formalmente stabiliti negli Urali meridionali lungo il torrente Aidaralash, in Kazakhstan. La definizione di questo limite è basata sulla prima comparsa (FAD) del conodonte *Steptognathodus isolatus*. Un limite ad ammoniti, precedentemente proposto, è presente 28 m al di sopra, e cioè alla base dello strato 20. Un limite "tradizionale" a fusulinidi figura 6.3 m sopra il limite a conodonti. Le faune a ammoniti, fusulinidi e conodonti del GSSP sono ben studiate e forniscono una base per correlazioni con altre sezioni marine entro gli Urali e internazionalmente. Tuttavia, il Permiano include estesi strati non-marini che non sono facilmente correlabili con i loro sostituti marini. La correlazione di strati non-marini con stratotipi marini richiede lo studio di faune che facciano da ponte tra i due. Recenti esami palinologici degli strati al limite tra il Carbonifero e il Permiano (da 24.2 m sotto a 26 m sopra) nella sezione di Aidaralash conferma l'evento di un'associazione a miospore simile alla Zona di Associazione *L. monstruosus - V. costabilis* di Utting, unita stratigraficamente ad abbondanti conodonti, ammoniti e fusulinidi.

INTRODUCTION

The global stratotype section and point (GSSP) for the base of the Permian has been officially defined (Davydov *et al.*, 1998) in strata at Aidaralash Creek, Aqtöbe (formerly Aktyubinsk) region, in the southern Ural Mountains of northern Kazakhstan. The base of the Permian System occurs at the first appearance of the conodont *Streptognathodus isolatus*. Establishment of the GSSP for the base of the Permian in the Ural Mountains represents the culmination of a long series of studies dating back to the original founding of the Permian System by Murchison (1841). Subsequent to its establishment, the lower boundary of the Permian has been repeatedly lowered in the Urals and various other regions, however, a GSSP for the base of the Permian had not been established until recently. Lacking a formal world standard, the lower boundary of many Permian successions outside of the Urals was drawn at locally convenient arbitrary levels without reference to the Ural sections; this practice was particularly flagrant in China and Australia.

Following the initial work by R.I. Murchison (1841), A.P. Karpinsky (1874), D.M. Rauser-Chernosova (1940), V.E. Ruzhencev (1936, 1945, 1950, 1951, 1952, 1954),

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I.V. Khvorova (1961) and S.E. Rosovskaya (1962), among others, conducted classic studies of Urals stratigraphy. More recently, M.F. Bogoslovskaya *et al.* (1995), V.I. Davydov *et al.* (1998), B.I. Chuvashov *et al.* (1993), V.V. Chernykh & S.M. Ritter (1994, 1997), V.V. Chernykh *et al.* (1997a, 1997b) among others, and the efforts of Permian and Carboniferous workers in preparation for the International Congress of the Permian in Perm, 1991 are particularly noteworthy. The Permian Congress provided the stimulus for renewed emphasis on Permian studies.

Many sections in the classic region of the southern Urals have been studied and in recent years two sections were contenders to serve as the GSSP – the sections at Aidaralash Creek and Usolka (Fig. 1). Strata of the Usolka section were deposited in the Ural sub-basin and the Aidaralash strata in the Aqtöbe sub-basin – both in the Pre-Uralian foredeep. Carboniferous-Permian strata of the Aqtöbe sub-basin were deposited on a marine shelf; those of the Ural sub-basin were deposited in deeper water (Snyder *et al.*, 1994). Strata of the deep marine facies of the Usolka section are exceptionally rich in conodonts, enabling precise correlation with Aidaralash and other sections in the Urals. On this basis, therefore, conodont workers favored Usolka as a potential GSSP. The section at Aidaralash was selected, however, because it meets many of the most important criteria for establishment of GSSPs. Especially significant is the diverse biota: abundant ammonoids, fusulinaceans, and conodonts offer great potential for worldwide correlation with other marine Permian strata. Abundant palynological (Dunn, 1998, 1999) and paleobotanical (Dunn, 1997; Naugoinykh, 1999) remains offer great potential for correlation to continental sections. The co-occurrence of marine and terrestrial biota is particularly important because the Permian contains extensive continental strata. Thus the Aidaralash GSSP may provide a bridge between marine and continental strata.

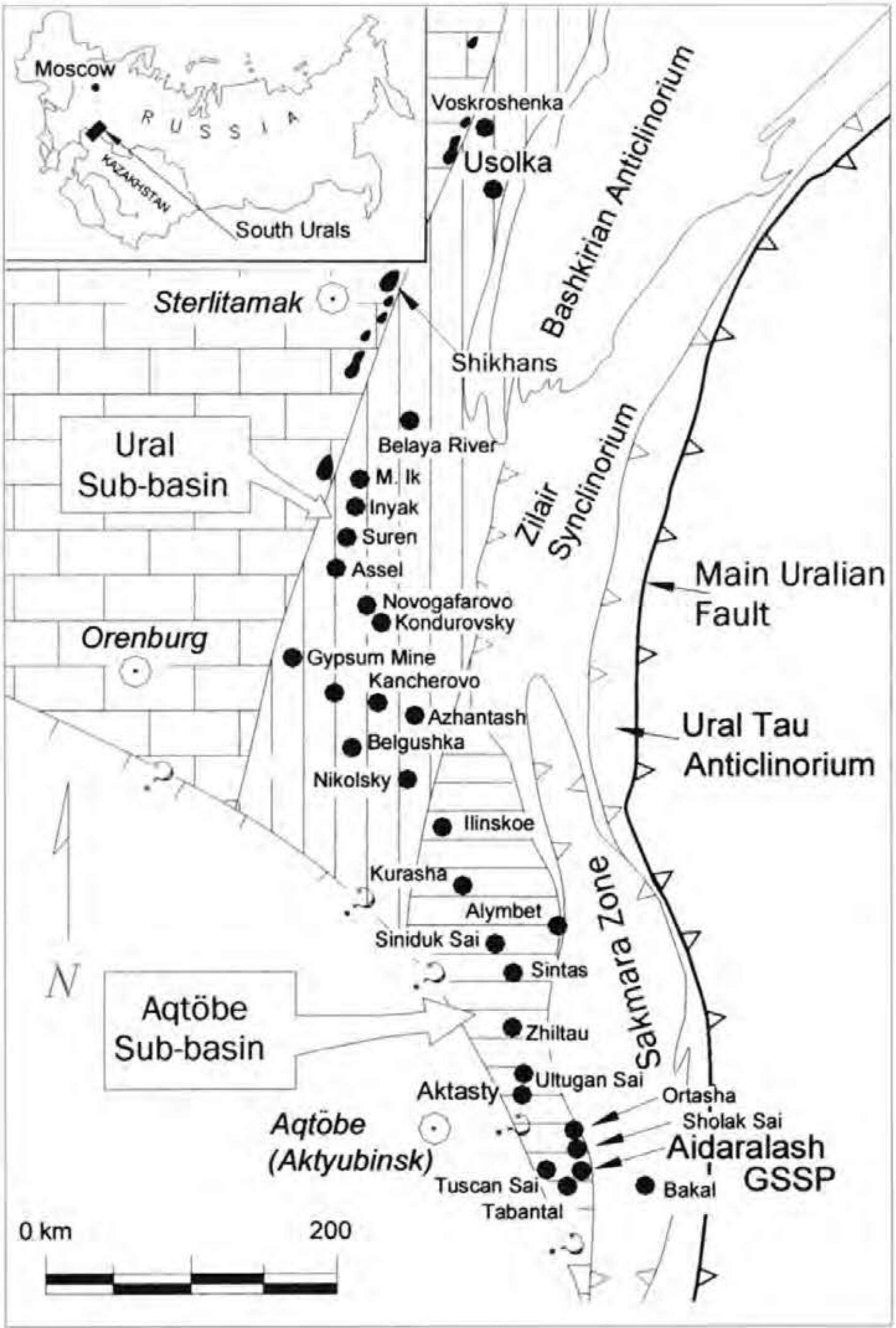

Fig. 1 – Regional Upper Paleozoic tectonostratigraphic map of the southern Ural Mountains and northern Pre-Caspian basin of Russia and Kazakhstan. Limestone pattern depicts the Russian platform. Black dots represent some of the more important Permian and Carboniferous sections of the region. Aidaralash is located in the southern region and Usolka in the north. Modified after Snyder *et al.*, 1994.

PERMIAN-CARBONIFEROUS BOUNDARY DEFINITIONS

Conodonts

The official boundary definition for the base of the Asselian Stage, and thus the base of the Permian and Carboniferous-Permian boundary, is based on the First Appearance Datum (FAD) of the conodont *Streptognathodus isolatus* that occurs 27 m above the base of bed 19 (Fig. 2).

Fusulinaceans

Previous fusulinacean boundary definitions (Ruzhencev, 1936, 1950) placed the base of the Permian in bed 19, located 6.3 m above the conodont-defined boundary (Fig. 2) and coincident with the base of the *Sphaeroschwagerina vulgaris aktjubensis-S. fusiformis* Zone that overlies the *Ultradaixina bosbytauensis-Schwagerina robusta* Zone. Fusulinaceans are abundant and diverse at Aidaralash and detailed evolutionary lineages are well preserved. However, fusulinacean species from Aidaralash, although widespread regionally, appear to be somewhat provincial and therefore have limited potential for international correlation.

Ammonoids

Ammonoid definitions of the Permo-Carboniferous boundary (Bogoslovskaya *et al.*, 1995) were based on the first appearance of the characteristically Permian families Metalegoceratidae, Paragastrioceratidae and Popanoceratidae. The *Juresanites-Svetlanoceras* genozone was considered the base of the Asselian, which overlies the *Shumardites-Vidrioceras* Genozone. Ammonoids are abundant at Aidaralash (Fig. 3) and the distribution of ammonoid genera is widespread, however, at the species level, similarly to fusulinaceans, ammonoid distribution is somewhat locally restricted.

Palynology

Approximately fifty meters of section (Fig. 4) across the Carboniferous-Permian boundary were sampled for palynomorphs (Dunn, 1998). No significant change was noted in the pollen and spore assemblage within this portion of the Aidaralash section. The palynoflora of the section studied is characterized by abundant and diverse species of *Vittatina* (polyplicate pollen), and a variety of taeniate disaccate pollen genera. As shown in Figure 5 these two suprageneric groups comprise 64% of the palynomorphs in the assem-

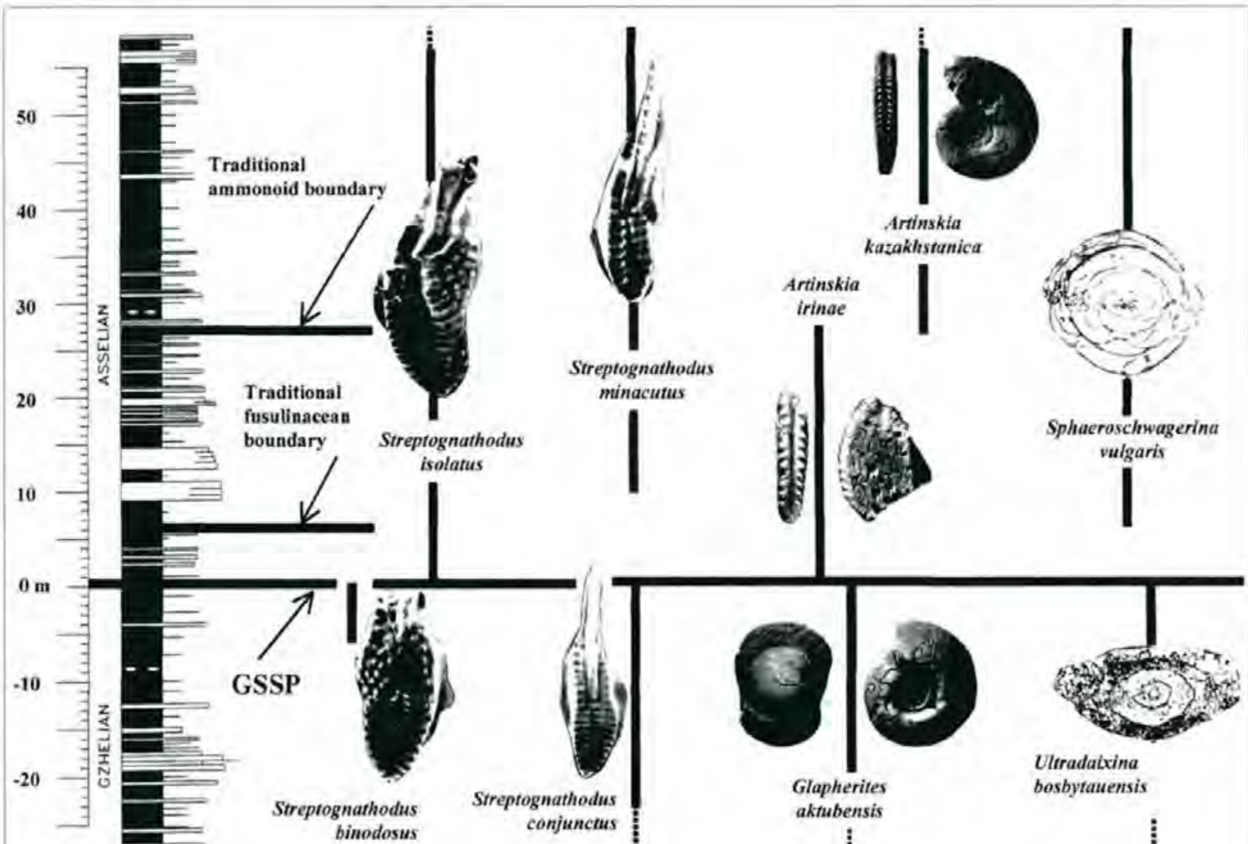

Fig. 2. – Stratigraphic column and ranges of selected representatives of significant invertebrate fossil groups in the GSSP for the base of the Permian System, Aidaralash Creek, Aktöbe (formerly Aktyubinsk) region, Northern Kazakhstan. Shown are the traditional boundaries based on ammonoids, fusulinaceans and the GSSP definition based on conodonts.

blage. Monosaccate and non-taeniate pollen comprise 8% and 7% of the population respectively. In situ trilete spores account for 13% and reworked Devonian and Lower Carboniferous trilete spores total more than 7%. The co-occurrence of *Limitisporites monstruosus* and *Vittatina costabilis* in conjunction with the abundance and diversity of taeniate disaccate and polyplicate pollen suggests that the palynological assemblage at Aidaralash Creek may be correlatable with the *Limitisporites monstruosus* - *Vittatina costabilis* Assemblage Zone of Utting (1989, 1994).

Important taxa include *Vittatina costabilis* Wilson, *Vittatina vittifera* (Lyuber) Samoilovich ex Hart, *Vittatina simplex* Jansonius, *Vittatina saccifer* Jansonius, *Vittatina subsaccata* Samoilovich, *Hamiapollenites bullaeformis* (Samoilovich) Jansonius, *Hamiapollenites tractiferinus* Samoilovich, the *Protohaploxypinus latissimus-Protohaploxypinus perfectus* complex, *Striatoabieites* sp., *Limitisporites monstruosus* Lyuber & Val'ts, *Illinites unicus* Kosanke, the *Lueckisporites-Scutasporites* complex, *Potonieisporites* spp., *Florinites luberae* Samoilovich, and *Cordaitina uralensis* Lyuber & Val'ts (Fig. 6). The most common trilete spores are generally not biostratigraphically useful, such as *Punctatisporites*, *Calamospora* and *Leiotriletes*, or are reworked such as *Densosporites* and *Apiculatisporis*.

Dyupina (1975) recognized that an increase in the quantity of acritarchs and a reduction in the diversity and quality of preservation of miospores correlate with an onset of a transgressive phase. At Aidaralash, acritarchs (*Inderites*) do not increase across the Permian-Carboniferous boundary, thus Dyupina's hypothesis is in agreement with the physical evidence (Snyder *et al.*, 1994) that no discernable transgressive cycle occurs in strata within 50 meters of the GSSP at Aidaralash.

The abundance of disaccate pollen and the polyplicate *Vittatina* with the relative paucity of spores and monosaccate pollen suggests that the macroflora of the Carboniferous-Permian boundary at Aidaralash Creek was dominated by gymnosperms. This gymnosperm dominance suggests that the terrestrial ecosystem of the region was that of an arid dry upland – typically Permian – rather than the coal swamps typical of the Upper Carboniferous. The dominance of gymnosperm pollen indicates that the aridity-tolerant Permian flora had been well established by the beginning of the Permian and that the change had occurred prior to the close of the Carboniferous. These data support the findings of Ziegler (1990) and of Utting & Piasecki (1995) who suggested that the climate of the southern Urals in Early Permian time was that of an arid tropical desert.

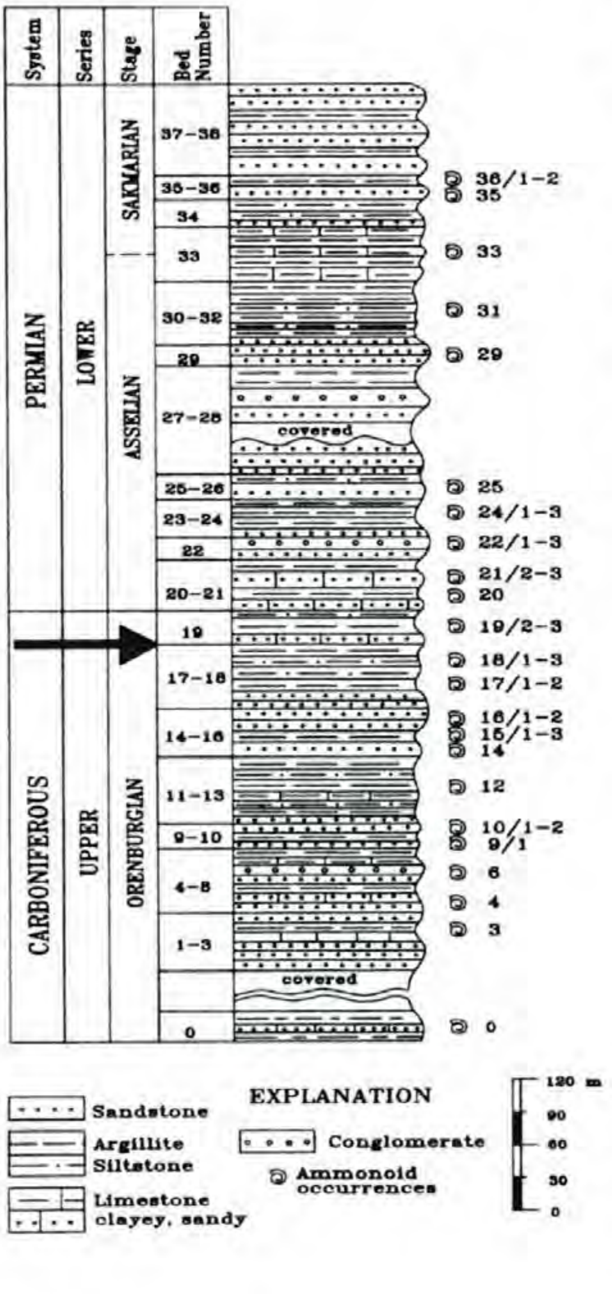

Fig. 3 – Diagrammatic columnar representation of the Aidaralash Creek section with indicated ammonoid occurrences and the Permian-Carboniferous boundary preferred by ammonoid workers. Arrow indicates location of internationally recognized GSSP. Bed numbers represent the terminology used in Russian literature. Adapted from Bogoslovskaya *et al.*, 1995.

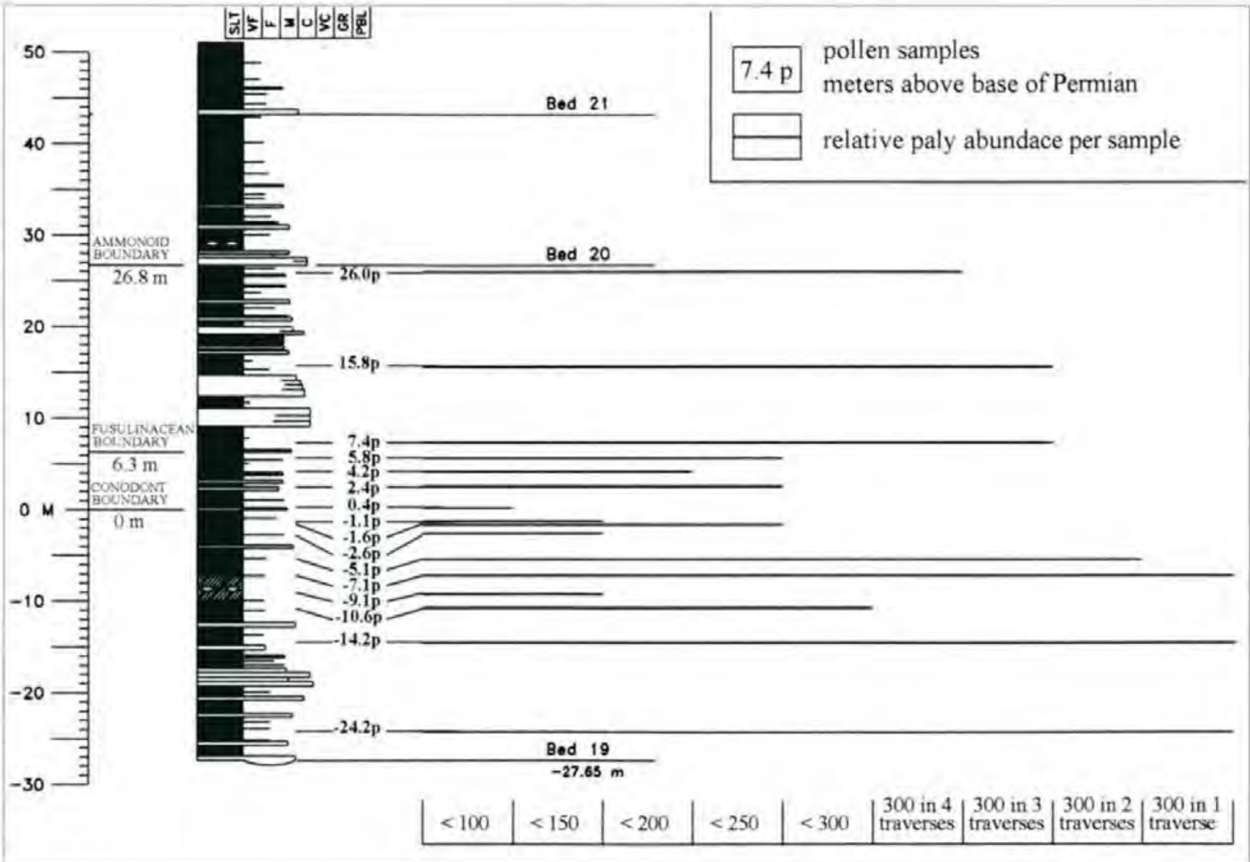

Fig. 4 – Relative abundance and location (meters above base of Permian) of palynomorph samples from a fifty-meter interval of the GSSP section at Aidaralash Creek, Aktöbe (formerly Aktyubinsk) region, Northern Kazakhstan. Adapted from Dunn, 1999.

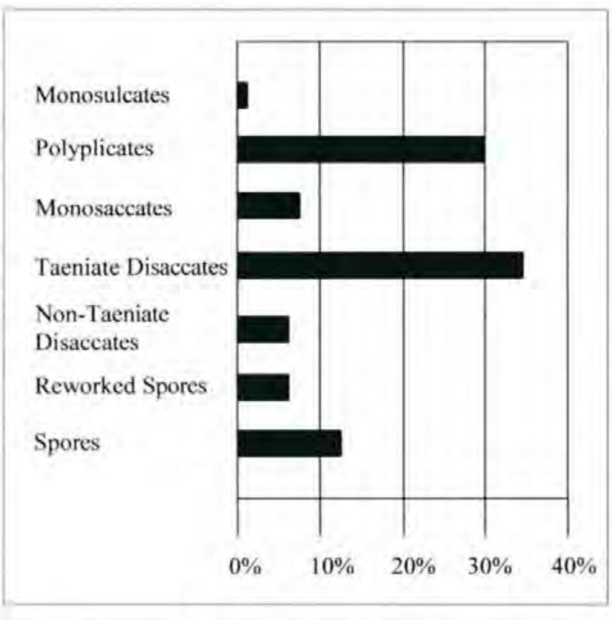

Fig. 5 – Percentage of suprageneric palynomorph groups from Aidaralash Creek, Aktöbe (formerly Aktyubinsk) region, Northern Kazakhstan.

Acknowledgements – We thank the staff of the Institute for Sedimentary and Petroleum Geology of the Geological Survey of Canada, Calgary, Alberta, for their hospitality and assistance; in particular Brenda Davies who processed the palynology samples and John Utting for his advice and critical review of the palynology. Paula Pittau and John Utting reviewed the manuscript and made significant improvements. Partial funding of this project was provided by the Geological Society of America Grants-in-Aid-Program (Student Research Grant No. 6019-97 to Dunn). The National Science Foundation (Grant EAR-9219428 to Snyder & Spinosa), and the Petroleum Research Fund (Grant 30215-AC8 to Snyder) provided funding for fieldwork. Boise State University provided financial support for the senior author. W.S. Snyder, V. I. Davydov & D. M. Gallegos provided field assistance.

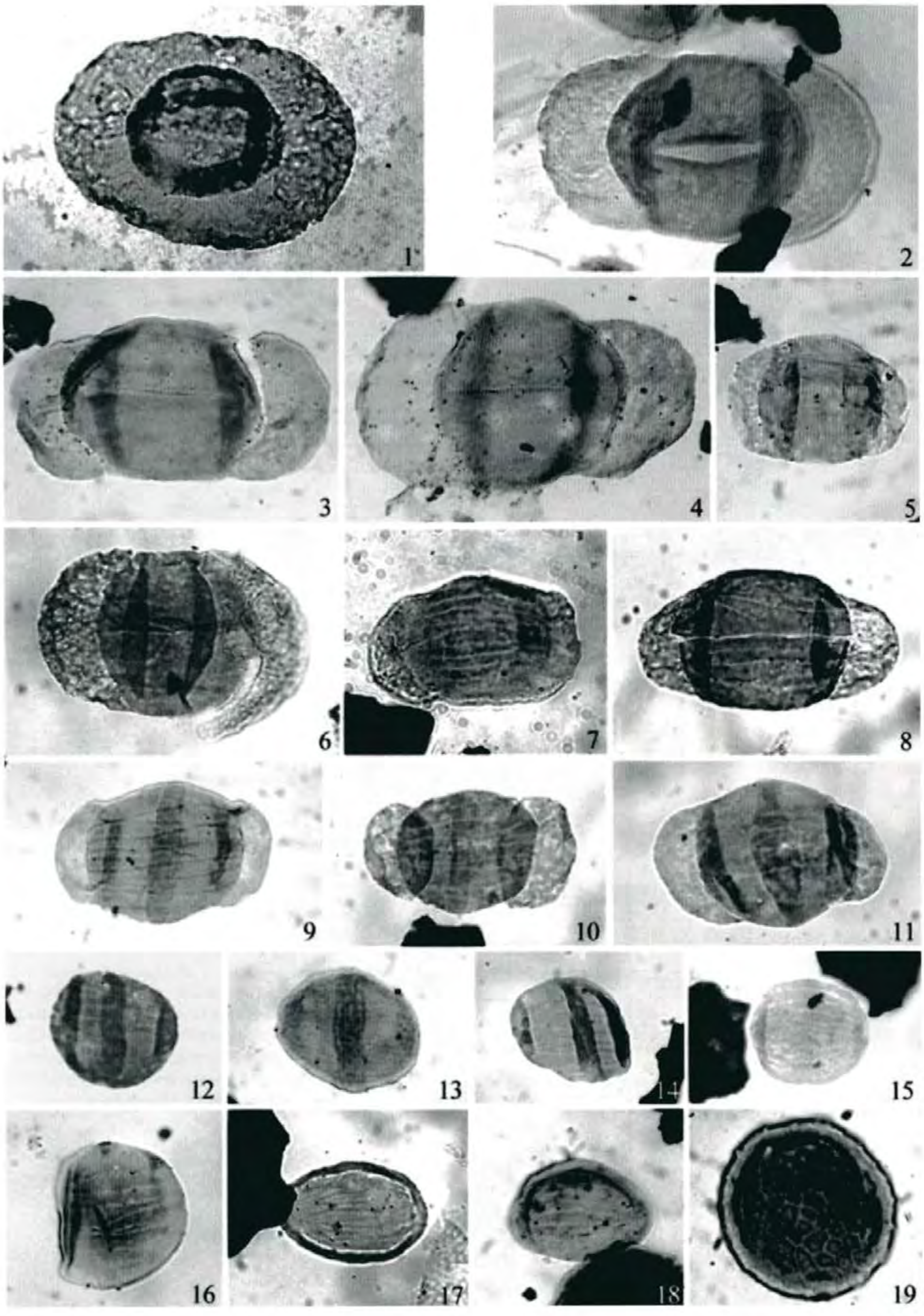
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Fig. 6 – Selected characteristic palynomorphs from the Aidaralash Creek section. Specimens are referenced by name, Geological Survey of Canada processing number, meters above the base of the Permian, and stage coordinates of Olympus Vanox microscope (BSU#31597), respectively. All photomicrographs are at a magnification of 500X.

6.1.	*Potonieisporites novicus* Bharadwaj, 1954;	4201-2, +20: -7.1MAB; 21x104
6.2.	*Limitisporites monstruosus* Lyuber and Val'ts 1941;	4201-2, +20: -7.1MAB; 12x92.
6.3.	*Limitisporites monstruosus* Lyuber and Val'ts 1941;	4202-8, +20: 26.0MAB; 21x86.
6.4.	*Limitisporites monstruosus* Lyuber and Val'ts 1941;	4202-6, +20: 5.8MAB; 20x70.
6.5.	*Protohaploxypinus latissimus-Protohaploxypinus perfectus* complex;	4202-3, +20: -9.1MAB; 13x84.
6.6.	*Protohaploxypinus latissimus-Protohaploxypinus perfectus* complex;	4201-7, +20: 4.2MAB; 11x87.
6.7.	*Protohaploxypinus latissimus-Protohaploxypinus perfectus* complex;	4202-8, +20: 26.0MAB; 18x73.
6.8.	*Striatoabieites spA*;	4204202-1, +20: -24.2MAB; 15x98.
6.9.	*Hamiapollenites bullaeformis* (Samoilovich) Jansonius 1962;	4202-1, +20: -24.2MAB; 19x74.
6.10.	*Hamiapollenites bullaeformis* (Samoilovich) Jansonius 1962;	4201-1, +20: -24.2MAB; 6x87.
6.11.	*Hamiapollenites bullaeformis* (Samoilovich) Jansonius 1962;	4201-1, +20: -10.6MAB; 21x79.
6.12.	*Vittatina costabilis* Wilson 1962;	4201-8, +20: 7.4MAB; 9x99.
6.13.	*Vittatina costabilis* Wilson 1962;	4202-7, +20: 15.82MAB; 18x81.
6.14.	*Vittatina costabilis* Wilson 1962;	4202-8, +20: 26.0MAB; 10x100.
6.15.	*Vittatina subsaccata* Samoilovich 1953;	4201-1, +20: -10.6MAB; 19x76.
6.16.	*Vittatina simplex* Jansonius 1962;	4202-3, +20: -9.1MAB; 10x98.
6.17.	*Vittatina vittifera* (Lyuber) Samoilovich ex Hart 1965;	4202-1, +20: -24.2MAB; 15x94.
6.18.	*Vittatina vittifera* (Lyuber) Samoilovich ex Hart 1965;	4202-7, +20: 15.82MAB; 13x88.
6.19.	*Inderites* sp. (Abromova and Marchenko) Dyupina, 1970;	4202-6, +20: 5.8MAB; 2.5x81.

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PERMIAN AND EARLY TRIASSIC PALYNOMORPH ASSEMBLAGES FROM CANADIAN ARCTIC ARCHIPELAGO, ALASKA, GREENLAND, AND ARCTIC EUROPE

JOHN UTTING[1]

Key words – Palynomorphs; Permian; Early Triassic; Arctic Canada; Alaska; Greenland; Arctic Europe.

Abstract – In the Sverdrup Basin of the Canadian Arctic Archipelago two pollen and spore zones have been established in Kungurian? to Wordian rocks at the basin margin. A further zone occurs in unconformably overlying beds which, based on ammonoids, have been assigned traditionally to the lower Griesbachian (lower Triassic), although recently some workers have suggested that the lowest of these beds may be uppermost Permian (Changhsingian), based on conodont data.
The palynomorph assemblages in the zones of Kungurian to Roadian, Wordian and Griesbachian age resemble those of approximately similar age on the North Slope of Alaska, northern and eastern Greenland, Spitsbergen, Svalis Dome, Finnmark, and Kolguyev Island, thereby enabling relatively precise age determinations, in terms of North American stages, to be made throughout the circum-polar area.
Many workers correlate the Ufimian stratotype in the Volga-Urals region of Russia with the Roadian of Texas, and the Kazanian stratotype of Russia with the Wordian and Capitanian of Texas. However, precise correlation between Texan and Russian stratotypes is hampered by the fact that, whereas the former contain a rich marine fauna, the Russian stratotypes contain only sparse brachiopods, ammonoids, corals and conodonts and a non-marine fauna of bivalves, ostracodes and conchostracans. Palynological comparisons cannot yet be made with assemblages from the Texan stratotypes because published data are unavailable, largely because the abundant carbonates lack palynomorphs.
However, palynomorphs are abundant in Kungurian? to Wordian rocks of the Canadian Arctic that also contain marine faunas, and in the Russian Ufimian and Kazanian stratotype areas, thus allowing comparisons to be made. These indicate that there are major differences in the composition of pollen and spore taxa of the two areas.
If the age correlations are valid, then the palynological differences may be the result of differences in floral province, paleoclimate, paleolatitude, facies and relative geographic location on Pangea.

Parole chiave – Palinomorfi; Permiano; Triassico inferiore; Artico canadese; Alaska; Groenlandia; Artico europeo.

Riassunto – Nel Bacino di Sverdrup, situato nell'Arcipelago Artico canadese, sono state stabilite due zone a pollini e spore in rocce kunguriane-wordiane poste al margine del bacino. Un'ulteriore zona è presente nei sovrastanti strati discordanti che, in base alla presenza di ammoniti, è stata tradizionalmente assegnata al Griesbachiano inferiore (Triassico inferiore), sebbene recentemente alcuni ricercatori abbiano proposto, sulla base di dati a conodonti, che il più basso di questi strati possa corrispondere al Permiano più alto (Changhsingiano). Le associazioni a palinomorfi nelle zone che vanno dal Kunguriano al Roadiano, Wordiano e Griesbachiano rassomigliano a quelle di età approssimativamente simili presenti sullo *slope* settentrionale dell'Alaska, in Groenlandia settentrionale e orientale, nelle Spitzbergen, nel "Duomo di Svalis", in Finnmark e nell'isola di Kolguyev, consentendo pertanto che determinazioni di età relativamente precise, in termini di piani nord-americani, siano compiute attraverso l'area circum-polare.
Molti ricercatori correlano lo stratotipo relativo all'Ufimiano nella regione Volga-Urali della Russia con il Rodiano del Texas, e lo stratotipo relativo al Kazaniano russo con il Wordiano e il Capitaniano del Texas. Tuttavia, una precisa correlazione tra gli stratotipi texani e russi è ostacolata dal fatto che, mentre i primi contengono una ricca fauna marina, gli stratotipi russi includono solo rari brachiopodi, ammoniti, coralli e conodonti ed una fauna non-marina a bivalvi, ostracodi e concostraci. Confronti palinologici non possono ancora essere fatti con associazioni provenienti dagli stratotipi texani in quanto i dati pubblicati sono inutilizzabili, in gran misura perché gli abbondanti carbonati sono privi di palinomorfi. Tuttavia, i palinomorfi sono diffusi nelle rocce kunguriane?-wordiane dell'Artico canadese, che pure contengono faune marine, e nelle aree a stratotipi dell'Ufimiano e del Kazaniano russi, consentendo pertanto che si possa procedere a confronti. Questi indicano che vi sono maggiori differenze nella composizione di taxa a pollini e spore delle due aree. Se le correlazioni cronologiche sono valide, le differenze palinologiche possono allora rappresentare il risultato di differenze legate alla provincia floristica, al paleoclima, alla paleolatitudine, alle facies ed alla relativa posizione geografica nell'ambito della Pangea.

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INTRODUCTION

Early to Middle Permian rock units in the marginal facies of the Sverdrup Basin include the Sabine Bay, Assistance and Trold Fiord formations (Figs 1 and 2). The Sabine Bay Formation (Tozer & Thorsteinsson, 1964) was probably deposited relatively rapidly in response to the Melvillian disturbance (Beauchamp *et al.*, 1989 b); it is dominated by sandstone, with intercalations of shale, and with rare coaly and carbonaceous shale intercalations (Fig. 2). The overlying Assistance Formation (Harker & Thorsteinsson, 1960; Fortier *et al.*, 1963; Thorsteinsson, 1974; Nassichuk, 1975) is mainly marine, fossiliferous, calcareous, glauconitic sandstone. It is overlain by the Trold Fiord Formation (Tozer & Thorsteinsson, 1964; Nassichuk, 1975; Beauchamp *et al.*, 1989 a, b), which consists of glauconitic, fossiliferous, calcareous sandstone, with some thin cherty intercalations. There is a hiatus between the Trold Fiord and the overlying lower Triassic Blind Fiord Formation (Tozer, in Fortier *et al.*, 1963); the latter is composed of green and grey shale and siltstone, considered to have been deposited on an offshore shelf (Embry, 1988, 1991).

The Permian formations have been dated with marine faunas of conodonts, brachiopods, and ammonoids (Fig. 2). These data suggest a Kungurian to Roadian age for the Sabine Bay Formation (Waterhouse, 1969; Nassichuk, 1995; Henderson, 1988; Beauchamp *et al.*, 1989 b; Beauchamp *et al.*, in press), a Roadian age for the Assistance Formation (Waterhouse, 1969; Nassichuk, 1995) a Wordian age for the Trold Fiord Formation (Waterhouse, 1969; Waterhouse in Thorsteinsson, 1974; Nassichuk, 1995). A review of the macrofaunal data has recently been provided by Fedorowski and Bamber (in press) for the Trold Fiord and Degerböls . The lower part of the unconformably overlying Blind Fiord Formation was dated as Early Triassic (early Griesbachian) on the presence of *Otoceras concavum* (Tozer, 1967). Conodonts are sparse in the basal beds, but Henderson & Baud (1997) suggested the former correlate with the uppermost Changhsingian of Meishan (China). This conclusion is based on the presence in the lowest 14-18 m of the Otto Fiord South Section, Ellesmere Island, of *Neogondolella* sp. cf. *subcarinata*, *N.* sp. aff. *changxingensis*, *N. meishanensis*, and questionable *N. deflectus*. However, Orchard & Tozer (1997, 1999) questioned this interpretation and recommended a detailed study of conodonts from the beds containing *Otoceras*. Until that is carried out an Early Triassic Griesbachian age is assumed in this paper for the lower part of the Blind Fiord Formation.

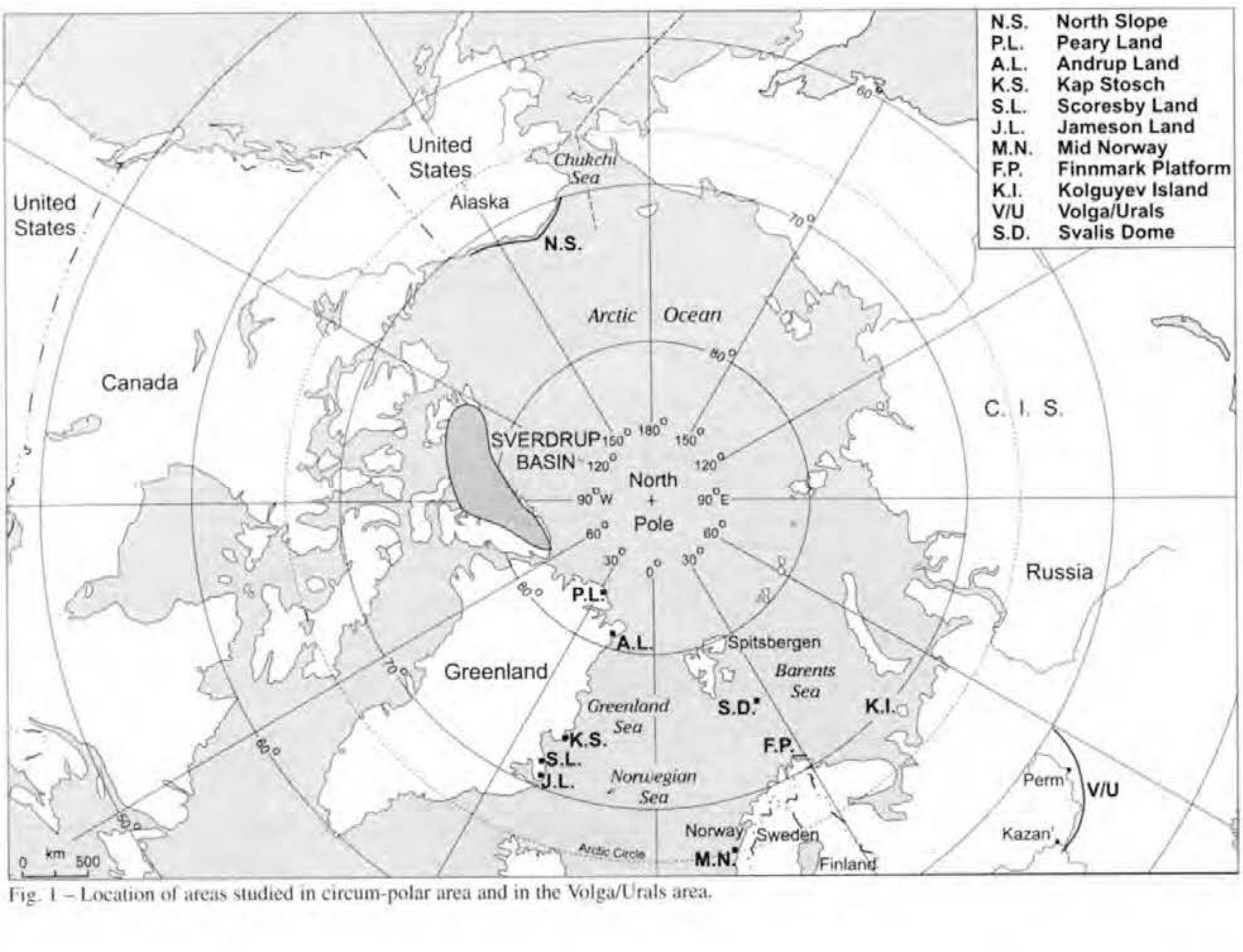

Fig. 1 – Location of areas studied in circum-polar area and in the Volga/Urals area.

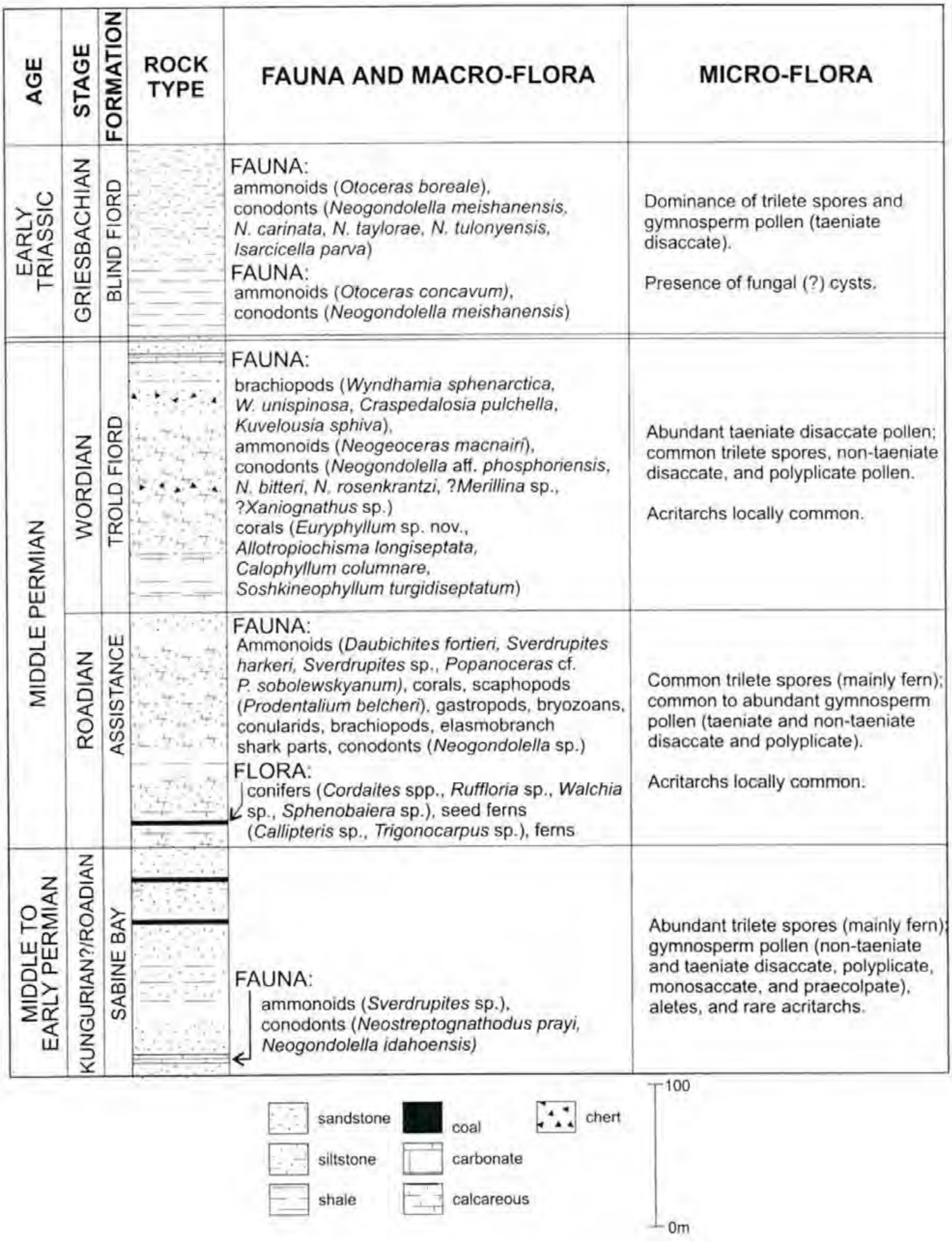

Fig. 2 – Age, formation, rock type, fauna, macro-flora, and micro-flora of rock units in Sverdrup Basin.

FLORAL PROVINCE

Macro-floras are rare in the Permian of the circum-polar study area, and interpretations of floral province are speculative. Mamay & Reed (1984) reported an Angaran flora from the central Alaska Range no younger than Early Permian. Wagner *et al.* (1982) and Wagner *et al.* (in press) described Late Permian (Kazanian) floras from northern Greenland that have Angaran and rare Cathaysian affinities. Recently in the Sverdrup Basin, study of macroflora in bioclastic limestone approximately equivalent to the Assistance Formation near the Svartevaeg Cliffs of northern Axel Heiberg has indicated elements in common with Angaran, European and Euramerian provinces (LePage, pers. comm. 2000). Also present are taxa unknown or rare in the Angaran floras and those that are more typical of the Upper Permian and Lower Triassic. The flora is dominated by cordaites (*Cordaites, Ruffloria, Vojnovskaya* and *Zamiopteris*), but abundant remains of *Utrechtia* indicate that this conifer co-dominated the regional vegetation throughout the extrabasinal regions (LePage, pers. comm. 2000). The flora is characterised by a diverse assemblage of pteridosperms. It contains medullosan seed ferns not normally associated with Angara. Also present are peltasperms, which are the dominant elements of the Upper Permian of Russia. The microflora in associated beds has been assigned to the upper part of the *Alisporites plicatus - Jugasporites compactus* Concurrent Range Zone of Roadian age (Fig. 2). The Sverdrup Basin was assigned to Sub-Angara using palynological criteria (Utting & Piasecki, 1995).

Fig. 3 – Stratigraphic ranges of taxa in palynomorph zones of the Sverdrup Basin.

PALYNOMORPH ZONES OF CANADIAN ARCTIC ARCHIPELAGO (SVERDRUP BASIN) AND COMPARISON WITH SELECTED CIRCUM-POLAR LOCALITIES

In the following sections Kungurian? to Wordian and Griesbachian assemblages of the Sverdrup Basin are compared with those of other circum-polar areas. This does not necessarily imply that the microfloras are restricted to that geographical area. For example, in Canada assemblages similar to the Griesbachian assemblages of the Sverdrup Basin occur some 1800 km to the south in northern Alberta (Jansonius, 1962).

SVERDRUP BASIN

The *Alisporites plicatus - Jugasporites compactus* (Kungurian? to Roadian) and the *Ahrensisporites thorsteinssonii - Scutasporites nanuki* (Wordian) Concurrent Range Zones (Utting, 1994a) are based on the lowest occurrence of a number of diagnostic taxa (Fig. 3). The palynomorph assemblages contain different proportions of trilete spores and gymnosperm pollen (Fig. 2), but fern spores are commonly represented, as are gymnosperm pollen including taeniate and non-taeniate disaccates, polyplicates and monosaccates. Acritarchs are rare to common. The ages of the Sverdrup zones (Figs 4, 5[1-3] and 6[1]) were derived from marine faunal ages summarised above, and are subject to change as more information becomes available concerning the latter.

The diversity of the palynoflora and the abundance of fern spores in the Sabine Bay and Assistance assem-

AGE		NORTHERN URALS-RUSSIAN PLATFORM	U.S.A. TEXAS		CANADA SVERDRUP BASIN	PALYNOMORPH ZONES OF THE SVERDRUP BASIN	FORMATION
TRIASSIC	EARLY	Induan	No data (Ochoan)		Griesbachian	*Tympanicysta stoschiana-Striatoabieites richteri*	Blind Fiord
PERMIAN	LATE	Tatarian	No data (Ochoan)		No data	No data Hiatus at basin margin	
PERMIAN	MIDDLE	Kazanian	Guadalupian	Capitanian	? Capitanian ?	?	?
PERMIAN	MIDDLE	Kazanian	Guadalupian	Wordian	Wordian	*Ahrensisporites thorsteinssonii-Scutasporites nanuki*	Trold Fiord
PERMIAN	MIDDLE	Ufimian	Guadalupian	Roadian	Roadian	*Alisporites plicatus-Jugasporites compactus*	Assistance Sabine Bay
PERMIAN	EARLY (PART)	Kungurian	Leonardian (part)		Kungurian	Not studied in detail	

Fig. 4 – Correlation of stages in Northern Urals-Russian Platform, Texas, U.S.A. and Sverdrup Basin with ages of palynomorph zones and formations.

AGE			STAGE	SELECTED PALYNOMORPH ZONES OF CIRCUM-POLAR AREA	MARINE ZONE FOSSILS
TRIASSIC	EARLY		GRIESBACHIAN (PART)	[5]*Tympanicysta stoschiana - Striatoabieites richteri*	Ammonoids and Conodonts
PERMIAN	LATE	LOPINGIAN	CHANGH-SINGIAN	? ?	None, Age Uncertain
PERMIAN	LATE	LOPINGIAN	WUCHIA-PINGIAN	[4]*Tympanicysta stoschiana*	Ammonoids
PERMIAN	MIDDLE	GUADALUPIAN	CAPITANIAN	[3]*Vittatina*	Conodonts
PERMIAN	MIDDLE	GUADALUPIAN	WORDIAN	? ? [2]*Ahrensisporites thorsteinssonii - Scutasporites nanuki*	Brachiopods, Ammonoids and Conodonts
PERMIAN	MIDDLE	GUADALUPIAN	ROADIAN	[1]*Alisporites plicatus - Jugasporites compactus*	Ammonoids and Conodonts
PERMIAN	EARLY (PART)	CISURALIAN	KUNGURIAN	?	
PERMIAN	EARLY (PART)	CISURALIAN	ARTINSKIAN		

Fig. 5 – Composite stages of the Permian of the world proposed by Jin *et al.*, 1999, and composite palynological zones in circum-polar area with associated marine zonal fossils. [5] *Tympanicysta stoschiana - Striatoabieites richteri* Assemblage Zone Utting 1994 (Sverdrup Basin and most circum-polar localities); [4] Lowest occurrence of *Tympanicysta stoschiana* in E. Greenland (Utting and Piasecki, 1995); [3] *Vittatina* Association Balme 1980a (E. Greenland); [2] *Ahrensisporites thorsteinssonii - Scutasporites nanuki* Concurrent Range Zone Utting 1994 (Sverdrup Basin and most circum-polar localities); [1] *Alisporites plicatus - Jugasporites compactus* Concurrent Range Zone Utting 1994 (Sverdrup Basin and most circum-polar localities).

blages suggest a diverse vegetation growing in a humid climate - an interpretation supported by the presence of thin coal seams and carbonaceous shales in the Sabine Bay Formation. On Axel Heiberg Island, possible equivalents of the Assistance Formation include carbonaceous shale associated with bioclastic limestone, which contain the macroflora described above as well as a microflora dominated by fern spores (Utting, 1994b). A relative decrease of fern spores in samples from the overlying Wordian Trold Fiord Formation suggests an increase in aridity, which is supported by lack of coal seams and the appearance of red beds. Beauchamp (1994) came to similar conclusions based on sedimentological and marine faunal criteria, although he also suggested that the climate in the Wordian was cold. This contrasts with the warm temperate climate proposed by Rees *et al.* (1999), who determined paleoclimates from floral and lithological data globally.

Unconformably overlying early Griesbachian (Early Triassic) beds contain the *Tympanicysta stoschiana - Striatoabieites richteri* Assemblage Zone (Utting, 1994a). Although the top has yet to be defined, the zone is known to locally extend 100 m above the base of the formation. The zone is characterised by trilete spores, abundant taeniate disaccate and rare non-taeniate disaccate and polyplicate pollen. Acritarchs are rare to common. Also present are entities of uncertain affinity (Kalgutkar & Jansonius, 2000), which Elsik (1999) attributed to *Fungi incertae sedis*. These have been described under various names, including *Tympanicysta stoschiana* Balme, 1980a; *Chordecystia chalasta* Foster, 1979 and *Reduviasporonites stoschianus*, Elsik, 1999. Another characteristic of the zone is the common occurrence of well preserved reworked Late Devonian spores.

There are significant differences between the Wordian *Ahrensisporites thorsteinssonii - Scutasporites nanuki* Concurrent Range Zone and the Griesbachian *Tympanicysta stoschiana - Striatoabieites richteri* Assemblage Zone. They have many genera, but virtually no species in common (Fig. 3). At some localities the basal part of the *Tympanicysta stoschiana - Striatoabieites richteri* Assemblage Zone is dominated by the trilete spore *Uvaesporites imperialis* (Jansonius) Utting, which suggests arid conditions. Visscher *et al.* (1996) pointed out that microspores of herbaceous and subarborescent lycopodiophytes were commonly abundant after the extinction event at the Permian/Triassic boundary. If the suggestion by Henderson & Baud (1997) that Changhsingian rocks are present in the lower part of the Blind Fiord Formation is correct then one might anticipate some changes in the palynomorph assemblages, but other than the abovementioned dominance of *Uvaesporites imperialis*, this does not occur in the sections investigated, including the Otto Fiord South section, Ellesmere Island studied by Henderson & Baud (1997).

With the exception of the basal samples, most contain assemblages with abundant taeniate disaccate pollen known to be well adapted to arid conditions (Foster, 1979). This is consistent with the presence of red shale and siltstone of overbank origin (Embry, 1991), and small carbonate nodules suggesting poorly developed caliches (Devaney, 1991) that indicate a seasonally dry, hot, subtropical savanna-type climate. *Tympanicysta stoschiana*, although present in many Blind Fiord samples, is generally rare, and there is no marked peak in abundance as occurs in some localities of the Italian Southern Alps, such as the Butterloch and Tesero sections (*e.g.* Conti *et al.*, 1986; Massari *et al.*, 1988, 1996, 1999) and the Badia Valley (Cirilli *et al.*, 1998). Recent work in other localities in northern Italy by Cirilli (pers. comm., 2000) indicates that abundance of *T. stoschiana* is not a single peak occurrence near the Permian/Triassic boundary, but may correspond to the upper parts of shallowing upwards sequences. This is not inconsistent with the gradual carbon-isotope shift at the boundary (Magaritz *et al.*, 1988). The suggestion was made by Visscher *et al.* (1996) that world-wide proliferation of *Tympanicysta stoschiana* in terrestrial and marine facies at the Permian-Triassic boundary implies an excessive heterotrophic presence, and reflects terrestrial ecosystem destabilization and collapse.

ALASKA (NORTH SLOPE)

The Sadlerochit Group in the Prudhoe Bay, North Slope area includes beds that have been assigned to the Permian and the Triassic (Moore *et al.*, 1994). In the lower part of the group is the Echooka Formation, which includes the lower part of the Joe Creek Member, consisting of calcareous mudstone, radiolarian chert and bioclastic, glauconitic limestone, overlain by the Ikiapauruk Member, consisting of glauconitic quartzose sandstone and siltstone (Moore *et al.*, 1994; Jones & Speers, 1976). These rock types resemble those of the Roadian (Assistance Formation) and Wordian (Trold Fiord Formation) of the Sverdrup Basin. The Joe Creek Member has been dated by a brachiopod fauna as Sakmarian to Kazanian, and the Ikiapauruk Member as Kazanian or late Guadalupian (Detterman *et al.*, 1975), or as Ufimian (late Leonardian) to Wordian (Dutro & Silberling, 1988). The Echooka Formation has been dated by ammonoids as Roadian or Wordian (Nassichuk, 1995). There is little published palynological data from this unit, although a mid-Permian age has been suggested based on the common occurrence of *Vittatina* and taeniate disaccate pollen and the lack of *Lueckisporites virkkiae* Potonié and Klaus (Jones & Speers, 1976).

A core sample of the Echooka(?) Formation from Kaparuk State Mobil #1 (17612 - 17621 ft. = 5370.88 - 5374.54 m), loaned to the present writer by Dr. B. Balme, University of Western Australia, yielded an assemblage with abundant gymnosperm pollen (53%), abundant trilete spores (35%), and common acanthomorph acritarchs (12%) consisting mainly of *Micrhystridium* spp. and rare *Veryhachium* spp. (Fig. 6[2]).

The assemblage comprises the following taxa:

Gymnosperm pollen:

Taeniate disaccates (18%): *Corisaccites stradivarius* Utting, *Protohaploxypinus latissimus* (Luber) Samoilovich, *P. perfectus* (Naumova) Samoilovich, *Protohaploxypinus* sp., *Striatoabieites striatus* (Luber & Waltz) Hart

Non-taeniate disaccates (18%): *Alisporites plicatus* Jizba, *A. splendens*(Leschik) Foster, *Vitreisporites pallidus* (Reissinger) Nilsson

Polyplicates (15%): *Weylandites cincinnatus* (Luber ex. Varyukhina) Utting, *W. striatus* (Luber) Utting

SELECTED TAXA	SVERDRUP BASIN[1]			ALASKA NORTH SLOPE[2-4]			NORTH GREENLAND: AMDRUP LAND[5]	NORTH GREENLAND: SLETTEN[5,6] PEARY LAND			EAST GREENLAND: KAP STOSCH, JAMESON LAND AND SCORESBY LAND[7-9]				SPITSBERGEN[10-15]			SVALIS DOME[16,17]			FINNMARK PLATFORM[18]			KOLGUYEV ISLAND[19]		MID-NORWAY[20]	VOLGA/URALS[21]	
	K/U	Ka	G	U	"G"	G	Ka	K/U	Ka	G	Ka	KW	W	G	K/U	Ka	G	K/U	Ka	G	K/U	Ka	G	U	Ka	G	U	Ka
Aratrisporites tenuispinosus																	X						X					
Proprisporites pocockii			X		X	X						X		X			X			X			X					
Klausipollenites schaubergeri					A	X					X						X											
Tympanicysta stoschiana			X			X						X	X	X			X						X					
Striatoabieites richteri			C		A	A						X		A			X						X					
Ephedripites steevesiae			X			A				X				A			X						X					
Klausipollenites staplinii			X											X		X					X	X	X					
Lunatisporites noviaulensis			A		X	A				X	X	X		A		X	X		X	X		X	X			X		
Protohaploxypinus samoilovichii			C		A	X					X	X		X		X	X		X									
Uvaesporites imperialis			A			A											X						X					
Densoisporites nejburgii														X		X	X			X			X					
Densoisporites playfordii														X			X						X			X		
Maculatasporites spp.			X													X	X			X			X					
Lundbladispora obsoleta														X			X						X			X		
Cycadopites spp.	X		X			A			X	X				X		X	X		X	X			X					
Lueckisporites virkkiae					"A"	"X"					X	X		"X"		X	"X"		X		X	X	"X"					X
Ahrensisporites thorsteinssonii		X							X													X	X		X			
Ahrensiporites multifloridus		X							X										X						X			
Lunatisporites beauchampii		X					X												X									
Piceapollenites nookapii		X					X		X																			
Scutasporites nanuki		X					X		X			X	X						X			X			X			
Striatoabieites borealis		C					X		X																			
Hamiapollenites erebi		X					X		X																X			X
Crinalites sabinensis	X	X																										
Cladaitina kolodae	X	X					X	X							X	X			X		X	X						
Sverdrupollenites agluatus	X	X																										
Sverdrupollenites connudatus	X	X					X																					
Alisporites plicatus	X	X		X			X		X																		X	X
Dyupetalum vesicatum	X	X					X														X			X	X			
Hamiapollenites bullaeformis	X	X					X												X								X	X
Jugasporites compactus	X	X																										
Diatomozonotriletes igluanus	X	X		X											X	X											X	
Corisaccites stradivarii	X	X		X			X																					
Weylandites striatus	C	X		X			X	X			X	A			C	C		X	X					X			X	X
Weylandites cincinnatus	X	X		X			X																					
Vittatina spp.	C	X	"X"				X	C	X		A	A	X	"X"	C	C	"X"	X	X	"X"	X	X	"X"	X			X	X
Leiotriletes ulutus	X	X		X			X									X		X				X		X	X			
Protohaploxypinus perfectus	X	X		X			X	C							X	X			X		X	X		X	X		X	X
Hamiapollenites tractiferinus	X	X																									X	X
Kraeuselisporites sverdrupensis	X	X		X				X	X						X	X		X	X					X	X			
Scutasporites unicus									X																			
Inaperturopollenites nebulosus		X						X			C	A				X	X		X			X	X					
Unellium spp.	X	C					X		X						A	A	X		X		X	X						
Micrhystridium spp.	X	C		X			X		C	C		C		A	A	A	D	X	X	X	X	X	X					
Veryhachium spp.	X	C		X					X	X		C		A	A	A	X		X	X	X	X	X					
Baltisphaeridium spp.								X						A														
Tasmanites spp.								X				A		A			X				X		X					

Fig. 6 – Selected taxa present in circum-polar localities.

K/U = Kungurian? to Ufimian (Roadian); Ka = Kazanian (Wordian and Capitanian); KW = Kazanian to early Wuchiapingian; W = Wuchiapingian; **"G" = Griesbachian with abundant Late Permian reworking; G = Griesbachian**

Monosaccates (1%): *Florinites luberae* Samoilovich
Monosulcates (1%): *Cycadopites* sp.

Trilete spores:
Apiculatisporis melvillensis Utting, *A.* sp.,
Densosporites sp., *Gordonispora obstaculifera* Utting,
Kraeuselisporites sverdrupensis Utting, *Leiotriletes ulutus* Utting, *Lophotriletes parryensis* Utting,
Neoraistrickia sp., *Raistrickia* sp., *Waltzispora* sp.

Some palynomorphs are fairly well preserved in spite of a moderate Thermal Alteration Index of at least TAI 3 suggesting that the organic matter is in the dry gas generation zone (Utting *et al.*, 1989). Although many specimens have mild to severe damage caused by the growth of sulphide pseudomorphs on the exines, sufficient taxa can be identified to tentatively propose a correlation with the *Alisporites plicatus - Jugasporites compactus* Concurrent Range Zone (Kungurian? to Roadian) of the Sverdrup Basin (Figs 3, 4 and 6). No taxa diagnostic of the Wordian *Ahrensisporites thorsteinssonii - Scutasporites nanuki* Concurrent Range Zone were seen, but such an age cannot be completely ruled out in view of the limited amount of material studied and the fact that some specimens are too poorly preserved for identification.

The unconformably overlying Ivishak Formation consists of fine to coarse grained clastic rocks deposited in marine and non-marine environments (Moore *et al.*, 1994). The Kavik Member, abruptly overlying the Echooka Formation, consists of dark-coloured, laminated to thin bedded, silty shale and siltstone, lithologically resembling the Blind Fiord Formation of the Sverdrup Basin. Ivishak rocks represent pro-delta deposits that grade upwards into massive deltaic sandstone and conglomerate of the Ledge Sandstone Member, similar sedimentologically to the Bjorne Formation of the Sverdrup Basin.

In outcrop the Kavik Member has been dated as Griesbachian by ammonoids (*Otoceras* and *Ophiceras*) and pelecypods (*Claraia*) (Detterman *et al.*, 1975; Jones & Speers, 1976), but in the subsurface, rocks of the same member have been dated by palynology as late Permian (Jones & Speers, 1976; Balme, 1980 b and pers. comm. 2000). The palynological assemblages (Fig. 6[14]) are very different in content from the Roadian? Echooka Formation, indicating a significant hiatus below the Kavik Member. Core samples from one locality (BP 9-11-13) contain abundant *Lueckisporites virrkiae* (up to 25%), but in most boreholes there is a variety and abundance of gymnosperm pollen, and rare trilete spores. Reworked Late Devonian spores are abundant (Jones & Speers, 1976; Balme, 1980 b and pers. comm. 2000).

Kavik taxa include *Striatoabieites richteri* (Klaus) Hart, *Lueckisporites virrkiae* (variants A, B and C of Clarke, 1965; Norms Aa to AC of Visscher, 1971), *Klausipollenites schaubergeri* (Potonié and Klaus) Jansonius and *Protohaploxypinus samoilovichii* (Jansonius) Hart. The assemblage was dated as Upper Permian (Tatarian) largely on the basis of abundant *Lueckisporites virrkiae*, *Klausipollenites schaubergeri* and *Striatoabieites richteri* and comparisons were made with the Upper Permian Zechstein microflora of Europe (Balme, 1980b). However, the fact that rare, but stratigraphically significant taxa, such as broadly taeniate *Lunatisporites* spp. ("*Taeniaesporites*"), *Proprisporites pocockii* Jansonius, *Ephedripites steevesiae*, *Uvaesporites imperialis* and *Tympanicysta stoschiana* (Balme, pers. comm. 2000) occur sporadically in core samples, suggests that another interpretation is possible and that all the Kavik Member assemblages are Griesbachian rather than Late Permian. With the exception of *L. virrkiae* and *K. schaubergeri*, all of the taxa listed above occur in the *Tympanicysta stoschiana - Striatoabieites richteri* Assemblage Zone of the Sverdrup Basin. Dating these subsurface assemblages as Griesbachian would eliminate the age discrepancy between the sub-surface and outcrop localities of the Kavik Member. However, if this interpretation proves correct then it would raise the important question of the age, provenance and significance of the abundant specimens of *Lueckisporites virrkiae*, normally characteristic of Upper Permian Zechstein assemblages of western Europe (*e.g.* Visscher, 1971).

The overlying Ledge Sandstone Member, which is partly a lateral facies equivalent of the Kavik Member, is virtually devoid of macrofauna or microfauna, but in the subsurface, samples from the Ledge Sandstone Member and the upper Kavik Member, contain rich palynomorph assemblages (Fig. 6[1]) with the following taxa (Balme 1980 b, pers. comm. 2000):

Gymnosperm pollen:
Taeniate disaccates: abundant *Striatoabieites richteri* (Klaus) Hart, *Lunatisporites* spp. (*Taeniaesporites* spp.), *Protohaploxypinus* spp, *Striatopodocarpites* spp., rare *Lueckisporites virrkiae* Potonié and Klaus
Non-taeniate disaccates: *Alisporites* spp.
Monosulcates: common (in some samples)
Ephedripites steevesiae (Jansonius) de Jersey and Hamilton, *Cycadopites* sp.

Trilete spores:
Uvaesporites imperialis (Jansonius) Utting,
Proprisporites pocockii Jansonius

***Incertae sedis*:**
Tympanicysta stoschiana Balme

Reworking:
common Late Devonian; Late Permian

The assemblage is comparable with those in rocks dated as Griesbachian by marine fauna from Kap Stosch, East Greenland (Balme,1980 b), and is similar to those recorded from the Griesbachian *Tympanicysta stoschiana - Striatoabieites richteri* Assemblage Zone. Thus a Griesbachian age is indicated, although a tentative Late Permian to Early Triassic age was proposed by Jones & Speers (1976).

The fact that reworking of Upper Devonian spores was prevalent in northern Alaska throughout deposition of the Ivishak Formation indicates that erosion of the Upper Devonian persisted through the early Griesbachian as the marine transgression took place. This is not surprising as the existence of Upper Devonian rocks in northern Alaska and the Sverdrup Basin is well documented (Ziegler, 1988). If the postulated Griesbachian age for the subsurface Kavik Member is correct, then there is presently no biostratigraphic evidence for Upper Permian rocks in this area. The nearest known Upper Permian locality with "Zechstein" affinities within the study area is East Greenland, where shallow marine carbonate and gypsum of the Karstryggen Formation contain many species (including *L. virrkiae*) in common with the Zechstein beds of Europe (see East Greenland below). However, in view of the Late Permian hiatus commonly present at the basin margin where most of the studied localities occur, it is possible that Upper Permian Zechstein soft evaporitic and argillaceous rock types, with an abundant micro-flora including abundant *L. virrkiae*, may have been deposited in many localities, but later relatively easily eroded during the Griesbachian transgression. Thus abundance of *L. virrkiae* in the lower Kavik Member may correspond to the active erosional phase, the numbers being dramatically reduced in the upper part of the member and in the Ledge Sandstone Member as the advancing transgression buried, or completely removed the Upper Permian rocks.

This problem of reworking versus *in situ* Griesbachian occurrence of species such as *L. virrkiae* was discussed in some detail by Ouyang & Utting (1990) for the Permian/Triassic boundary at Meishan in China. They postulated that *L. virrkiae* may have survived in refuges into the Griesbachian. It is of course possible for *L. virrkiae* to be reworked from the Upper Permian, and also for its stratigraphic range to extend into the Griesbachian. Thus, with such a scenario all rare and sporadic occurrences of *L. virrkiae* in rocks dated by marine fauna as Griesbachian, such as those in the *Protohaploxypinus* Association of East Greenland (Balme,1980 a), may be either reworked or *in situ*.

Overlying the Griesbachian is an assemblage with *Aratrisporites tenuispinosus* Playford; a Smithian (pers. comm. Balme 2000) age is supported by the presence of the ammonoid *Euflemingites* (Silberling, 1971).

NORTH GREENLAND (WANDEL SEA BASIN)

At Ingolf Fiord (lat. 80° 30'N; long. 18° W) in Amdrup Land (Fig. 1) an un-named unit, approximately 65 m thick, consisting of sandstone and siltstone, contains taxa diagnostic of the *Ahrensisporites thorsteinssonii - Scutasporites nanuki* Concurrent Range Zone (personal observation); these include *Lunatisporites beauchampii* Utting, *Scutasporites nanuki* Utting and *Hamiapollenites erebi* Utting (Figs 4 and 6^{5}).

At Sletten (lat. 82°40' N, long. 22° W) in Peary Land (Fig. 1), the Kim Fjelde Formation consists of well-bedded chert-rich, biogenic limestone, dated by foraminifers and conodonts as Artinskian to Kungurian (Stemmerik *et al.*, 1996). The Sletten palynomorph assemblage is relatively poor but contains *Cladaitina kolodae* Utting (*Maculatasporites* sp.), common *Vittatina* spp., *Weylandites striatus* (Luber) Utting, *Protohaploxypinus perfectus* (Naumova) Samoilovich, *Kraeuselisporites sverdrupensis* Utting, *Inaperturopollenites nebulosus* Balme, and rare spinose acritarchs (Stemmerik *et al.*, 1996; this paper; Fig. 6^{7-8}). The assemblage is correlated with the *Alisporites plicatus - Jugasporites compactus* Concurrent Range Zone (Kungurian? to Roadian).

The Kim Fjelde Formation is overlain by the Midnatfjeld Formation, which consists of shale and carbonate in eastern Kim Fjelde and shale and sandstone in northern Kim Fjelde, dated from small foraminifers as late Kungurian to Kazanian (Stemmerik *et al.*, 1996). It contains assemblages that are similar to the *Ahrensisporites thorsteinssonii - Scutasporites nanuki* Concurrent Range Zone of the Sverdrup Basin. Diagnostic taxa include *Ahrensisporites thorsteinssonii* Utting, *A. multifloridus* Utting, *Hamiapollenites erebi*, *Piceapollenites nookapii* Utting, *Striatoabieites borealis* Utting, *Scutasporites nanuki*, and *Scutasporites unicus* Klaus. Acritarchs of the genus *Micrhystridium* are common, and *Veryhachium* spp. and *Unellium* spp. are present (Stemmerik *et al.*, 1996; this paper; Fig. 6^{9-10}). The assemblage contains relatively few pollen and spores, and it is not possible to determine whether the climate was humid or dry. However, the macro-flora from a nearby locality suggests a warm temperate humid climate (Wagner *et al.*, 1982; Wagner *et al.*, in press). This contrasts with the dry, cold climate proposed by Beauchamp (1994) for rocks of similar age in the Sverdrup basin. However, it is possible that the adjacent sea water was cooled by circulation of ocean currents, whereas the land was warm.

A few productive samples investigated in the present study from the fine to coarse grained sandstone of the overlying Parish Bjerg Formation, contain mainly taeniate disaccate pollen and monosulcate pollen. Species include *Ephedripites steevesiae* and *Lunatisporites novi-*

aulensis (Leschik) Foster, suggesting correlation with the Griesbachian *Tympanicysta stoschiana* - *Striatoabieites richteri* Assemblage Zone (Figs 4 and 6[5]).

EAST GREENLAND

Assemblages in the Karstryggen Formation (marine carbonate and gypsum) are similar to those of the Zechstein (Utting & Piasecki, 1995). They contain *Klausipollenites schaubergeri*, early forms of *Lueckisporites virrkiae*, *L. tatooensis* Jansonius, *Jugasporites delasaucei* (Potonié and Klaus) Clarke, *J. paradelasaucei* Klaus, *J. lueckoides* Klaus, *Lunatisporites noviaulensis*, *Perisaccus granulosus* (Leschik) Clarke, *Protohaploxypinus samoilovichii*, *Vittatina costabilis* Wilson and *Weylandites striatus* (Fig. 6[7]). Unlike the Zechstein they contain common and diverse *Vittatina* spp. The abundance of taeniate and polyplicate grains in rocks containing evaporite deposits suggests a hot dry climate for the Karstryggen.

The overlying Wegener Halvø and the Ravnefjeld formations contain the *Vittatina* Assemblage (Fig. 5[3]) of Balme (1980a) and Piasecki (1983). It is the upper Ravnefjeld Formation where the conodont *Neogondolella rosenkrantzi* has been recorded. This suggests an age from Wordian to early Wuchiapingian according to Kozur & Mostler (1995). Present are *Striatoabieites richteri*, *Lunatisporites noviaulensis*, *Protohaploxypinus samoilovichii*, *Lueckisporites virrkiae*, *Scutasporites nanuki* (= *Scutasporites* sp. cf. *S. unicus*) and *Weylandites* sp. Abundant is *Inaperturopollenites nebulosus* (Fig. 6[8,9]). *Scutasporites nanuki* has its earliest occurrence in the *Ahrensisporites thorsteinssonii* - *Scutasporites nanuki* Concurrent Range Zone, but the assemblage also resembles the *Tympanicysta stoschiana* - *Striatoabieites richteri* Assemblage Zone in that it contains *Striatoabieites richteri*, *Lunatisporites noviaulensis* and *Protohaploxypinus samoilovichii*. The presence of rare *Lueckisporites virrkiae* indicates some similarities with the Zechstein of western Europe (Visscher, 1971).

In the overlying Oksedal Member (Fig. 5[4]) of the Schuchert Dal Formation of Jameson Land the ammonoid *Paramexicoceras* occurs approximately 4 m below the lowest fish horizon of the Lower Triassic Wordie Creek Formation (Perch-Nielsen *et al.*, 1972; Nassichuk, 1995; and pers. comm. 2000). In the Kap Stosch region, stratigraphic relationship between ammonoids recorded from the traditional lithological unit, the "Martinia Limestone", and the units Oksedal member and the Ravnefjeld Formation is not clear. All units are dark mudstone and locally there is a potential for stratigraphic errors due to poor exposure and significant faulting (Piasecki pers. comm. 2000). However, it appears that *Paramexicoceras* is associated with the Dzhulfian (Wuchiapingian) *Cyclolobus* (Nassichuk, 1995). In the Oksedal Member (Fig. 5[4]) there is an abundance of taeniate and non-taeniate disaccates and *Vittatina* is less common than in underlying beds (Utting & Piasecki, 1995). *Tympanicysta stoschiana* and acritarchs are rare. Trilete spores are rare, but become increasingly common in the uppermost few metres of the member (Fig. 6[7]).

Overlying beds of the Wordie Creek Formation contain *Otoceras woodwardi boreale* and thus are of Griesbachian age. They contain the *Protohaploxypinus* Zone of Balme (1980a) which resembles the *Tympanicysta stoschiana* - *Striatoabieites richteri* - Assemblage Zone (Fig. 6[8]). Diagnostic taxa include *Proprisporites pocockii*, *Tympanicysta stoschiana*, *Striatoabieites richteri*, *Ephedripites steevesiae*, *Protohaploxypinus samoilovichii*, and *Densoisporites playfordii* (Balme) Dettmann. *L. virrkiae* occurs rarely. Overlying this zone is the *Taeniaesporites* Association also of Griesbachian age (Balme, 1980a) containing *Lundbladispora obsoleta* Balme, *Cycadopites follicularis* Wilson and Webster, *Ephedripites* spp., *Klausipollenites staplinii*, *Tympanicysta stoschiana*, *Lunatisporites noviaulensis*, *Densoisporites nejburgii* (Schultz) Balme, and *Endosporites* spp. (Fig. 6[8]).

SPITSBERGEN

Mangerud & Konieczny (1991, 1993) described palynomorph assemblages from a number of formations including the Kapp Starostin Formation (Fig. 6[10,11]). The lower to middle parts of the formation consist of shale, sandstone, carbonate with cherty intercalations, and the upper part shale with chert nodules and carbonate intercalations. The formation contains conodonts, brachiopods and corals (Nakamura *et al.*, 1992, and Fedorowkski & Bamber, in press). Nakamura *et al.*, (1992) suggested a Roadian to early Dzhulfian age based on brachiopods, although Fedorowski & Bamber (in press) pointed out that there were no biostratigraphic data to indicate an age younger than early Capitanian. The lower to middle parts of the formation contain the *Kraeuselisporites* assemblage, but the upper part lacks identifiable palynomorphs. The lower part of the *Kraeuselisporites* assemblage was correlated by Mangerud & Konieczny (1991,1993) with the *Alisporites insignis* - *Triadispora* sp. Assemblage Zone of Utting (1989), which is equivalent to the *Alisporites plicatus* - *Jugasporites compactus* Concurrent Range Zone of Utting (1994a), and the upper part of the *Kraeuselisporites* assemblage was correlated with the

Taeniaesporites sp. Assemblage Zone of Utting (1989), which is equivalent to the *Ahrensisporites thorsteinssonii - Scutasporites nanuki* Concurrent Range Zone of Utting (1994a). Significant is the presence of *Lueckisporites virrkiae* (Fig. 6) in the upper part of the *Kraeuselisporites* assemblage zone suggesting a Kazanian or younger age based on the Volga/Urals stratotype area (Utting *et al.*, 1997).

The overlying Deltadalen Member (greenish grey silty shale and minor sandstone) of the Vardebukta Formation contains the *Densoisporites nejburgii* Zone of Griesbachian age. Significant taxa include *Proprisporites pocockii*, *Striatoabieites richteri Ephedripites steevesiae*, *Lunatisporites noviaulensis*, *Protohaploxypinus samoilovichii*, *Uvaesporites imperialis*, *Densoisporites nejburgii*, *Maculatasporites* sp. and *Tympanicysta stoschiana* (Hochuli *et al.*, 1989; Mørk *et al.*, 1990, 1992, 1999). This assemblage (Fig. 6[12-15]) closely resembles the *Tympanicysta stoschiana - Striatoabieites richteri* Assemblage Zone of the Sverdrup Basin, the top of which has yet to be defined. However, in Spitsbergen the equivalent zone, Assemblage P (?Early Griesbachian), is defined by the range and common occurrence of *T. stoschiana* (= *C. chalasta*). In other respects it is difficult to differentiate from Assemblage O (Dienerian) (Hochuli *et al.*, 1989; Mørk *et al.*, 1992).

SVALIS DOME AND FINNMARK PLATFORM

Mangerud (1994) and Nilsson *et al.* (1996) correlated subsurface rock units of the Svalis Dome and Finnmark with those of the Svalbard Archipelago. In the lower Tempelfjorden Group of Svalis Dome they recognised the *Kraeuselisporites* Assemblage and in the upper part of the group the *Scutasporites* sp. cf. *S. unicus - Lunatisporites* spp. Assemblage Zone (Fig. 6[16]). In the Havert Formation, which consists of laminated and slightly bioturbated greenish grey shale, they recorded the *Pechorosporites* assemblage, which on the basis of a high percentage of *Aratrisporites* spp. pollen appears to be slightly younger than the *Tympanicysta stoschiana - Striatoabieites richteri* Assemblage Zone (Fig. 6[17]). A late Griesbachian age is indicated from the marine fauna (Vigran *et al.*, 1998).

In Finnmark Mangerud (1994) recognised three zones (Fig. 6[18]) and correlated them with those established for the Sverdrup Basin (Utting, 1994a). These were:

i) *Dyupetalum* sp. (*Dyupetulam vesicatum* Utting, 1994 a) - *Hamiapollenites bullaeformis* Assemblage Zone (Kungurian? - Ufimian), equivalent to the *Alisporites plicatus - Jugasporites compactus* Concurrent Range Zone (Utting, 1994 a).

ii) *Scutasporites* sp. (*Scutasporites nanuki* Utting 1994 a) - *Lunatisporites* spp. Assemblage Zone (Kazanian - Tatarian?) equivalent to the *Ahrensisporites thorsteinssonii - Scutasporites nanuki* Concurrent Range Zone (Utting, 1994 a).

iii) *Lundbladispora obsoleta - Tympanicysta stoschiana* Assemblage Zone (early Griesbachian) equivalent to the *Tympanicysta stoschiana - Striatoabieites richteri* Assemblage Zone.

The *Dyupetalum* sp. - *Hamiapollenites bullaeformis* Assemblage Zone was defined by the earliest occurrence of *Dyupetalum* sp. (= *D. vesicatum* of Utting, 1994 a), and the latest occurrence of *Hamiapollenites bullaeformis* (Samoilovich) Jansonius. In the Sverdrup Basin *D. vesicatum* has its earliest occurrence in the *Alisporites plicatus - Jugasporites compactus* Concurrent Range Zone, but *H. bullaeformis* extends into the *Ahrensisporites thorsteinssonii - Scutasporites nanuki* Concurrent Range Zone (Fig. 3). The base of the *Scutasporites* sp. cf. *S. unicus - Lunatisporites* sp. Assemblage Zone was defined at the earliest occurrence of *Scutasporites* sp. cf. *S. unicus* (= *Scutasporites nanuki*) of Wordian age, and the *Lundbladispora obsoleta - Tympanicysta stoschiana* Assemblage Zone is defined by the presence of the *Tympanicysta stoschiana*, *Striatoabieites richteri*, *Proprisporites pocockii*, and the *Uvaesporites imperialis* morphon. There is a similar abundance of *Uvaesporites imperialis* in the lower part of the Havert Formation to that found in the basal Blind Fiord assemblages. Interesting is the presence in the early Griesbachian of poorly preserved specimens of *Vittatina* and well preserved specimens of *Lueckisporites virrkiae*. Mangerud (1994) considered the former to be reworked but the latter, which are rare to common, to be *in situ*. As discussed above in the section concerning Alaska, it is difficult to resolve this problem of *in situ* versus reworked taxa. In view of the hiatus present between the Wordian and Griesbachian in Finnmark, *L. virrkiae* could have been derived from post-Wordian rocks that have susequently been eroded, on the other hand its rare, but consistent presence in circum-polar Griesbachian rocks could indicate that the parent plant was not extinct.

KOLGUYEV ISLAND

Ufimian and Kazanian assemblages from Kolguyev Island and the Kungurian? to Roadian and Wordian of the Sverdrup Basin have many stratigraphically diagnostic taxa in common and correlation of the spore zones was proposed by Grigoriev & Utting (1998). For example the Kazanian (Wordian) Assemblage II of Kolguyev includes *Ahrensisporites thorsteinssonii*, *A. multifloridus*, *Scutasporites nanuki*, and *Hamiapollenites* *erebi* (Fig. 6[19]). Nevertheless there are differences: *Lueckisporites* sp. (a small representative of the *L. virkiae* morphon) is present on Kolguyev, but absent in the Sverdrup Basin. Also present on Kolguyev, but absent from the Sverdrup basin in rocks of Kungurian? to Wordian age, are taxa such as *Limitisporites monstruosus* (Luber and Waltz) Hart and *Crucisaccites ornatus* (Samoilovich) Dibner. In addition Kolguyev assemblages differ in that they are dominated by pteridophyte spores and this suggests that the climate may well have been more humid than that of the Sverdrup Basin. Griesbachian rocks are probably absent from the island (Hochuli *et al.*, 1989; Mørk *et al.*, 1992).

MID-NORWAY

In this area data concerning the Kungurian? to Wordian microfloras are sparse (Fig. 6[20]), but more details are available concerning the overlying late Griesbachian to Smithian rocks (Vigran & Mangerud, 1991). The latter are dominated by trilete cavate spores including *Uvaesporites imperialis*. The fact that *Aratrisporites* spp. were common was taken to indicate an age younger than early Griesbachian.

VOLGA/URALS REGION

When comparisons are made between the Roadian and Wordian assemblages of the Sverdrup Basin with those from the Ufimian and Kazanian stratotype areas of the Volga-Urals region, Russia, it is evident that there are major differences in composition (Utting *et al.*, 1997; and Fig. 6[21]). For example the Sverdrup basin contains abundant trilete spores in the Roadian and Wordian, but although these are common in the type Ufimian, they are rare in the Kazanian. Many taxa found in the Wordian of the Canadian Arctic are absent from the Kazanian (Utting *et al.*, 1997). For example taxa that are present in the Wordian, but lacking in the Kazanian of Russia, include stratigraphically diagnostic taxa such as *Ahrensisporites multifloridus*, *A. thorsteinssonii*, *Lunatisporites beauchampii*, *Piceapollenites nookapii* Utting, *Scutasporites nanuki*, *Striatoabieites borealis* Utting, *Diatomozonotriletes hypenetes* Utting, *Lunatisporites arluki* Utting, *Grandispora jansonii* Utting and *Inaperturopollenites nebulosus*. The only taxon present in both the Wordian and Kazanian assemblages is *Hamiapollenites erebi*. Taxa restricted to the Ufimian and Kazanian are *Hamiapollenites tractiferinus*, *Limitisporites monstruosus*, *Crucisaccites ornatus*, *Cordaitina subrotata* var. *isopolaris* (Varyukhina) Varyukhina, *Cordaitina uralensis* (Luber) Samoilovich, *Kraeuselisporites papulatus* Virbitskas and, *Entylissa caperata* (Luber) Varyukhina. Those restricted to the Kazanian, and absent from the Wordian, are *Lueckisporites virrkiae* and *Discernisporites* sp.

If, as suggested by the marine faunas the Wordian and Kazanian are correlative (see summary by Fedorowski & Bamber, in press), then the fundamental differences in the palynomorph assemblages between the Volga/Urals area, and the Sverdrup Basin and other circum-polar areas may have been caused by a variety of factors independent of time including paleoclimate, facies, latitude, topography, and relative position on Pangea. For example, the climate was probably hot and arid in the Volga/Urals region from the Ufimian to the Tatarian (Utting *et al.*, 1997), whereas in the Sverdrup Basin it was humid and cool in the Kungurian? to Roadian, and cool and dry in the Wordian (Beauchamp, 1994; Utting, 1994a).

CONCLUSIONS

1. Assemblages of Kungurian? to Wordian, and Griesbachian age are similar throughout the circum-polar area studied.
2. With the exception of East Greenland, there are no palynological data available from the Wuchiapingian and Lopingian stages.
3. Middle Permian assemblages on Kolguyev Island have many features in common with other circum-polar localities, but there are also some taxa in common with the Volga/Urals region.
4. The circum-polar assemblages of Kungurian? to Roadian and Wordian age differ markedly from the Ufimian and Kazanian assemblages of the Russian stratotype areas in the Volga /Urals region, probably reflecting differences in a number of factors including climate, facies, latitude, elevation and location in Pangea.

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FORMATION AND EVOLUTION OF THE PERMIAN STRATA OF THE EASTERN TIANSHAN MOUNTAIN IN XINJIANG, CHINA

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Key words – Xinjiang; eastern Tianshan Mountain; Permian; molasse facies; homologous flysch facies; uniform intracontinental lake basin.

Abstract – Owing to the inhomogeneity of orogeny with the compression force chiefly from the southern flank, the uplift folding (orogeny) formed since the Late Carboniferous at the southern and northern margins of the eastern Tianshan Mountain in Xinjiang, China is asymmetrical. In the early to Middle Permian, a set of piedmont molasse facies sedimentary associations were developed at the southern margin of the eastern Tianshan Mountain, near the core of the orogeny; at the northern margin there existed a set of assemblages of homologous flysch facies formed in a foreland basin. By the end of the Middle Permian, the difference between the basement topographies of the southern and northern margins disappeared gradually, and the eastern Tianshan Mountain and its periphery area thus became a broad, uniform, intracontinental lake basin. The depositionally continuous terrestrial sediments of the Permian to Triassic boundary are found broadly in these areas.

Parole chiave – Xinjiang; Tianshan orientale; Permiano; facies molassica; facies flyshoidi omologhe; uniforme bacino lacustre intracontinentale.

Riassunto – A causa della disomogeneità dell'orogenesi derivante dalla forza di compressione esercitata principalmente dal fianco meridionale, il sollevamento per pieghe determinatosi a partire dal Carbonifero superiore sui margini meridionale e settentrionale del Tianshan orientale in Xinjiang, Cina, è asimmetrico. Durante il Permiano inferiore e medio, un gruppo di associazioni sedimentarie caratterizzate da facies molassiche pedemontane si sviluppò lungo il margine meridionale del Tianshan orientale, vicino al nucleo dell'orogenesi; sul margine settentrionale esisteva un gruppo di associazioni di facies omologhe di flysch formatesi in un bacino di avampaese. Entro la fine del Permiano medio, le differenze topografiche relative al basamento dei margini meridionale e settentrionale scomparvero gradualmente, e il Tianshan orientale e la sua periferia divennero pertanto un esteso, uniforme, bacino lacustre intracontinentale. I sedimenti terrestri deposizionalmente continui al limite P/T sono ampiamente presenti in queste aree.

INTRODUCTION

Xinjiang Uygur Autonomous Region (1.6 million km^2 in area) is the largest provincial region in China. The Tianshan mountain range lies across the central part of Xinjiang. Habitually, taking the highway from Urumqi City, the provincial capital of Xinjiang, to Toksun county as a boundary, to the west of this boundary is called the western Tianshan Mountain, and to the east is the eastern Tianshan Mountain which chiefly comprises Mount Bogda and Mount Karlik. On the southern margin of the eastern Tianshan Mountain is the Turpan-Hami Basin and on the northern margin is the southeastern part of the Junggar Basin.

Permian rocks are widespread throughout Xinjiang, especially in the north of Xinjiang where they are best developed. The sedimentation types are varied and fossils are abundant in the Early, Middle and Late Permian strata of the eastern Tianshan Mountain studied in this paper. There are many complete and continuous stratigraphic sections (Fig. 1) in this region, which has gradually become the "hot spot" for studies of the continental Permian strata and the continental Permian- Triassic boundary beds in recent years in China (Cheng Zhengwu *et al.*, 1997; Zhou Tongshun *et al.*, 1997; Sheng Jinzhang & Jin Yugan, 1994).

TECTONIC SETTING OF THE FORMATION OF THE EARLY AND MIDDLE PERMIAN STRATA OF EASTERN TIANSHAN MOUNTAIN

The middle and late Hercynian Orogeny, and the two plate suture belts which between them pinned the eastern Tianshan Mountain, were the chiefly controlling factors affecting the formation, development, and distribution of the

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Lower and Middle Permian Series. The Karamaili suture belt on the northeastern side of eastern Tianshan Mountain was the convergent margin between the Siberian plate and the Kazakhstan - Junggar plate, and finally closed in the Middle-Late Devonian (Li Jinyi *et al.*, 1989, 1990), but did not result in strong compressive folding and metamorphism. It resulted only in formation and uplift of transitional crust (Xiao Xuchang *et al.*, 1992) over a vast area of the southeastern part of the Junggar Basin and the continuous shrink of the residual Carboniferous sea basin, which changed to give near-shore limnetic facies sediments during the Early Permian. Another suture belf, the Erenhabirga-Kangguer suture belt on the southwestern side of eastern Tianshan Mountain, was the convergent junction of the Tarim plate and the Kazakhstan-Junggar plate. It produced sharp stratigraphic folds and volcanic activity during the Late Carboniferous, with typical piedmont molasse facies at the southern margin of eastern Tianshan Mountain, under the influence of strong and sustained compression from the south during the Early and Middle Permian. At the northern margin of eastern Tianshan Mountain, a paralic transitional facies developed consisting of very thick coarse to fine clastic sediments of foreland basin-type.

THE FORMATION AND GENERAL FEATURES OF THE EARLY AND MIDDLE PERMIAN MOLASSE FACIES AT THE SOUTHERN MARGIN OF EASTERN TIANSHAN MOUNTAIN

During uppermost Late Carboniferous times, with the loss of balance of the originally roughly uniform tectonic-sedimentary framework of the Bogda-Karlik volcanic arc (Li Jiliang, 1989), the nuclear zone of the orogenic belt close by the southern margin of eastern Tianshan Mountain rapidly became folded and uplifted. Seawater withdrew rapidly towards the east and northeast, and the Carboniferous volcanics and pyroclastic rocks as well as carbonate rocks higher than the erosion surface were sharply denuded. In assocation with some Early Permian volcanic eruptive rocks, this debris was rapidly transported from the piedmont belt to the nearshore depression and deposited there along a belt from Ewirgol through Taoxigou and Yiwanquan to Kulai. This sequence of red molasse sediment is characterised by the following:

1. It has a narrow ribbon-shaped distribution in the southern margin area of eastern Tianshan Mountain, about 600 km long and only a few tens of km in width. The parent rocks of this suite of coarse clastic sediments are composed chiefly of the underlying disintegrated Late Carboniferous volcanics and pyroclastic rocks, as well as minor carbonate rocks containing Late Carboniferous fossils. This is a suite of typically continental piedmont molasse (Liao Zhuoting *et al.*, 1999), as a whole dominated by river fan and alluvial fan deposits, with only occasional limnic sediments in the upper part.
2. This suite of continental molasse sediments displays the normal cyclic sequence, coarse below and fine above, which may be looked upon in general as a complete magnacyclothem The interbeds of volcanics only occur in the middle and lower part, indicating the intensity of the Early and Middle Permian tectonic activity and the sedimentation rate changing from strong to weak.

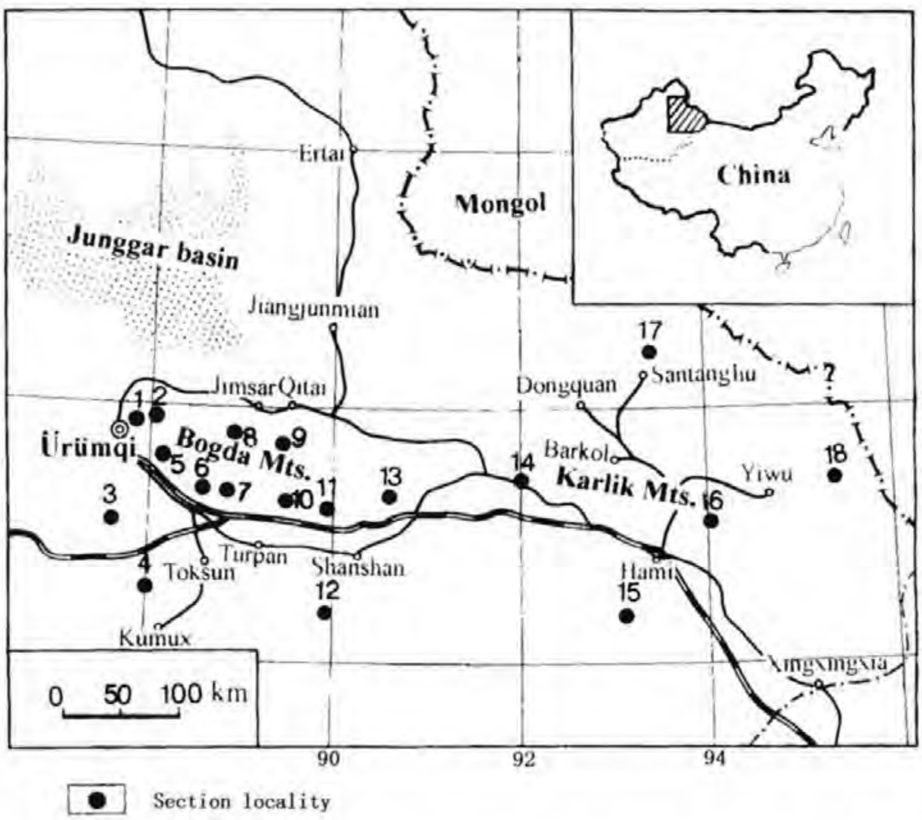

Fig. 1 – Map showing the section localities of Permian strata in eastern Tianshan Mountain and adjacent regions, Xinjiang.

1. Jingjingzigou; 2. Lucaogou; 3. Ewirgol; 4. Mishigou; 5. Guodikeng; 6. Taoxigou; 7. Taotonggou; 8. Dalongkoul; 9. Quanzijie; 10. meiyaogou; 11. Ertanggou; 12. Dikanr; 13. Zhaobishan; 14. Yiwanquan; 15. Dananhu; 16. Kulai; 17. Santanghu; 18. Naomaohu.

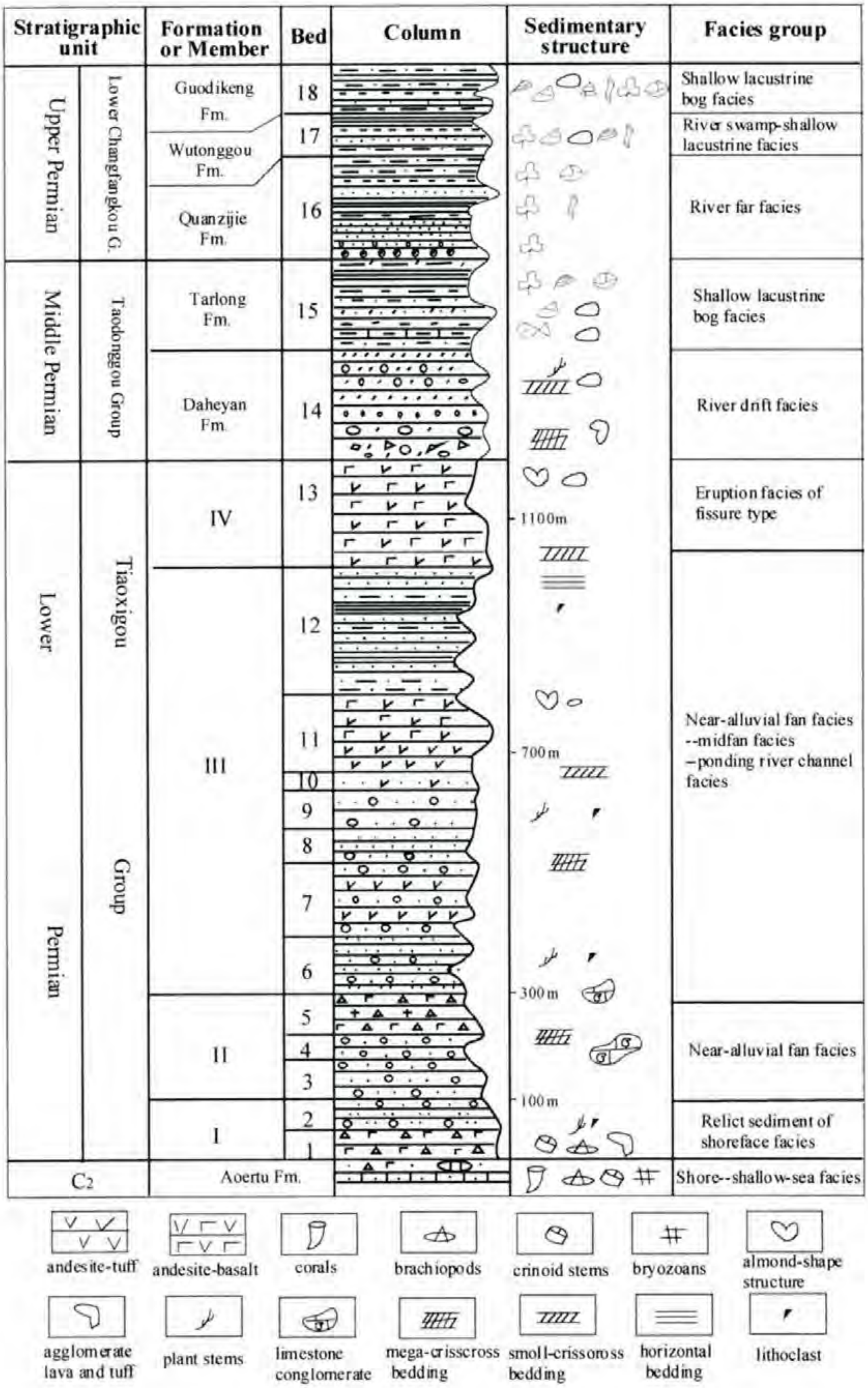

Fig. 2 – Columnar section of the Permian strata on the southern margin of the eastern Tianshan Mountain, Xinjiang, China.

3. There are neither flysch facies sediments below this suite of molasse, nor transitional assemblages intermediate between these two types of sedimentary associations (Li Jiliang, 1978). This occurrence suggests that an abrupt change in the change of the volcanic arc facies from a marginal island arc sedimentary environment into a continental sedimentary environment marks the occurrence of the "bald head" in the orogenic belt of eastern Tianshan

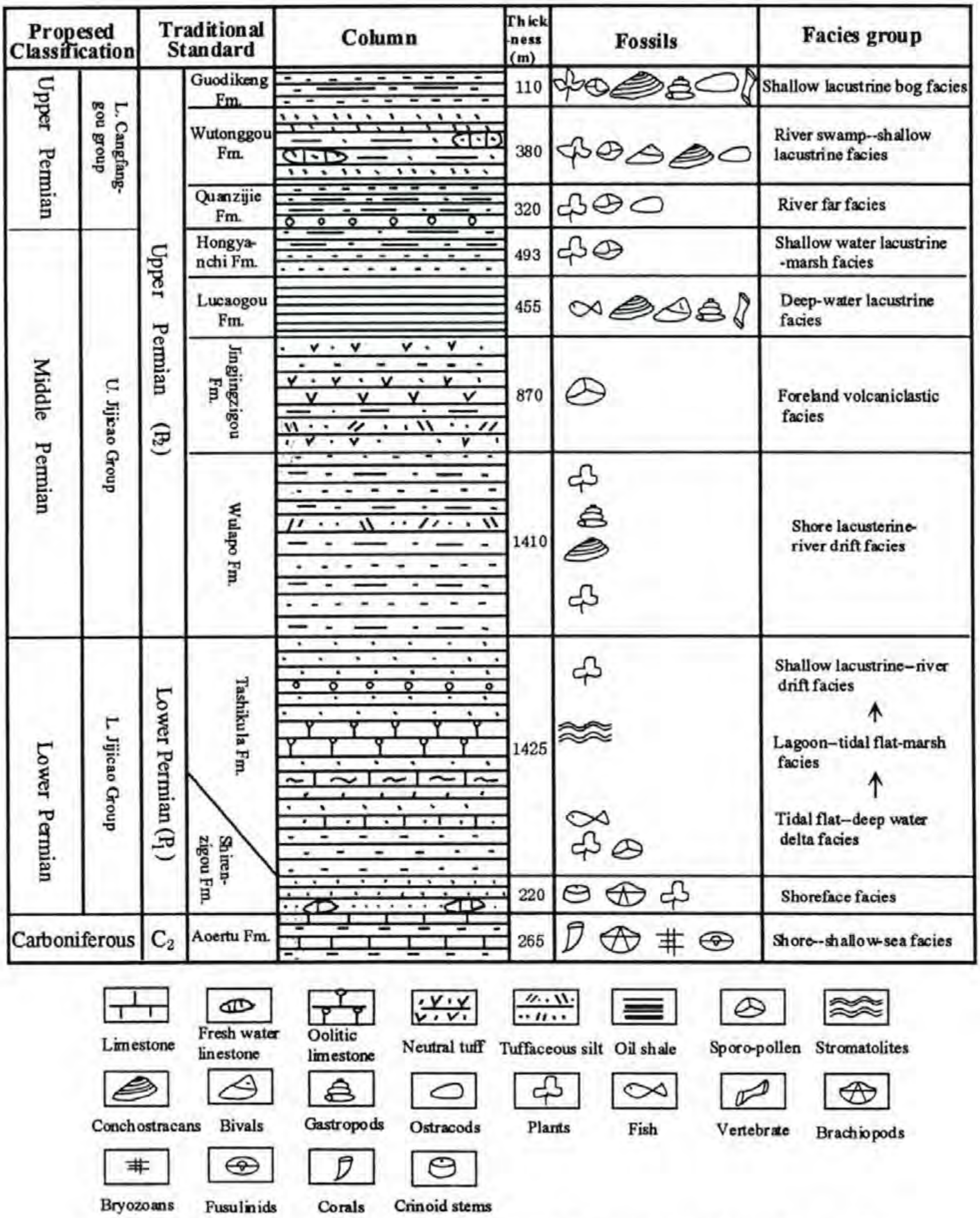

Fig. 3 – Columnar section of the Permian strata on the northern margin of the eastern Tianshan Mountain, Xinjiang, China.

Mountain, and the end of the Hercynian Orogeny. Evidently, the beginning and end orogenic movement is slightly earlier at the southern margin of the eastern Tianshan Mountain than at the northern margin (Fig. 2).

THE FORMATION AND MAIN FEATURES OF THE EARLY AND MIDDLE PERMIAN FORELAND BASIN FACIES AT THE NORTHERN MARGIN OF EASTERN TIANSHAN MOUNTAIN.

Owing to the inhomogeneity of the orogenesis and the influence of compression from the south, the northern margin of eastern Tianshan mountain was resisted by the rigid transitional-type crust at the southeastern part of the Junggar Basin, giving rise to a sharply downwarping and eastward- opening foreland basin. Although the seawater in this region began to withdraw eastward like that in the southern margin during the very beginning of the Early Permian, the rate of its withdrawal to the east was not as rapid as at the southern margin and the period of its complete withdrawal was also not as short, exhibiting a slow and continuous process. Therefore a flysch facies suite consisting chiefly of sandy and argillaceous sediments, in the basal part of which are intercalated slumps composed of shallow-water marine carbonate rocks (Liao Zhuoting *et al.*, 1987). Upwards there are transitional sandstone beds containing large numbers of plant fossils, ooid-bearing dolomite, and interbeds of stromatolites, and upwards again there are medium-and fine-grained clastic rooks containing coal streaks and thin coal seams. The top part is similar to the Middle Permian of the southern margin, in the dominance of limnetic oil shales, dolomite, argillaceous limestone and mudstone. In this Middle and Lower Permian Series, the sediment was not found to be derived from the underlying, reworked, Upper Carboniferous fossil-bearing carbonate rocks like those in the southern margin, and the volcanics and interbedded volcanic sedimentary rocks are not so well-developed as those in the southern margin.

The nature of the above-mentioned transitional sedimentary assemblage indicates that the volcanic arc of eastern Tianshan Mountain changed from a strongly subsiding phase to the late stage of orogenic uplift, as the regional tectono-sedimentary environment at the northern margin experienced a continuously changing process from marine through to transitional and continental fasies during the Early and Middle Permian. This suite of foreland basin transitional sediments is quite different from the contemporary piedmont molasse facies of sedimentary association at the southern margin, in terms of its genesis, lithologies, sedimentation type and spatial extent (Fig. 3).

FORMATION OF THE LATE PERMIAN UNIFORM INTERIOR BASIN FACIES IN EASTERN TIANSHAN MOUNTAIN

Above paragraph, we have expounded the changes in the Early and Middle Permian tectono-sedimentary environment, and the features of stratigraphic development at the southern and northern margins. As a whole, during the Early Permian, the two regions were distinctly different, until the Middle Permian when the differences between them reduced, and the sedimentary succession and texture gradually came to be identical, especially in the uppermost horizons where the extrusive volcanics and volcaniclastic rocks that occur are restricted to the Upper Jijicao Group, indicating that the regional volcanic activity ended in the Middle Permian (Liao Zhuoting & Wu Guogan, 1998). The overlying Cangfanggou Group is a suite of time-transitional (from Late Permian to Early Triassic), depositionally continuous, widespread continental clastic rocks. The regional unconformity between them and the underlying Upper Jijicao Group represents the end of the Hercynian Orogeny and the beginning of the non-orogenic, within-plate developmental stage. From then on, the difference between the southern and northern margins of the eastern Tianshan Mountain orogenic belt, in structure, landform, relief, biological assemblage, and entire sedimentary environmental differences resulting from the originally inhomogeneous orogeny, began to diappear, and a vast inland lake basin occupied the whole eastern Tianshan Mountain (Zhou Tongshun *et al.*, 1997). This new and uniform tectonic paleogeographical framework lasted from the Late Permian to the Early Triassic. Thus, judging from the history of tectonic development and the sedimentary features, it is reasonable to divide the eastern Tianshan Permian into three series: the Lower, Middle and Upper Series.

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EARLY PERMIAN TETRAPOD TRACKS - PRESERVATION, TAXONOMY, AND EURAMERICAN DISTRIBUTION

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Key words – Tetrapod footprints; Lower Permian; red beds; taxonomy; stratigraphy.

Abstract – After the first comparative studies, most of the material known from the European Permian red beds corresponds to the track fauna of the Abo Formation of Wolfcampian age in North America. All differing stratigraphic interpretations of European track sites in red bed facies come under question. The spectrum of ichnotaxa in red beds of Lower Permian age is restricted to *Batrachichnus*, *Limnopus*, *Amphisauropus*, *Hyloidichnus*, *Ichniotherium*, *Dromopus* (*D. lacertoides* = *D. agilis* = *D. didactylus*), *Tambachichnium*, *Dimetropus* and *Gilmoreichnus*. These ichnotaxa can be correlated with temnospondyls, seymouriamorphs, diadectids, araeoscelids, and pelycosaurs, taxa characteristic of the Lower Permian. The vast majority of occurrences of tetrapod tracks in red bed facies of the Permian show no significant differences when examined using the taxonomic criteria accepted by the authors. The ichnofauna of the Abo, and related formations of Wolfcampian age, are consequently, a potential standard for both ichnofaunal and stratigraphic purposes. The distribution in time of most Euramerican Permian red beds with tetrapod tracks, inclusive of the so-called Upper Rotliegend, is confined to the lowermost Permian, correlating to the Asselian, Sakmarian, and Artinskian.

Parole chiave – Impronte di tetrapodi; Permiano inferiore; *red beds*; tassonomia; stratigrafia.

Riassunto – Dopo i primi studi comparativi, la maggior parte del materiale conosciuto nei *red beds* del Permiano europeo corrisponde alla fauna ad impronte della Formazione di Abo del Wolfcampiano nord-americano. Tutte le varie interpretazioni stratigrafiche di siti di impronte europee in facies di *red beds* sono poste in discussione. Lo spettro di icnotaxa nei *red beds* del Permiano inferiore è ristretto a *Batrachichnus*, *Limnopus*, *Amphisauropus*, *Hyloidichnus*, *Ichniotherium*, *Dromopus* (*D. lacertoides* = *D. agilis* = *D. didactylus*), *Tambachichnium*, *Dimetropus* and *Gilmoreichnus*. Questi icnotaxa possono essere correlati con temnospondili, seymuriamorfi, diadectidi, areoscelidi e pelicosauri, taxa caratteristici del Permiano inferiore. L'ampia maggioranza di eventi di tracce di tetrapodi nei *red beds* permiani non mostra tra esse alcuna significativa differenza allorché sono prese in esame usando i criteri tassonomici accettati dagli autori. L'icnofauna dell'Abo, e le connesse formazioni di età wolfcampiana, sono conseguentemente un potenziale standard per scopi icnofaunistici e stratigrafici. La distribuzione cronologica della maggior parte di *red beds* permiani euroamericani con tracce di tetrapodi, comprensive del così chiamato Rotliegende superiore, è confinata al Permiano più basso, correlabile all'Asseliano, Sakmariano e Artinskiano.

INTRODUCTION

About 140 ichnogenera related to Permian tetrapods have hitherto been named in the literature. This high number of taxa could be understood as evidence of 1) high faunal diversity , and 2) a possible basis for refined stratigraphical subdivision of the track-bearing formations. However, most such results are derived from regional studies and relatively few local samples. The trigger for a critical revision was the discovery by J. P. MacDonald of extensive tracksites in the "Abo Tongue of the Hueco Formation" (Robledo Mountains Formation of Hueco Group) of southern New Mexico in the late 1980s. From a large sample size at a Lower Permian megatrack-site, many preservational variations along trackways can be demonstrated. Isolated segments of these trackways were formerly often understood as different ichnotaxa worthy of species- and genus-level recognition. The traditional interpretation of a high diversity in tetrapod ichnofauna is opposed by a low-diversity interpretation, both documented in the volume edited by Lucas & Heckert (1995). The characteristic tetrapod ichnogenera of the low-diversity interpretation of the Lower Permian Wolfcampian red beds in North America are *Batrachichnus*, *Limnopus*, *Hylodichnus*, *Dromopus*, *Gilmoreichnus*, and *Dimetropus* (Haubold *et al*., 1995

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a). In the continuation of this study, an attempt was made at a reorganisation of the taxonomy, classification and stratigraphical value of the tetrapod tracks of the Permian (Haubold, 1996). As one result, the cosmopolitan character of tetrapod ichnofaunas of Early Permian age became evident (Hunt & Lucas, 1998). These results and analyses were continued through 1998 and 1999 by further fieldwork in North America and Europe, with additional collections at track sites of red bed facies.

TRACKBEARING FORMATIONS OF WOLFCAMPIAN AGE

The collections from the Abo and Sangre de Cristo Formations of central and northern New Mexico were extended in 1999, in addition to the occurrences in coastal plain environments of the Robledo Mountains Formation of the Hueco Group (Lucas *et al.*, 1998) in southern New Mexico (Fig. 1). Track sites with large number of samples have been excavated and provide most of the obligate ichnotaxa in the broad scale of muddy to sandy red-bed facies. After comparative studies of Permian tetrapod ichnofaunas from Europe, most of the trackbearing Permian formations of red-bed facies correspond in age to those of the Hueco, Abo, and Sangre de Cristo Formations of Wolfcampian age in New Mexico. Note that the Wolfcampian Stage-Age is a provincial dating term used in North America to refer to much of Early Permian time; it is approximately equivalent to the Asselian, Sakhmarian and part of the Artinskian using the standard global chronostratigraphic scale.

Over the last 100 years, Permian ichnofaunas and their geological occurrences have been described by several authors. However, the determinations and many of the names listed by some authors are relatively incompatible taxonomically with each other. For a principal revision, all known track-bearing formations and the majority of specimens have been investigated and re-examined, particularly during the last decade. Hitherto published results of reinvestigations, revisions and new collections, concern tracks from the Hermit Shale (Haubold *et al.*, 1995 a), the Standenbühl/Nierstein Formation (Haubold & Stapf, 1998) and the Tambach Formation (Haubold, 1998).

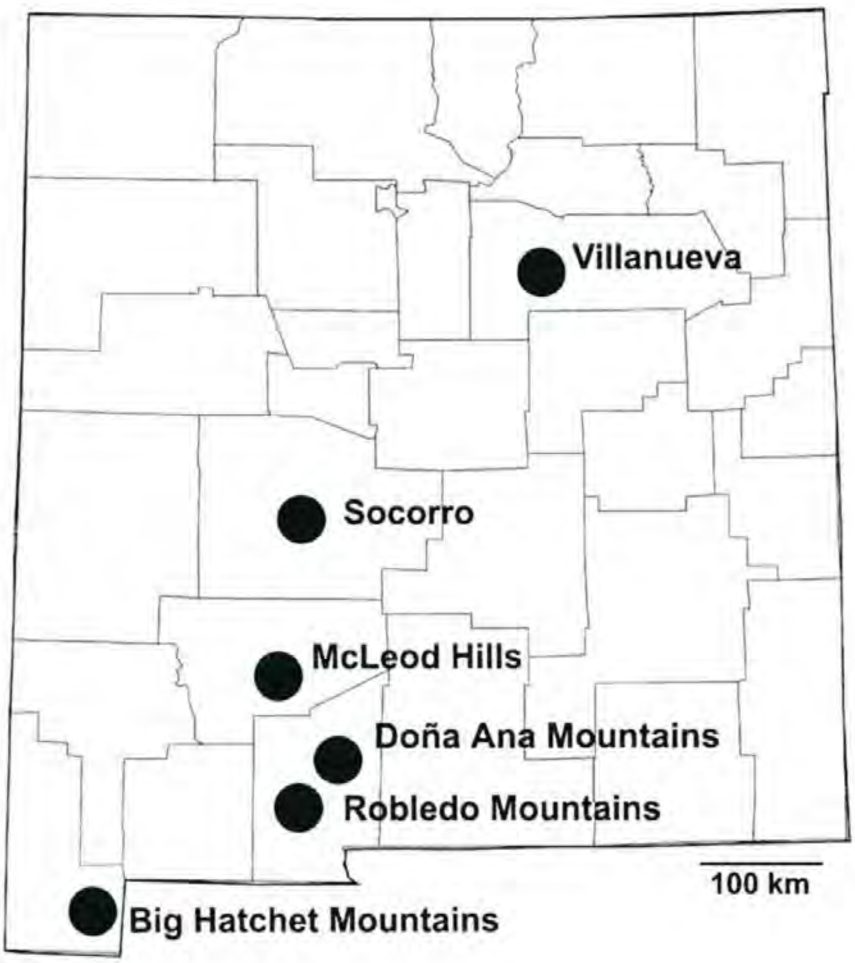

Fig. 1 – Geographical distribution of the Lower Permian track-sites of Wolfcampian age in New Mexico. The occurrences belong to, from south to north, the Robledo Mountain, the Abo and Sangre de Cristo formations.

In summary, the following trackbearing units correlate to the Wolfcampian tetrapod track faunas in the Hueco (Robledo Mountains), Abo and Sangre de Cristo Formations in New Mexico:

- USA – Arizona, and Colorado: Hermit Shale, and Cutler Formation
- Northern Italy – Collio Formation of the Southern Alps
- Southern France – Permian sequence of the Lodève Basin up to the Rabejac Formation; Permian sequences of the Luc, Bas Argens and Estérel Basins up to the Mitan and Gonfaron Formations; Permian sequence of the St. Affrique Basin up to the pelites of St. Pierre.
- Germany – Rotliegend sequence of the Saar-Nahe Basin up to the Nierstein/Standenbühl Formation of the Wadern Group; Hornburg Formation of the eastern Harz Foreland Basin; Rot-

liegend sequence of the Thüringer Wald up to the Tambach Formation.

Other track sites of Permian red beds with the same ichnofaunal elements are known in southern Poland, nothern Bohemia (Czechien), England, northern Spain, Nova Scotia, Argentina, and southern Russia.

Of separate age are the tetrapod ichnofaunas of the Val Gardena Sandstone in northern Italy, the Salagou Formation (site La Lieude) of the Lodève Basin in southern France, and the Choza Formation in Texas.

Different in facies and somewhat different in age are the principally aeolian track occurrences, the so-called *Chelichnus* ichnofacies (*Laoporus* ichnofacies of Lockley *et al.*, 1994) of the Supai Formation, Coconino Sandstone, DeChelly Sandstone, Corncockle and Locharbriggs Sandstones, Hopeman Sandstone, and Cornberger Sandstein (Haubold *et al.*, 1995 b; Lockley *et al.*, 1995; McKeever & Haubold, 1996; Morales & Haubold, 1995).

As a result, the Wolfcampian track fauna can be characterised by five to six basic track types, of about nine ichnogenera. This morphological subdivision is in agreement with the range of presumed perpetrators of the tracks, the osteologically-recorded Lower Permian temnospondyls, seymouriamorphs, diadectids, araeoscelids, captorhinomorphs, and pelycosaurs (Fig. 2). As currently understood, the record of tracks of captorhinomorphs and pelycosaurs is comparatively more diverse. Tracks of temnospondyls, araeoscelids, and seymouriamorphs are most frequent, whereas the record of diadectid tracks – *Ichniotherium* – appears restricted.

FORMER ATTEMPTS AT PERMIAN TETRAPOD ICHNOSTRATIGRAPHY

All former stratigraphic subdivisions using tetrapod tracks come into question as a result of the new data. Indeed, not one of the taxa used for the tetrapod track zonation by Boy & Fichter (1988) and Kozur (1989), such as *Saurichnites*, *Telichnus*, *Varanopus*, *Hardakichnium*, *Palmichnus*, *Anhomoiichnium* and *Actibates*, can be accepted as valid. In other words, the age of the formations in red bed (Rotliegend) facies – as reflected in the zonations proposed by these authors, as well as the associations I to III of Gand (1987), Gand & Haubold (1988), and the so-called Saxonian distinguished by tetrapod footprints by Haubold & Katzung (1972, 1975) – is restricted to the Wolfcampian.

These results – which conflict with several former opinions – are related to two basic tenets of biostratigraphy. Firstly, similar biological forms found as fossils are potential evidence of a similar age for related formations. Secondly, differences in recorded/observed biological forms are evidence of different ages.

If one uses these premises positivistically, just a handful of divergences may lead to a wide range of subjective interpretations. Every ichnological form of tetrapod track that looks different may be new and may be

Ichnotaxon	**anatomical interpretation**
***Batrachichnus* Woodworth, 1900** *B. salamandroides* (Geinitz, 1861) *B. delicatulus* (Lull, 1918) ***Limnopus* Marsh, 1894** *L. cutlerensis* Baird, 1965 *L. zeilleri* (Delage, 1912)	Temnospondyli
***Amphisauropus* Haubold, 1970** *A. latus* Haubold, 1970 *A. imminutus* Haubold, 1970	Seymouriamorpha
***Ichniotherium* Pohlig, 1895** *I. cottae* (Pohlig, 1885)	Diadectidae
***Dromopus* Marsh, 1894** *D. lacertoides* (Geinitz, 1861) ***Tambachichnium* Müller, 1954** *T. schmidti* Müller, 1954	Araeoscelida
***Hyloidichnus* Gilmore, 1927** *H. microdactylus* (Pabst, 1897) *H. bifurcatus* Gilmore, 1927 *H. major* (Hey. &. Less., 1963) ***Gilmoreichnus* Haubold, 1970** *G. hermitensis* (Gilmore, 1927) ***Dimetropus* Romer & Price, 1941** *D. nicolasi* Gand & Haubold, 1984	Captorhinomorpha /Pelycosauria

Fig. 2 – The Lower Permian – Wolfcampian – principal tetrapod ichnotaxa of Euramerican distribution, and their interpretation.

Acibates Acrodactylichnium Acutipes Agostopus Akropus Allopus Amblyopus Amphisauroides Amphisauropus Anhomoiichnium Anthichnium Anthracopus Attenosaurus Auxipes "Antianhomopus" "Archaeotheriopus" "Arktopus" "Asteipodiscus"
Baropezia Baropus Barypodus Batrichnis Batrachichnus Brontopus
*Camunipes Cardiodactylum Chelaspodus Chelichnus *Chirotherium Cincosaurus Collettosaurus Crenipes Cursipes "Caseipus" "Cyclopus"*
*Devipes *Dicynodontipus Dimetropus Diversipes Dolichopodus Dromillopus Dromopus "Dromicopezus"*
Erpetopus Eumekichnium "Edaphopus" "Eobatrachidopus" "Eobarachipodiscus" "Eocercopithecopus" "Eocynodontipus" "Eodicynodontipus" "Eomacrochlichnus" "Eootokosaurus" "Eotheriopodiscus" "Eulaoporoides" "Eulaoporus" "Exopodiscus"
Fichterichnus Foliipes
Gampsodactylum Gargalonipes Gilmoreichnus Gonfaronipes Gracilichnium "Gnamptonycopus"
Hardakichnium Harpagichnus Herpetichnus Hyloidichnus Hylopus
Ichniotherium Ichnium "Isolaoporus"
Jacobiichnus Janusichnium
Korynichnium "Keraunopus"
Labyrinthodon Laoporus Limnopus Luneapes Loxodactylus
"Laodromus" "Laopezus" "Laoporoides" "Leptotheriopodiscus"
Margennipes Merifontichnus Microsauropus Moodieichnus
"Macrochelichnus" "Megalaoporus" "Micropodiscus" "Moschopopus"
Nanopus Nanipes Notalacerta
Oklahomaichnus Onychichnium Opistopus "Okypus"
Pachypes Palaeopus Palmichnus Parabaropus Paradoxichnium Permomegatherium Phalangichnus Planipes Prochirotherium Procolophonichnium Protritonichnites Pseudobradypus Pseudosynaptichnium "Paranodontipus" "Plagionycopus" "Praefoliipes" "Pseudoanthropopus" "Pseudopithecopus" "Psililaoporus"
Quadropedia
**Rhynchosauroides*
*Salichnium Saurichnis Saurichnites Serripes Sphaerodactylichnium Stenichnus Strictipes *Synaptichnium "Sphenacopus"*
Tambachichnium Telichnus Testudo Tetrapodichnus Thecodontichnus Tridactylichnium "Therocephalopus" "Theriobematistes" "Theriopezus" "Theriopodiscus"
Varanopus

Fig. 3 – Ichnogeneric names related to tetrapod tracks of Permian formations, as an alphabetical overview of about 140 names. In "quotation marks" are undefined names listed by ELLENBERGER (1983, 1984) from the Permian formations of the Lodève Basin. *Taxa that were originally related to tracks from Triassic beds.

separated taxonomically. This common position has produced about 140 ichnogenera, or ichnogeneric names used or introduced for tetrapod tracks of Permian age (Fig. 3), and thereby it has been the basis for several biostratigraphical subdivisions of continental Permian sequences.

During earlier investigations, up to the early 1980s, one of the authors (HH) was influenced by some of these aspects, especially by the potentially attractive idea of detailed biostratigraphical subdivision of red beds using tetrapod tracks. However, in the face of increasingly divergent and contradictory opinions published by some colleagues, a more skeptical and critical position was established. This position has been justified by the extensive studies of tetrapod tracks in collections and in the field since 1990 in Germany, northern Italy, southern France, northern Spain, and particularly in 1994 and 1999 in New Mexico. The discovery of the Robledo tracksites in southern New Mexico was a test case for the Permian tetrapod ichnotaxonomy and ichnostratigraphy. After the initial analyses it seemed to be the most diverse Permian ichnofauna, with about 30 different taxa (Hunt *et al.*, 1995). However, after the transfer of the huge amount of material collected by J. P. MacDonald to the collection of the New Mexico Musem of Natural History, followed by detailed study, this position was fundamentally modified by the joint investigations of the authors.

In short, in our present state of knowledge the main question is not which 140 generic names and published binominal ichnospecies may be valid, but whether it is the case that only six and no more than 10 different morphotypes are real? All these morphotypes are comparable to and principally represented in the ichnofauna known from the Abo Formation and its equivalents. This interpretation will not be accepted without protest by some other students of the same topic. This might become obvious in several lists of names (ichnotaxa) used in local or regional related descriptions previously. However, the only possible conclusion is that these very different names and interpretations are based only on the same few basic morphotypes and their variations.

One of the most extended examples of ichnotaxonomic oversplitting may be *Batrachichnus* (Fig. 4). It is one of the most common track types of the Permian and Permocarboniferous red beds. The perpetrators of *Batrachichnus* can be interpreted as juvenile temnospondyls. Their locomotory abilities together with their small size are the reasons for the large range of variability in track preservation. Like *Batrachichnus*, the other basic morphotypes *Amphisauropus*, *Dromopus* and *Dimetropus*, are discussed by Haubold (1996). *Tambachichnium* is closely related to *Dromopus* (Haubold, 1998), while for *Ichniotherium* see Voigt (this volume). *Hyloidichnus* and *Gilmoreichnus* have yet to be discussed in their taxonomic context together with an objective documentation of their variation.

POSSIBILITY OF A CONSENSUS

The difficulities in reaching a consensus on the outlined basic morphotypes might be due to

- subjective positions,
- ignorance of the phenomenon of extramorphology in track preservation,
- the use of less representative type specimens (phantom taxa), and
- principally, differences in ichnotaxonomic procedures.

These aspects seem to be the background of some formalistic, less constructive discussions, which have as their only aim the conservation of some personal, subjective results in faunal analyses related to problematic ichnotaxa and derived biostratigraphic subdivisions. We may summarise the experience of two positions in the understanding of Permian tetrapod track faunas from the red beds:

The concentration of individual discoveries and local separation reflects increasing degrees of difference.

The larger the sample sizes and the broader the range of comparison, the smaller the differences in taxonomy and stratigraphy.

At the present time, we see problems in making the second position understandable, if not acceptable.

- Ichnofaunas hitherto separate in time and taxonomy are much more uniform,
- so the age of these faunas and the trackbearing formations may be more or less restricted to Wolfcampian age, and
- our position is in conflict with many traditional and personal opinions.

However, if tetrapods and tetrapod footprints are a key element of Permocarboniferous stratigraphy, then it should be accepted consequently as one possible stratigraphic model. The current increasingly extensive record of tetrapod tracks in the Abo Formation of New Mexico is evidence of the Wolfcampian age of most tetrapod track faunas known in Europe.

Some seemingly problematic exceptions concern the Permian trackbeds in southern France, particularly eastern Provence. Visscher (1968) determined the flora of sites at Agay (Pradineaux Formation, in the Estérel Basin) and le Muy (Muy Formation, in the Bas-Argens Basin) as Upper Permian. He concluded that the basin of Estérel, an important centre of volcanic activity, was of Thuringian, Upper Permian, age. In their analyses of the ichnofaunas, Demathieu & Gand (in Demathieu *et al.*, 1991) argued for a lower Kazanian or Kungurian age for the Pradineaux Formation. Gand *et al.* (1995) described from the site St. Se-

Krkonoshe Basin, Czech Rep. Kalna-Formation	GEINITZ, 1861 FRITSCH, 1895, 1901 PABST 1908	*Saurichnites salamandroides* *? Saurichnites caudifer* *? S. comaeformis* Ichn. rhopalodactylum kalnanum
Thüringer Wald, central Germany Goldlauter and Oberhof Formations	PABST, 1908 HAUBOLD, 1970, 1973 etc.	*Ichn. anakolodactylum* I. brachydactylum kabarzense *Anthichnium salamandroides* *Gracilichnium jacobii* *Gilmoreichus brachydactylus* *Gilmoreichnus minimus* *Jacobiichnus caudifer*
Southern Alps, Collio Formation	DOZY, 1935 CEOLONI *et al.*, 1988	*Anhomoiichnium orobicum* *Gracilichnium berrutii*
Lodève and St. Affrique Basins, France Formation of Tulières to Rabejac and Pelites of St.Rome to St.Pierre	HEYLER & LESSERTISSEUR, 1963 GAND, 1987, 1993, GAND & HAUBOLD, 1984	*Diversipes proclivis* *Foliipes abscisus* *Devipes caudatus* *Crenipes abrectus* *Crenipes abscurvus* *Acutipes decessus* *Strictipes regularis* *Nanipes minutus* *Serripes pectinatus* *Anthichnium salamandroides* *Limnopus regularis* *Salichnium pectinatus* *Salichnium decessus*
Saar-Nahe Basin, SW Germany Altenglan Formation to Nahe-Group	FICHTER, 1983 a, b, 1984 BOY & FICHTER, 1988	*Saurichnites slamandroides* *S. intermedius* *S. incuvatus* *Amphisauroides sp.* *Foliipes abscissus* *Gilmoreichnus minimus* *G. kablikae (brachydactylus)* *Hyloidichnus arnhardti* *Jacobiichnus caudifer* *Limnopus palatinus*
Hermit Shale, Arizona	LULL, 1918 GILMORE, 1927	*Exocampe delicatula* *Batrachichnus delicatulus* *Batrachichnus obscurus* *Dromillopus parvus*
Hueco Formation, Robledo Mountains Member, New Mexico	SCHULT, 1995 HAUBOLD *et al.*, 1995 a	*Anthracopus ellangowensis* *Batrachichnus plainvillensis* *Chelichnus bucklandi* *Cursipes dawsoni* *Dromillopus quadrifidus* *Foliipes caudatus* *Hyloidichnus bifurcatus* *Limnopus regularis* *Laoporus cf. L. nobeli* *Nanopus caudatus* *Quadropedia prima* *Salichnium becki* *Batrachichnus delicatulus*

Fig. 4 – *Batrachichnus salamandroides*, an extended example of ichnotaxonomic oversplitting. Overview of the presumed synonyms and phantom taxa, separated by formations and authors.

bastien (in Saint Raphael of the Estérel Basin) tracks from an extended surface with both traditional and several new names. The determinations, as well as the stratigraphic interpretation, raise some questions. Our own investigations of the ichnofauna of St. Sebastien show exceptional extramorphological preservation on a rhyolitic tuff surface. It is, in contrast to Gand *et al.*, a member of the Mitan Formation, which is intercalated between the Pradineaux and Muy formations. The stratigraphic position and the track assemblage of St. Sebastien, as well as all track faunas of the Pradineaux, Mitan, and Muy formations, can be interpreted as Lower Permian, or Wolfcampian.

More compatible with a Wolfcampian interpretation appears to be the correlation of the track formations from the Saint Affrique and Lodève basins (except La Lieude) as Asselian, Sakmarian and ?Artinskian, after Gand (1993). Only a few names listed by Gand (1993) differ from the above-suggested basic morphotypes, such as *Anthichnium*, *Salichnium*, and *Varanopus*. In our interpretation they are synonyms of *Batrachichnus* and *Hyloidichnus*.

As a result, we can present evidence of only one Early Permian Euramerican terrestrial tetrapod fauna, and in consequence a restriction in time for most related red bed sequences in Europe to approximately Wolfcampian age. This conclusion seems compatible with the trend derived from other observations, as well as from biostratigraphy and radiometric data, which point more and more to a restriction of most European Permocarboniferous red bed sequences to the lowermost Permian. We hope to have presented our results with care and acceptance of divergent opinions. Our result may be understood as the initiation of a forthcoming constructive discussion.

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A PERMIAN TIME SCALE 2000 AND CORRELATION OF MARINE AND CONTINENTAL SEQUENCES USING THE ILLAWARRA REVERSAL (265 MA)

MANFRED MENNING[1]

Key words – Permian; time scale; geochronology; isotopic ages; magnetostratigraphy; global correlation.

Abstract – Synthesis of an increased number of reliable U-Pb, Ar-Ar, and Rb-Sr ages, and field indicators of the duration of Permian stages provide a time scale that differs significantly from those of Harland *et al.* (1982, 1990), Odin (1982, 1994), and Gradstein & Ogg (1996). The duration of the Lower Permian Series (Cisuralian: Asselian, Sakmarian, Artinskian, Kungurian) is comparable with the duration of the Middle Permian Series (Guadalupian: Roadian, Wordian, Capitanian) plus the Upper Permian Series (Lopingian: Wuchiapingian, Changhsingian).
U-Pb zircon SHRIMP ages from the southern Ural Mountains indicate 292 Ma for the Carboniferous-Permian (Gzhelian-Asselian) boundary. In contrast, Ar-Ar sanidine ages from Central Europe provide about 296 Ma for that boundary. Using available data, the best estimate for the Permian-Triassic boundary is 251 ±1 Ma.
The 265 Ma Middle Permian Illawarra Reversal (IR) of the Earth´s magnetic field (Irving & Parry, 1963) is not positioned consistently with traditional Permian correlations. Referred to regional mapping units the Illawarra Reversal is within the Lower Permian of Central Europe (Rotliegend, continental) and South China (Maokou, marine) but in the Upper Permian of East Europe (Tatar, continental), North America (Guadalupe, marine), and North China (Upper Shihhotse Formation, continental). Thus, the duration of deposition of sequences mapped as "Upper Permian" ranges from about 7 Ma (Zechstein) to 23 to 20 Ma (Ufa+Kazan+Tatar ~ Guadalupe+Ochoa).
A preliminary estimation of the duration of lithostratigraphic units for the described stratigraphic profiles, together with SE Australia, is provided.

Parole chiave – Permiano; geocronologia; età isotopiche; scala cronologica; magnetostratigrafia; correlazione globale.

Riassunto – La sintesi di un crescente numero di attendibili età U-Pb, Ar-Ar e Rb-Sr, nonché le indicazioni di età e durata dei piani permiani forniscono una scala cronologica che differisce sensibilmente da quelle di Harland *et al.* (1882, 1990), Odin (1982, 1994) e di Gradstein & Ogg (1996). La durata del Permiano inferiore (Cisuraliano: Asseliano, Sakmariano, Artinskiano, Kunguriano) è confrontabile con la durata complessiva del Permiano medio (Guadalupiano: Rodiano, Wordiano, Capitaniano) e del Permiano superiore (Lopingiano: Wuchiapingiano, Changhsingiano).
Le età U-Pb ottenute mediante microsonda SHRIMP da zirconi presenti negli Urali meridionali indicano 292 Ma per il limite Carbonifero-Permiano (Gzheliano-Asseliano). Al contrario, le età Ar-Ar inerenti a cristalli di sanidino provenienti dall'Europa centrale forniscono all'incirca 296 Ma per questo limite.
L'inversione medio-permiana "Illawarra" (IR) di 265 Ma del campo magnetico terrestre (Irving & Parry, 1963) non è posizionata in modo consistente con le tradizionali correlazioni permiane. La cartografia regionale pone l'"Illawarra Reversal" entro il Permiano inferiore dell'Europa centrale (Rotliegende, continentale) e della Cina meridionale (Maokou, marino), ma entro il Permiano superiore dell'Europa orientale (Tatar, continentale), del Nord America (Guadalupa, marino) e della Cina settentrionale (Formazione di Shihhotse Superiore, continentale). Conseguentemente, il tempo di deposizione di successioni cartografate come "Permiano superiore" va da circa 7 Ma (Zechstein) a circa 23-20 Ma (Ufa+Kazan+Tatar ~ Guadalupe+Ochoae). È data una stima preliminare delle unità litostratigrafiche per i descritti profili stratigrafici, annettendo anche l'Australia sud-orientale.

INTEGRATIVE TIME SCALE CALIBRATION

Refinement of the geologic time scale has proceeded by steps. In an integrative calibration, all available time proxies of different types and sources should be used. The most reliable isotopic data must be compared with one another and with field indicators such as the weighted number of biozones, weighted average thicknesses, and sequences of comparable duration located elsewhere.

Using Ar-Ar sanidine and Rb-Sr biotite ages from Central Europe (Lippolt *et al.*, 1984; Lippolt & Hess, 1989) the Gzhelian-Asselian boundary is at about 296 (297) Ma (Menning, 1995) whereas the boundary age is at about 292 Ma according to Pb-U SHRIPM data from zircon from easternmost Europe (Chuvashov *et al.*, 1996; Jin *et al.*, 1997:13; Menning *et al.*, 2000: Fig. 7). For the Kasimovian to Asselian time span the U-Pb zircon ages seem to be systematically younger than stratigraphically corre-

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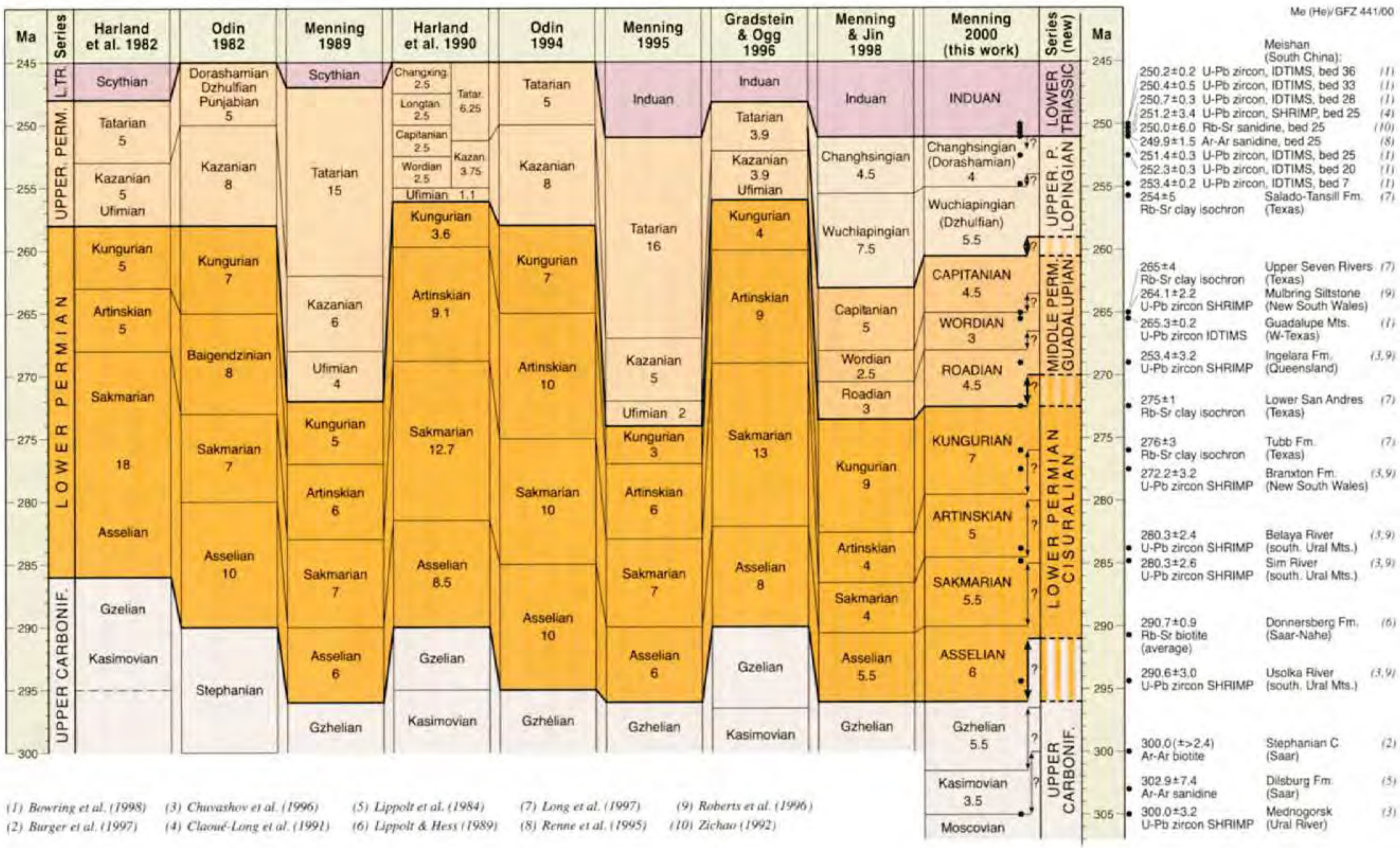

Fig. 1 – Permian time scales. The ages of Harland *et al.* (1982, 1990), Odin (1982, 1994), and Gradstein & Ogg (1996) are significantly younger in part compared with the integrative estimated ages of Menning (1989, 1995) and Menning & Jin (1998). The isotopic ages are shown according to their stratigraphic position. Permian standard stages (Jin *et al.* 1997) accepted by the Subcommission on Permian Stratigraphy of the IUGS are in CAPITAL letters.

sponding Ar-Ar sanidine and Rb-Sr biotite ages (Fig. 1). These age differences correspond to those in the pre-Kasimovian Carboniferous (Menning *et al.*, 2000: Fig. 6, Time Scales A and B). Thus, in the lower part of our Permian time scale 2000 (Fig. 1) the ages, based on Ar-Ar and Rb-Sr data in combination with field-based time indicators, are about 5 Ma older than a calibration using U-Pb SHRIMP ages (Fig. 1, upward arrows).

The 264.1 ± 2.2 Ma age from the Mulbring Siltstone (New South Wales) fits with the time scale if it is allocated to the Wordian-Capitanian boundary (Fig. 1) using the magnetostratigraphic results of Théveniaut *et al.* (1994). But the age of 253.4 ±3.2 Ma from the Ingelara Formation (Queensland) does not fit if a Kazanian stratigraphic position (Chuvashov *et al.*,1996; Roberts *et al.*, 1996) is used (Fig. 1).

Claoué-Long *et al.* (1991) used $^{238}U/^{206}Pb$ SHRIMP data on zircon to obtain 251.2 ±3.4 Ma for the Permian-Triassic boundary. This age is further supported by sanidine dating by the Rb-Sr (Zhang, 1992) and Ar-Ar (Renne *et al.*, 1995) methods, and by U-Pb dating of zircon by ID-TIMS (Bowring *et al.*, 1998). A combination of all of these data available provides 251 ±1 Ma as the best estimate for the Permian-Triassic boundary (Fig. 1) whereas an age of ~253 Ma is favoured by Mundil *et al.* (2000).

According to the number of conodont, ammonoid and fusulinid zones the duration of the Lower Permian Series (Cisuralian: Asselian, Sakmarian, Artinskian, Kungurian) is comparable to that of the Middle Permian Series (Guadalupian: Roadian, Wordian, Capitanian) plus the Upper Permian Series (Lopingian: Wuchiapingian, Changhsingian). If we assign 296 Ma to the Carboniferous-Permian boundary (Menning, 1995) and 251 Ma to the Permian-Triassic boundary (Claoué-Long *et al.*, 1991), this results in a Kungurian-Ufimian boundary age about halfway between: *i.e.* 275 Ma to 270 Ma (Menning, 1995).

According to a duration of the Permian of about 45 Ma and to weighted average thicknesses of 11 stratigraphic sections from five continents the Middle Permian (Guadalupian) Illawarra Reversal is 10 ±4 Ma older than the Permian-Triassic boundary (Menning, 1986). Consideration of the number of fusulinid, ammonoid, and conodont zones suggests that the reversal is 14 Ma older than the Permian-Triassic boundary, which, according to Menning (1995), is 265 Ma (Illawarra Reversal).

The Kungurian could be the longest Permian stage if its base is defined with the first occurrence of the conodont *Neostreptognathodus pnevi* and the base of the Roadian coincides with the first occurrence of *Mesogondolella nankingensis* (Kozur, 1995, 1997). Such a definition of the Kungurian would include the uppermost part of the original Artinskian and the lower part of the Ufimian (Kotlyar, 2000) (Figs 1, 3).

In summary, recently obtained isotopic ages (Chuvashov *et al.*, 1996; Roberts *et al.*, 1996; Long *et al.*, 1997; Burger *et al.*, 1997; Bowring *et al.*, 1998) from samples that can be positioned within the new Permian stratigraphic standard scale confirm the ages estimated by Menning (1989, 1995) on geological grounds. Discrepancies, which are no more than 5 Ma, fit mostly within the $\pm 2\sigma$ errors of the isotopic ages. These ages differ significantly from the time scales of Harland *et al.* (1982, 1990), Odin (1982, 1994), and Gradstein & Ogg (1996) (Fig. 1), but only moderately from the fusulinid-related time scale of Davydov *et al.* (1999) and the conodont-related time scale of Wardlaw & Schiappa (this volume). More data of diverse sorts (U-Pb, Ar-Ar, palaeontological, sequence-stratigraphic) are necessary to provide more reliable estimates of the duration of deposition of Permian stratigraphic units.

BIOSTRATIGRAPHIC CORRELATION

Extremly strong provincialism is a characteristic of all Permian fossil groups. Conodonts, fusulinids and ammonoids serve as index fossils for global correlation of marine sequences. Tetrapod body and trace fossils, shark teeth, macroplants, arthropod tracks, insect wings, and palynomorphs are used to correlate continental sequences.

Only few stratigraphic sections contain an interfingering of continental and marine beds, making intercalibration possible. Palynomorphs are the most usable fossil in this regard, but they are not found in red beds unless gray intercalations are also present – a situation that is rare in Pangaea. Moreover, Permian floral realms were distinctive and for this reason interregional palynological correlation is limited.

GEOLOGICAL MAPPING UNITS

Mapping units are defined for each geographic area on a practical basis, by observable lithologic properties. A mapped succession of rocks is assigned roughly to a unit of the global stratigraphic reference scale (Fig. 2). For instance, the Rotliegend of Central Europe was assigned to "Lower Permian" and the Zechstein to "Upper Permian" (Fig. 2). However, according to recent biostratigraphic, magnetostratigraphic and isotopic age evidence, the base of the Rotliegend lies in the Upper Carboniferous Gzhel and its top lies in the Tatar of Upper Permian age (Fig. 2). The Chihsia and Maokou limestones in South China were mapped as "Lower Permian" and Longtan and Changhsing rocks as "Upper Permian" (Fig. 2). However, the time span represented by the "Lower Permian" Rotliegend

Group is more than 300 Ma to 260 Ma, whereas the accumulation of "Lower Permian" Chihsia and Maokou Formations was between about 280 Ma to 260 Ma, a much shorter duration (Fig. 3).

Correlation of mapping units that are ostensibly "Lower Permian" (Fig. 2) but which accumulated over different time spans (cf. Fig. 3) can create confusion. For instance, consider the information from early Permian palaeomagnetic poles. A more reliable palaeogeographic reconstruction results if we combine palaeomagnetic poles from South China (Chuanshan = "Upper Carboniferous" + Chihsia = "lower Lower Permian") with "Lower Permian" poles from East Europe and North America (Figs 2, 3).

Fig. 2 presents stratigraphic units of five areas assigned in regional geological maps respectively to the "Lower Permian" and "Upper Permian". The inconsistent position of the Illawarra Reversal (IR) (Irving & Parry, 1963) argues that the correlation of most of those Permian mapping units used in several correlation charts must be incorrect (Fig. 2).

Durations of deposition of "Upper Permian" sequences (Fig. 2) vary from about 7 Ma (Zechstein), to about 9 Ma (Longtan+Changhsing), to about 16 Ma (Upper Shihhotse+Shihchienfeng), to about 23 to 20 Ma (Ufa+Kazan+Tatar ~ Guadalupe+Ochoa) (Fig. 3).

Ar-Ar and Rb-Sr ages (Lippolt *et al.*, 1984; Lippolt & Hess, 1989) and our integrative time analysis assign a maximum duration of 10 Ma (301-291 Ma) for deposition of the Lower Rotliegend of the Saar-Nahe Basin (Kusel+Lebach+Tholey). For the Upper Rotliegend of Central Europe the duration increases to about 34 Ma which is several times longer than previously thought (Fig. 3). In Rotliegend basins the boundary between "Lower" and "Upper" Rotliegend deposits seems to be significantly time transgressive. This is indicated in Fig. 3 by a diagonal dashed line.

MAGNETOSTRATIGRAPHIC CORRELATION

The IR is the best magnetic time marker within Palaeozoic

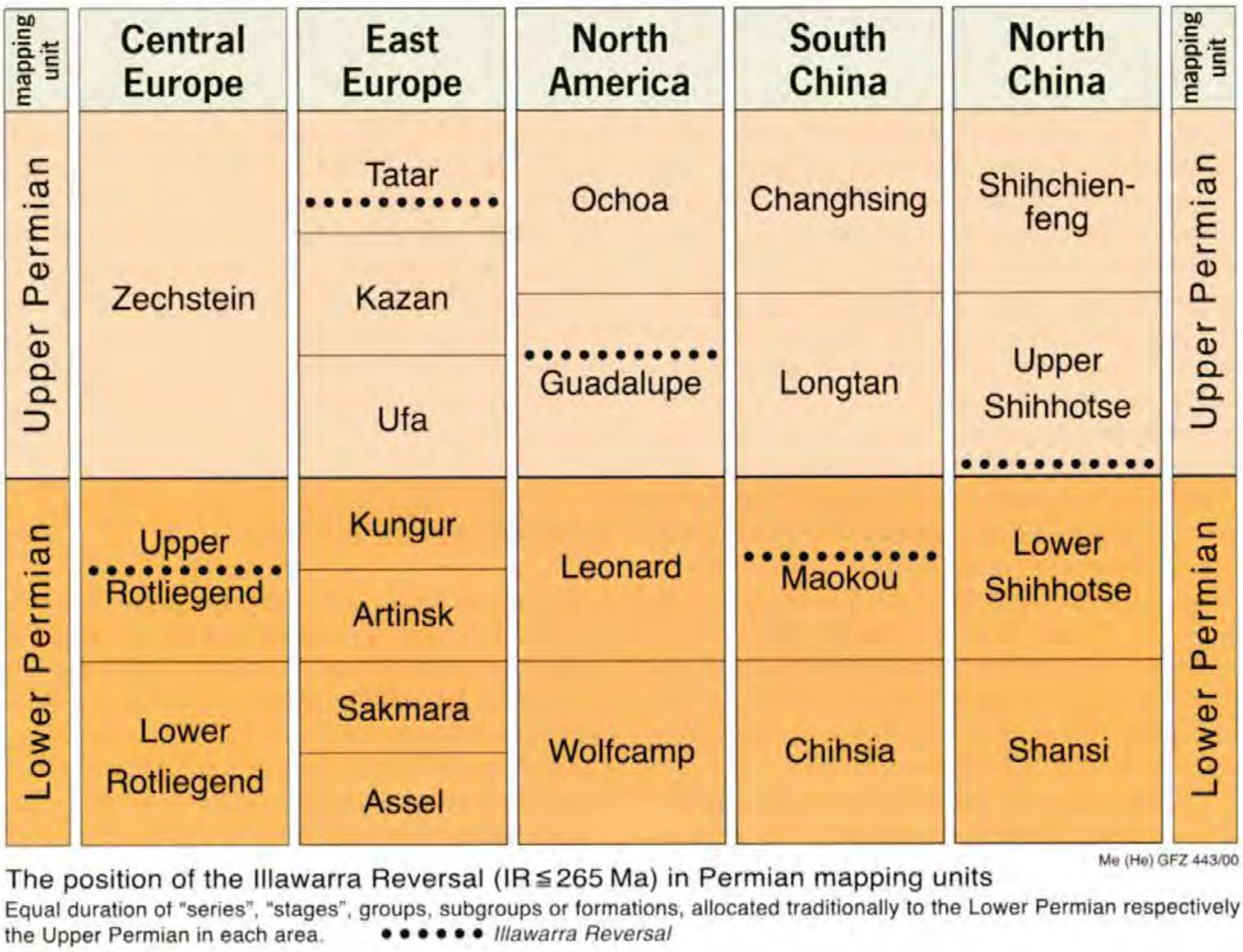

Fig. 2 – Stratigraphic scheme showing Lower Permian and Upper Permian mapping units of five areas. The Illawarra Reversal (dotted lines) is within the Upper Permian (Tatar, Guadalupe, Upper Shihhotse) but also in the Lower Permian (Upper Rotliegend, Maokou). It is detected within marine (Guadalupe, Maokou) and continental (Tatar, Rotliegend, Shihhotse) units. Equal duration of "series", "stages", groups, subgroups or formations.

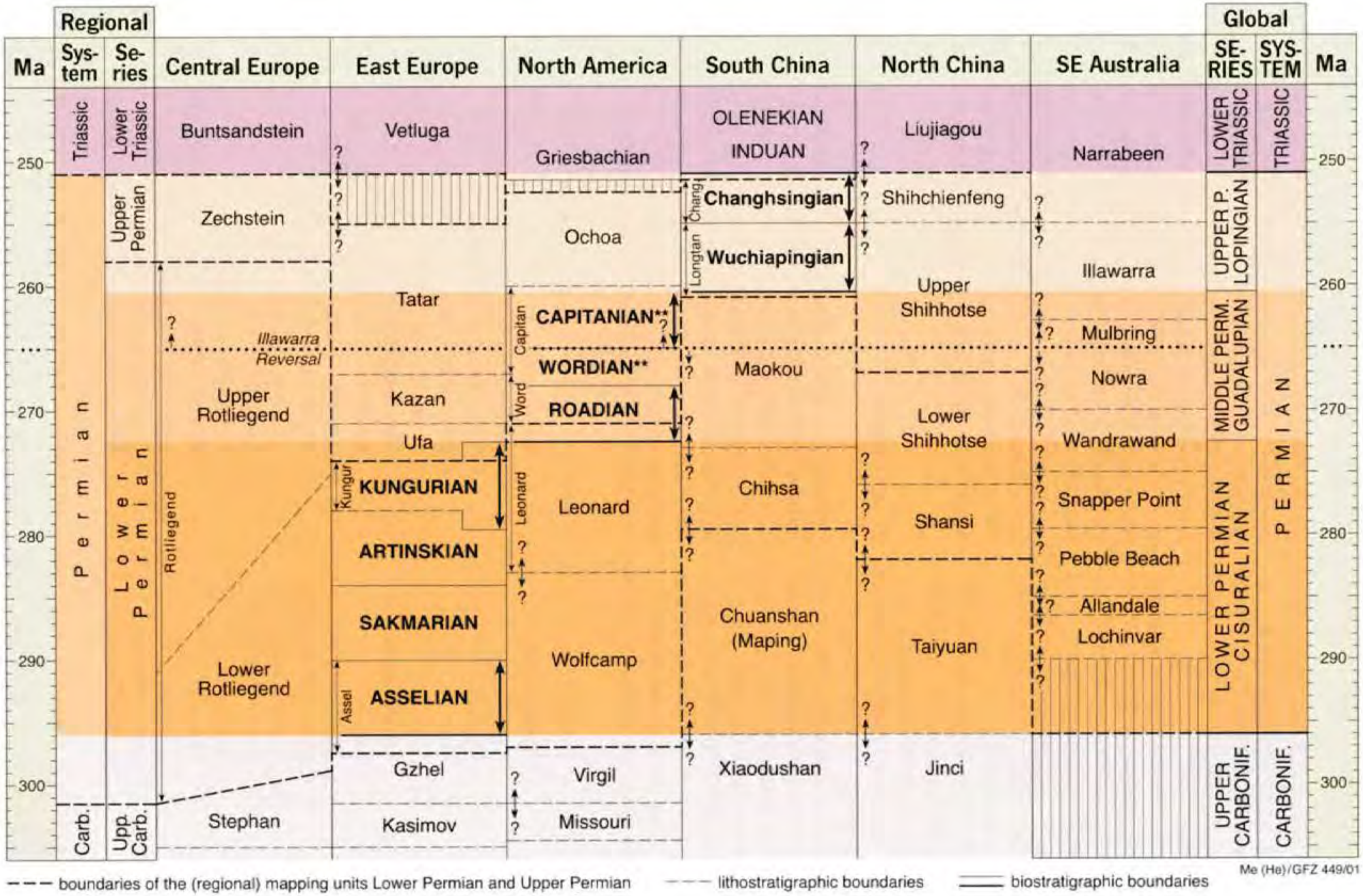

Fig. 3 – Revised stratigraphic scheme using an integrative time scale calibration, particularly magnetostratigraphic information. Preliminary estimation of age and duration of important Permian stratigraphic units. Permian standard stages (Jin *et al.*, 1997) are shown in bold. Standard stages accepted by the Subcommission on Permian Stratigraphy of the IUGS are in CAPITAL letters.

*The (increased) KUNGURIAN is defined by the first occurrence of *Neostreptognathodus pnevi* at the base and the first occurrence of *Mesogondolella nankingensis* at the base of the Roadian (Kozur, 1995, 1997).

**The chronostratigraphic units WORDIAN, CAPITANIAN, ASSELIAN, KUNGURIAN, ARTINSKIAN, WUCHIAPINGIAN, and CHANGHSINGIAN (Jin *et al.*, 1997) differ from the lithostratigraphic units Word, Capitan, Assel, Kungur, Artinsk, Longtan, and Changhsing as used earlier (*e.g.* Dunbar *et al.*, 1960).

rocks. It marks the boundary between the Carboniferous-Permian Reversed Megazone (CPRM = Kiaman Magnetic Division of Irving & Parry, 1963) and the Permo-Triassic Mixed Megazone (PTMM = Illawara (RN) Hyperzone of Molostovsky *et al.*, 1976). CPRM and PTMM refer to rock, and the corresponding time terms are Permo-Carboniferous Reversed Superchron (PCRS = Kiaman Magnetic Interval of Irving & Parry, 1963) and Permo-Triassic Mixed Superchron (PTMS).

The number of magnetic zones within the CPRM and PTMM is unknown, and consequently many global correlations are speculative. In the pre-Illawarra Permian System (ca 31 Ma) there could be as many as four zones of short-duration normal polarity. The post-Illawarra Permian (ca 14 Ma) may contain about 15 magnetic zones (Menning, 1995; Embleton *et al.*, 1996).

The very existence of the CPRM is under discussion although nearly all uppermost Carboniferous and Lower Permian sequences are reversely magnetized. Have some normally polarized rocks been overlooked or ignored as they were interpreted as the result of remagnetization?

Among all Permian magnetostratigraphic markers the IR can be best used for intercontinental correlation. Its numerical age is about 265 Ma (Menning, 1995; Menning, 1992 (pers. comm.) in Opdyke, 1995:42), which is significantly older than estimated earlier. The numerical age of the IR and its stratigraphic position according to the East European reference section changed as shown in Tab. 1.

The IR has been found within the continental lower Tatarian (Upper Permian) of East Europe (Khramov, 1963 ff.; Gialanella *et al.*, 1997) whereas its age has been interpreted to be Ufimian (Théveniaut *et al.*, 1994; Klootwijk *et al.*, 1994; Opdyke, 1995; Embleton *et al.*, 1996) or Kungurian (Ogg, 1995) (Tab. 1). Menning (2001) summarizes details of the arguments. Moreover, the IR is said to be positioned in the continental Lower Permian (Upper Rotliegend) of Central Europe (Dachroth, 1976), the back-reef Upper Permian (Guadalupe) of North America (Peterson & Nairn, 1971), the marine Lower Permian (Maokou) of South China (Heller *et al.*, 1995), and the continental Upper Permian (Upper Shihhotse) of North China (Embleton *et al.*, 1996) (Figs 2, 3).

Guadalupian marine sequences in SW North America and in South China are correlated using conodonts (Kozur, 1997). It is important to discover the accurate position of the IR to confirm their biostratigraphic correlation.

CONCLUSIONS

Our integrative time analysis (Fig. 1), the inconsistent position of the IR in units mapped as "Lower Permian" and "Upper Permian" (Fig. 2), and the revision of the age of the IR to an older value (Tab. 1) have important consequences for a Permian global stratigraphic correlation, particularly for the Upper Permian, and for estimates of the duration of important Permian stratigraphic units (Fig. 3).

In the future, all regional Permian stratigraphic units

Age [Ma]	Stratigraphy	Reference
—	Tatarian	Khramov (1963)
235	Tatarian	Khramov *et al.* (1974, 1982)
---	Tatar	Dachroth (1976)
250	Tatarian	Harland *et al.* (1982)
255*/257*/259**	Lower Tatarian	Menning (1986, 1991, 1992)
261	Tatarian	Haag & Heller (1991)
—	Lower Tatarian	Solodukho *et al.* (1993)
—	Ufimian/Kungurian	Théveniaut *et al.* (1994)
>267	Ufimian/Kungurian	Klootwijk *et al.* (1994)
≤265***	Lower Tatarian	Menning (1995)
261	Kungurian	Ogg (1995)
262	Ufimian	Opdyke (1995), Opdyke & Channel (1996)
—	Ufimian	Embleton *et al.* (1996)
—	Lower Tatarian	Gialanella *et al.* (1997)
265***	Lower Tatar(ian)	Menning & Jin (1998), Menning (this work)

*10 ± 4 Ma older than the Permian Triassic boundary (Menning, 1986)
**10 + 2 Ma older than the Permian Triassic boundary
***10 + 4 Ma older than the Permian Triassic boundary

Tab. 1 – Numerical age and stratigraphical position of the Illawarra Reversal according to the East European reference section.

should be referred (if possible) to the revised global Permian chronostratigraphic reference scale (Jin *et al.*, 1997) which is subdivided into a Lower Permian Series (Cisuralian), a Middle Permian Series (Guadalupian), and an Upper Permian Series (Lopingian) (Figs 1, 3). Until the significant disagreement of Permian time scales (Fig. 1) is resolved, one should quote the reference and the stratigraphic source of each numerical age.

In stratigraphic figures and tables, the global chronostratigraphic and geochronologic terms that are recommended by the International Stratigraphic Commission should be written in CAPITAL letters to support the scientific communication (Figs 1, 3).

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TOWARD A REFINED PERMIAN CHRONOSTRATIGRAPHY

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Key words – Permian scale; conodont zonation; numerical ages.

Abstract – The generalized standard marine conodont zonation for the Permian of 24 zones is placed within a framework utilizing the presently available reliable radiometric as well as estimated numerical ages. The base of the Permian is taken to be 291.6 Ma and the top to be 251.4 Ma.

Parole chiave – Scala del Permiano; zonazione a conodonti; età numeriche.

Riassunto – La generalizzata zonazione standard a conodonti marini relativa al Permiano di 24 zone è inserita in uno quadro strutturato che utilizza le attendibili età radiometriche e le stimate età numeriche attualmente disponibili. Il limite inferiore del Permiano è correntemente riferito a 291.6 Ma e il limite superiore a 251.4 Ma.

INTRODUCTION

This is a working document to: 1) encourage more rigorous studies on radiometric age dating within biostratigraphically well constrained sections; 2) further discussions of the 'standard' Permian scale; and 3) to better correlate the marine to continental sections. The premise was to use recent age-dates thought reliable (Chuvashov *et al.*, 1996; Menning, 1995; Bowring *et al.*, 1998) and fit them into the evolving generalized standard marine conodont zonation as currently recognized by the authors. Each conodont zone represents time, and therefore, radiometric ages that suggested no time elapsed for a zone were extended to their maximum error range to reflect an elapsed time.

CONODONT ZONATION

The Lower Permian (Cisuralian Series) conodont zonation is from a variety of published and unpublished sources. The Asselian and Sakmarian zonation, based on the succession of *Streptognathodus* species is from Chernykh *et al.* (1997), Boardman *et al.* (1998), and Wardlaw *et al.* (1999). This succession is well represented in Kansas and the southern Ural Mountains of Russia.

The Artinskian zonation is based on species of *Sweetognathus*, *Streptognathodus*, and *Neostreptognathodus* and reflects the major changeover in the forms dominating shelf faunas during this interval. It is largely based on unpublished material from the southern Urals, Russia and Kazhastan and the Great Basin, USA.

The Kungurian zonation is based on the succession of *Neostreptognathodus* species modified from Wardlaw & Grant (1987) from West Texas, USA.

The Middle Permian (Guadalupian Series) conodont zonation is from Wardlaw & Lambert (1999) except that the rapid succession of upper Guadalupian *Jinogondolella* (*altudaensis*, *prexuanhanensis*, *xuanhanensis*, and *crofti*) are all overlapped by *J. altudaensis* and considered as subzone indicators of that zone. This succession is well represented in West Texas and South China.

The Upper Permian (Lopingian Series) conodont zonation is from Mei *et al.* (1994, 1998) as modified by Wardlaw & Mei (1998) and reflecting the change of selecting the first appearance *Clarkina dukouensis* in an evolutionary cline from *C. postbitteri* as a more appropriate base of the Wuchiapingian. This succession is well represented in both South China and the Dzhulfa area of Iran and Transcaucasia.

In all these generalized zones, more local zones (or subzones) are recognized. In particular, the Asselian and Sakmarian Stages of Kansas (based on many more species of *Streptognathodus*), the Kungurian Stage of West Texas (based on the concurrent succession of *Mesogondolella* species with the *Neostreptognathodus* species), and the Lopingian Series of China (based on more species of *Clarkina*).

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Informal names such as *"exsculptus"*, *postwhitei*, *florensis*, and *trimilus* are those proposed in various taxonomic works involving one or both of the authors currently in various stages in the publication process.

IMPORTANT RADIOMETRIC AGES

Chuvashov *et al.* (1996) report one very well biostratigraphically constrained SHRIMP (Super High Resolution Ion Microprobe) analysis date for the base of the *constrictus* (and corresponding local fusulinid zone) of 290.6 ± 3.0 Ma which we feel is very close to on-the-mark. By projection we suggest that the *isolatus* zone is no more than 1 million years in duration and that 291.6 probably provides a very good age for the base of the Permian. Chuvashov *et al.* (1996) also report less well constrained ages for the uppermost Sakmarian (280.3 ± 2.4) and the lowermost Artinskian (280.3 ± 2.6). We project an age of 283 Ma for the boundary as defined by conodonts, which falls close to the margin of error for both dates. The dates utilized by Chuvashov *et al.* (1996) from Australia, although important, are not well enough constrained to be used in building the standard time scale.

Bowring *et al.* (1998) discuss several important ID-TIMS (Isotope Dilution Thermal Ionization Mass Spectrometry) dates, of which a few are very well constrained within the proposed stratotypes for the Middle and Late Permian. In particular, an age of 265.3 ± 0.2 Ma from just below the recently approved GSSP for the base of the Capitanian coincides with the estimated age of 265 Ma by Menning (1995) for the Illawarra Reversal. Menning (in Glenister *et al.*, 1999) places the Illawarra Reversal within this important section between the horizon isotopically dated and the conodont defined base of the Capitanian.

Bowring *et al.* (1998) report several dates from the Meishan section in China, the proposed GSSP for both the Changhsingian and the Permian-Triassic boundary. A date of 253.4 ± 0.2 Ma is derived from bed 7, immediately below the first occurrence of *Clarkina subcarinata (sensu strictu)*, which is the proposed definition of the base of the Changhsingian Stage. Also, from immediately below the proposed conodont-defined base of the Triassic, the first occurrence of *Hindeodus parvus*, they report an age of 251.4 ± 0.3 Ma.

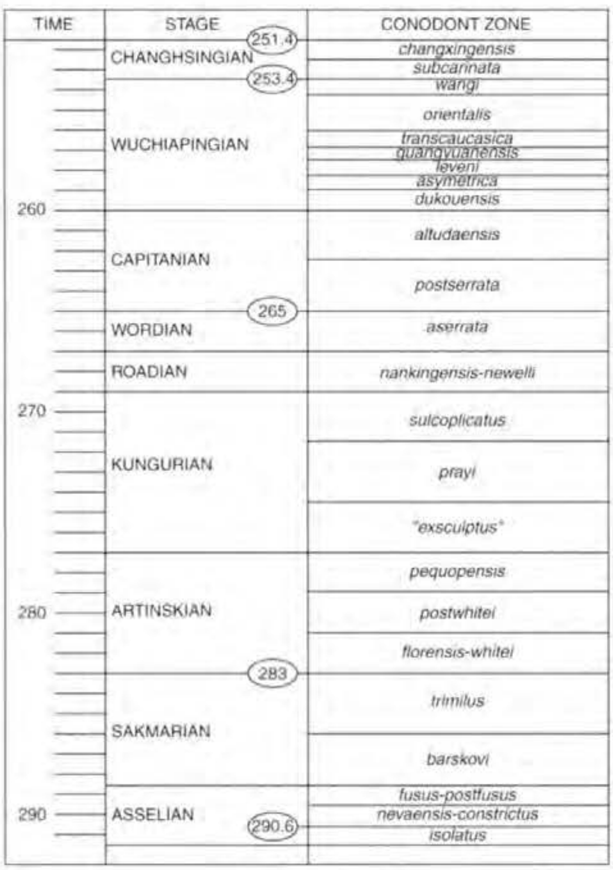

Fig. 1 - Chronostratigraphic chart with age in million years before present, Permian stages, and generalized marine conodont zonation. Ages thought to be reliable are highlighted in ellipses.

CONCLUSIONS

Permian conodont zones appear to range in age from 0.7 to 3.0 million years. This initial study suggests that the stages as now defined represent the following span in years:

Changshingian	2.0 million years
Wuchiapingian	6.6 million years
Capitanian	5.0 million years
Wordian	2.0 million years
Roadian	2.0 million years
Kungurian	8.0 million years
Artinskian	6.0 million years
Sakmarian	5.6 million years
Asselian	3.0 million years

The base of the Permian is taken to be 291.6 Ma at the base of the *isolatus* conodont zone. The base of the Triassic (and therefore, the top of the Permian) is taken to be 251.4 Ma at the base of the *parvus* conodont zone.

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MEETING REPORT

GIUSEPPE CASSINIS[1]

GENERAL

The international Congress on "The continental Permian of the Southern Alps and Sardinia (Italy). Regional reports and general correlations" was held in Italy from 15-25th September 1999. It followed another meeting, in July 1986, which was set up by a team of Italian researchers involved in the IGCP project No. 203 ("Permo-Triassic events of East Tethys region and their intercontinental correlation"), who have been working for years in the Southern Alps. Brescia, in eastern Lombardy, was the official centre of both the initiatives.

The last meeting was inspired by a number of scientific activities. In particular, the wish of the IUGS Subcommission on Permian Stratigraphy (SPS) to create a "Continental Permian Group", and the consequent decision of its Officers to give impetus to this research, was a determining factor in organising the Brescia Field Conference. The meeting has also greatly benefited by funds from the Ministry of University and Scientific and Technological Research (MURST), and the National Research Council (CNR), for a new study programme on the Alpine and Apennine Late Paleozoic evolution, compared with the contemporaneous framework of Sardinia and other western Mediterranean areas. Further, the Brescia Museum of Natural Sciences generously gave financial support for the publication of these proceedings.

Due to the participation of foreign experts in Carboniferous to Triassic continental successions, and the large number of oral and poster contributions outside the selected areas of the meeting (Southern Alps and Sardinia), the subject of these Proceedings has been changed accordingly. The new title "Permian continental deposits of Europe and other areas. Regional reports and correlations" corresponds better to the contents of this volume.

The Congress consisted of a scientific Conference held in the Civic Museum of Natural Sciences in Brescia, and two field trips carried out in Sardinia and in the central-eastern Southern Alps, respectively. Over 100 participants came generally from Europe, and some from North American and Asiatic countries.

THE PRELIMINARY EXCURSION IN SARDINIA (15-18 SEPTEMBER)

The visit to some major Late Paleozoic continental basins on the island was the aim of this trip, of which the itinerary and stops are indicated in Fig 1. The participants met in the maritime station of Porto Torres, in northwestern Sardinia.

15 September (Speakers: Barca, Del Rio, Pittau)

The first day was dedicated to the examination of the Carboniferous to Permian continental deposits of the small San Giorgio (Iglesiente) and Guardia Pisano (Sulcis) basins. However, due to the long distances involved, the or-

Fig. 1 - Excursion itinerary (15-18 September) and main stops on some Late Paleozoic continental basins of Sardinia.

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ganisers made a preliminary stop near Barrumini village, to visit the famous archaeological site of "Su Nuraxi". The participants appreciated very much the opportunity to explore this fortress ("Nuraghe"), which is dated to 1500-1300 BC.

The San Giorgio Basin essentially consists of a detrital succession, about 30 m thick, which unconformably overlies a Cambrian basement. Polymict conglomerates pass upwards to dolostones alternating with sandstones and conglomerates, pelites, shales and coals rich in fossil plants; coarse- to fine siliciclastic sediments, again bearing floristic remains, follow above. The macro- and microflora are generally related to "Stephanian" times.

The Guardia Pisano Basin is made up of both fluvial-to-lacustrine clastic sediments and volcanic products (Fig. 2). This succession, about 100 m thick, consists of grey shales more or less carbonaceous and rare lenticular sandstone with plants and palynomorphs; volcanic breccias of rhyolitic-rhyodacitic lavas (295±5 Ma) alternating with coal-shales, sandy dolostone and tuffs rich in conifer fragments; coarse- to fine siliciclastics; and reddish shales with intercalations of silty sandstone-to-conglomerate, locally with cross-bedding structures and burrows. On the top, Eocene limestones crop out unconformably. The palynological data may correlate with the "Stephanian-Autunian" of Western Europe, the Wolfcampian of the North America Midcontinent and the Ghzelian-Asselian of the Donetz and Urals basins.

Later, after a long trip, the group reached the Sant'Elmo Beach Hotel (Costa Rey), on the southeastern coast of the island.

16 September (Speakers: Broutin, Cortesogno, Gaggero, Pittau, Ronchi)

The second day was spent visiting the Escalaplano (Gerrei) and Perdasdefogu (Ogliastra) volcanic and sedimentary basins. In the former, the investigated S. Salvatore section, about 170 m thick, can be subdivided into two sequences. The lower one (80 m), which unconformably rests on the Variscan crystalline basement, generally consists of, in ascending order: a basal conglomerate (reddish breccia); varicoloured sandstones-pelites alternating with tuffs; rhyolitic cinerites, more or less silicified, with interbedded limestones, and including near the top a black chert layer; and massive rhyolitic ignimbrites. The upper sequence (about 80-90 m) is made up of a breccia body, with metamorphic and volcanic rock-fragments; a volcaniclastic deposit (interpreted as a lahar), dacitic to andesitic in composition; reddish conglomerate, bearing only the rock-elements of the underlying presumed lahar; brownish and dark red volcaniclastic pelites with carbonate-siderite nodules and, at the top, silicified cinerites and limestone with chert layers (hyaloclastites); and an andesitic lava and a massive breccia body. Algae and fossil plants (rare) suggest an Autunian age.

Fig. 2 – Observing the Guardia Pisano Permian outcrops, Sulcis, SW Sardinia.

The second stop in the Escalaplano Basin was at its southeastern edge, along the road going up to the village. Here, Buntsandstein-type red clastic deposits, with chalky and marly-clayey intercalations, unconformably overlie the Variscan metamorphic basement. On the basis of palynomorphs, these deposits generally relate to Anisian times.

The Perdasdefogu Basin, north of this village, crops out over an area of about 25 km^2, with deposits which show marked thickness changes, and lateral sedimentary and volcanic variations. The succession of the Punta Gardiola area can be subdivided into two parts. The lower one, over 100 m thick, initially consists of dark-grey, sandy-shaly clastic sediments, associated with thin conglomerates and a number of calcalkaline, intermediate to acidic volcaniclastic breccias and lavic products. Lacustrine limestones and dolostones (from 30-70 m), affected by diffuse silicification and with black chert of volcanic origin, in places including conglomerates and rare coal layers, follow above. A huge dacite lava flow (Bruncu Santoru, ca. 180 m thick) occurs on the top; elsewhere, this manifestation is associated with, or followed by, rhyolitic volcaniclastic deposits, mostly in the form of tuffs and ignimbrites; dacite and rhyolite dykes, cross-cutting the Variscan basement and the overlying units, also crop out. Middle Jurassic carbonate rocks rest unconformably on the Permian succession.

The above-mentioned finer laminated and carbonate sediments yield a fossil macro- and microflora, amphibians, algal stromatolites, ostracods and fish teeth, which have been related to the Early Permian ("Autunian").

On returning to the Hotel S. Elmo, the participants enjoyed a performance of traditional Sardinian dances and songs.

17 September (Speakers: Broutin, Cassinis, Cortesogno, Gaggero, Sarria)

The Seui Basin, in central Sardinia (Barbagia di Seulo), was the topic of the third day. At the first stop, a panoramic view from a site (the small church of San Sebastiano), located in the southern sector, led the participants to examine the bulk of this trough, which can be considered the most example on the island when studying and refining the local sedimentary and volcanic systems, and the geological evolution.

Sedimentation in the basin again began with an alluvial "Basal Conglomerate". Rhyolite pyroclastic layers suggest early volcanic activity. Upwardly, fine fluvio-lacustrine-to-palustrine clastic and coal sediments generally follow. They are cut by and intercalated with andesitic bodies ("Porfiriti" *Auct.*), in the form of plugs and lavas. The succession is topped by rhyolitic ignimbrites, probably originating in the north of the basin, where they attain more than 500 m in thickness at Mount Perdedu. Meanwhile, along the northern boundary of this small basin, large calcalkaline diorite dykes fed dacitic and rhyolitic domes, which intruded at shallow levels and inflated the Variscan crystalline basement and/or the overlying sediments. Due to dome intrusions, slices of metamorphic basement were emplaced locally over the Lower Permian, volcanic and sedimentary deposits.

A large part of the fossil macroflora ascribes the Seui sediments to the Autunian. However, the chronostratigraphy of the overlying volcanic rocks, due to the irregular radiometric dates and the need for more careful correlation with other volcanic sections of Sardinia, deserves additional study.

The second stop in the basin was preceded by a rustic but substantial lunch in the field (with "porchetta" and other Sardinian foods, accompanied by local red wine, myrtle liqueur and "Filu Ferru" brandy), kindly prepared by the native Calzia family (Fig. 3). This surprise lunch allowed the participants to appreciate greatly some of the typical and famous produce of the island.

Stop 2, on the northern side of the Seui Basin, allowed the group to collect from fine clastic sediments a large number of plant-bearing samples, related by the accompanying paleobotany specialists to a presumed "Autunian interval".

Later, after a long trip, the group reached Alghero, on the northwestern coast.

18 September (Speakers: Cortesogno, Gaggero, Fontana, Neri, Oggiano)

The Permian-Triassic succession of Cala Viola-Porto Ferro was the main topic. However, an intense downpour forced the organisers to make the first presentation on the local sections in a bar. Once the sun returned, the Torre del Porticciolo (Stop 1) and the Cala Viola (Stop 2) successions were examined in the field (Fig. 4). They are generally represented by red clastics. In particular, two units are recognisable. The lower one is made up of sandstones and subordinate pelites, both rich in well defined sedimentary structures (trough cross-bedding, fluvial channels, bioturbation, etc.). The onset of the overlying unit is marked by a disconformity. The lowermost part consists of a well-sorted quartz conglomerate, 5-7 m thick, passing upwards to more sandstones and minor pelites. Plant remains, ascribed to *Equisetum mougeotii*, occur locally. At Cala Viola, the upper boundary of this unit, which resembles the typical Buntsandstein of Western Europe, is in tectonic contact with Keuper deposits.

Stop 3 was at Torre Nera, on the rocks which, to the north, surround the large beach of Porto Ferro. Reddish coarse sandstones and polygenic conglomerates, along with subordinate finer lithotypes, crop out. Trough cross-bedded sandy or gravel bodies appear mainly amalgamated and thick (> 1 m) bars with step foreset occur. The sedimentary facies indicate an overall braided-river setting.

Fig. 3 – Enjoying a "light" Sardinian lunch near the village of Seui, Barbagia, central Sardinia.

Fig. 4 – Taking a break in front of the Porticciolo hillock, Nurra, NW Sardinia.

This tract of section (about 50 m thick) lies below the previous Torre del Porticciolo and Cala Viola successions, but the age is as yet unknown. However, as these red beds developed later then the Autunian plant-bearing P.ta Lu Caparoni Fm., which crops out in this sector, their general attribution to late Early Permian and/or slightly younger times may be suggested. Therefore, the described interval corresponds, at least in part, to the "Saxono-Thuringian" of the French authors.

Later, outside the programme and the topic of the meeting, an additional stop was made, in order to illustrate the structural setting of the Monte Santa Giusta area, on the way to the last stop.

Stop 4 highlighted the Permo-Triassic sequence, probably found in other parts of northwestern Sardinia, near Porto Torres. The investigated section led to the recognition of basal rhyolitic volcanic products (ignimbrites and tuffs), heavily weathered and tectonised. Reddish clastic sediments crop out above but, although their alluvial environment is obvious, detailed facies analysis is prevented by poor exposure. On the top, a quartz-conglomerate bank (ca. 2 m thick) occurs (Fig. 4). It is regarded, at least so far, as the lateral equivalent of that confined in the basal Buntsandstein between the Torre del Porticciolo and the Cala Viola successions. The conglomerate bank is overlain by a thin silty-sandy succession and by, a few metres above, the Middle Triassic Muschelkalk. The thickness (about 100 m) of these local red beds records a slightly reduced subsidence rate. According to some authors, it is also noteworthy that the volcanic products are alkaline in nature, and consequently demonstrate the presence of a second anarogenic magmatic cycle in Nurra, as in Corsica, southern France and the Pyrenees.

Later, the participants reached Porto Torres, in order to embark for Genova.

19 September

Following an early morning arrival in Genova, the participants were transported to Brescia for the Conference.

THE CONFERENCE IN BRESCIA (20-22 SEPTEMBER)

20 September

According to the programme, the opening ceremony began at 10.00h. First, speeches were given by the Mayor of Brescia, Prof. Paolo Corsini, and Prof. Mario Vanossi, as Director of the Earth Science Department and Head of the "Alps Group" in Pavia, from where the meeting initiative started. They are reported at the beginning of this volume.

Later, after a welcoming reception, Prof. Peter A. Ziegler from the University of Basel opened the Conference with a general lecture on the "Late Palaeozoic-Early Mesozoic Plate Boundary Reorganization: Collapse of the Variscan Orogen and Opening of Neotethys". This extensive and updated geological framework, from the Late Carboniferous to Middle-Late Triassic times, was greatly appreciated by the participants. Later, there was a guided tour of the Science Museum by its curator Dr. Paolo Schirolli.

In the afternoon, the oral sessions began. The first one was devoted to paleontological, stratigraphical, sedimentological and paleogeographical contributions. It consist-

Fig. 5 – Participants' group photograph outside the Natural Sciences Museum of Brescia.

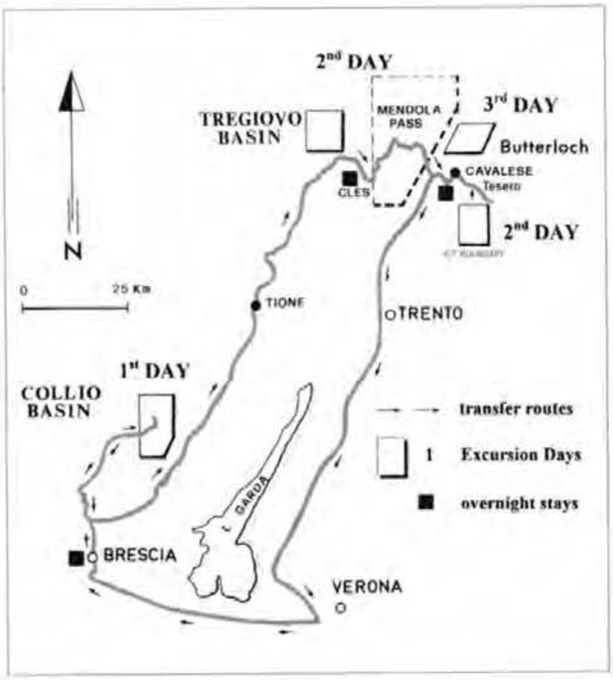

Fig. 6 – Excursion itinerary (23-25 September) and stops on the Permian of the central-eastern Southern Alps.

ed, in this first day, of six oral presentations (H. Kerp, P. Pittau, J. Utting, C. Spinosa, S. Voigt and U. Nicosia). A poster session was drawn up. Independent meetings on differing Permian subjects were also promoted early in the evening.

After dinner, the participants enjoyed an organ concert by Mrs. Eva Frick Galliera in the church of S. Gaetano, in Brescia.

21 September

The previous session (1) continued, and another one (2), on the Permian-Triassic boundary in central to eastern European areas, followed. Both the sessions included 13 oral presentations (H. Haubold, R. Wernenburg, D. Sciunnach, A. Schaefer, J.P. Deroin, C. Virgili, J. Schneider, A. Vozarova, M. Popa, A. Del Bono, E. Malysheva, Z. Liao and S. Radrizzani). A poster session followed. As on the first day, independent meetings on various subjects were held.

A conference dinner, in the picturesque "Porta Bruciata" restaurant, was enjoyed by a part of the group.

22 September

Session 2 and three others, on the Late Carboniferous to Permian volcanism and tectonics of some European and external regions, as well as on the subdivision and discussion of some time-scales, represented the aim of this last day. Nine presentations were given by C. Neri, V. Lozovsky, N. Capuzzo, P. Brack, C. Breitkreuz, J. Feijth, N. Esaulova, M. Menning and B.R. Wardlaw, respectively. A poster session was again drawn up.

The Conference ended with some general reflections on the topics highlighted by three-days of speeches and debates on the continental Permian, thanks to the involvement of the participants (Fig. 5).

A guided tour of the Museum of S. Giulia, in Brescia, concluded the programme.

THE FINAL EXCURSION IN THE CENTRAL-EASTERN SOUTHERN ALPS (23-25 SEPTEMBER)

The itinerary and general stops are indicated in Fig. 6.

23 September (Speakers: Brack, Breitkreuz, Cassinis, Cortesogno, Gaggero, Nicosia)

The day was devoted to examining the continental, sedimentary and volcanic Permian of the Brescian Pre-Alps. The visited outcrops occur between the Upper Trompia Valley and the Giogo della Bala, along the Maniva-Croce Domini road, south of the Tertiary intrusive Adamello massif. The lithostratigraphical succession is among the most notable in the Southern Alps, thanks to the numerous stratigraphical, sedimentological, paleontological and petrographical studies.

The first stop was a panoramic view of the Permian succession from the mountain crest beyond the Refuge Bonardi. The observed section reaches, in the upper Val Dasdana, up to 1500 m in thickness. From the base, the following units generally stand out: rhyolitic ignimbrites, unconformably overlying the Variscan crystalline basement; tuffs interbedded with some alluvial-fan clastic deposits; fluvial to lacustrine variegated and laminated sandy-shaly sediments (lower Collio Fm.); the M. Dasdana volcaniclastic mass-flow deposit, about 15 m thick; turbiditic, fluvial to lacustrine green-brownish coarse to fine clastic sediments (upper Collio Fm.); varicoloured conglomerates and sandstones intercalated with finer sediments which pass laterally into the Collio Fm. (lower Dosso dei Galli Conglomerate); stratified and bioturbated red-brown sandstones and siltstones ("Pietra Simona" Member of the Dosso dei Galli Conglomerate); coarse-grained reddish conglomerates, including metamorphic basement and volcanic rock-fragments, which upwards and laterally, towards the boundaries of the basin, progressively substitute part of the underlying Permian units; and massive rhyolitic-rhyodacitic ignimbrites ("Auccia Volcanics"). The lower volcanic and sedimentary succession (first Cycle or Cycle 1) is unconformably covered by the fluvial red beds of the Verrucano Lombardo Fm., which marks the beginning of Cycle 2.

Stop 2 allowed the participants to see in detail, along the Maniva-Croce Domini road, the boundary between the Variscan metamorphic substrate and the basal volcaniclastic unit (over 50 m thick) of the Permian succession. This boundary spans a gap of as-yet-unknown duration, which however, from local radiometric dates (339±8 Ma, Del Moro, pers. comm. and 283±1, Shaltegger & Brack, 1999, respectively), could be evaluated as about 60 Ma. The contact between the weathered basement and the overlying unconformable volcanics is locally affected by a slight Alpine dislocation. Furthermore, well-bedded varicoloured pyroclastic products (about 30-90 m thick), which include some alluvial bodies, come into tectonic contact with the underlying prominent ignimbrites. A brick-red bed with accretionary lapilli occurs at the top of this section. Immediately above, a conglomerate band (10-13 m), interpreted as a distal fan, marks the beginning of the Collio Basin. The subsequent, typical Collio beds (*Auct.*) are made up of laminated sandstones and shales, up to about 200 m thick, progressively green, reddish and black in colour. Researchers from Genova University gave explanations related to the aforementioned basal volcanics.

Generally, these Collio sediments mark a cycle (from distal alluvial fan to sandflat, mudflat and lacustrine environments) towards deeper water conditions. However, the presence of ripple-marks, mud-cracks, raindrops, and so on, as well as of tetrapod footprints indicate that the basin was never very deep and was often exposed. Carbonate lenses and nodules also occur, especially concentrated in the lower and middle parts of the section.

Stop 3 was along the watershed between the Trompia and Caffaro Valleys, at the sharp Maniva-C. Domini road-bend (2100 m a.l.m). The panoramic view enabled the group to examine in detail some units also seen from Stop 1, in particular the upper dark band of the lower Collio Fm., the M. Dasdana volcaniclastic beds, and the lower member of the Dosso dei Galli Conglomerate.

Fig. 7 – Taking shelter in the Val Trompia Basin, Mt. Dasdana, Brescian Prealps.

Specialists from Rome University presented a poster, showing the stratigraphic position and the names of the tetrapod footprints found in the Val Trompia-Val Caffaro area, from the lowermost Collio up to the first levels of the "Pietra Simona" Member of the Dosso dei Galli Conglomerate.

The stop also allowed the participants to note the local dark shales, traditionally rich in fossil plants and ichnofaunas, cropping out on the side of the road (Fig. 7). The included layers and nodules, which show yellow-orange in colour due to weathering, consist of spherulitic danburite, ankeritic carbonates with minor dravite and trace gold. These mineralisations are interpreted as precipitation from hydrothermal activity associated with the emplacement of the overlying volcaniclastic mass-flow deposits ("Dasdana Beds").

These deposits were the topic of nearby Stop 4. Recent investigations carried out by German geologists, in partial collaboration with Italian researchers, led to the interpretation of this and other subsequent key-beds as the result of a dome activity in the eastern Collio Basin. At Mount Dasdana, they consist of (1) a lower crystal-rich gravelly greyish sub-unit and (2) well-bedded, green, sandy-to-pelitic turbidites. In the former, black pelites (lacustrine Collio rip-up clasts), and various volcanic and metamorphic porphyritic SiO_2-rich lava fragments are included. According to the above authors, a presumed sublacustrine/subaerial eruption column formed as a consequence of the lava dome fragmentation, from which, in the first instance, dense crystal-rich mass flows originated. Much of the foamy lava fragments remained in the column and sedimented later in a second phase from dilute turbidity currents together with sandy-pelitic deposits, and from fall-out.

Stop 5 was at the upper Dosso dei Galli Conglomerate, on the western side of the type-locality. The unit, which is characterised by the inclusion of large Variscan metamorphic rock fragments and Permian volcanics, can be interpreted as a result of debris flows, in alluvial fan deposits.

Stop 6 was located at the topmost part of this "Conglomerate", near the boundary (not exposed along the road) with the overlying Auccia volcanics, reaching a maximum thickness of 130-140 m.

Stop 7 allowed the participants to examine the transition from the aforementioned volcanics to the Verrucano Lombardo, *i.e.* the contact between Cycles 1 and 2 of the Permian. The Auccia volcanics, which consist of massive violet rhyolitic/rhyodacitic ignimbrites, calcalkaline in composition, are subjected at the top to intense weathering and erosion processes. Therefore, a paleosol originated locally.

The Verrucano fluvial red clastics, which initially correspond to sandy braided-stream deposits, unconformably cover the Collio Basin area, and step down outside on to the Variscan crystalline basement. On the whole, the unit ranges from approximately 200 m to 500 m in thickness, and could be related to the onset of a new geodynamic regime.

Logistical and time problems forced the organisers to stop at the Giogo della Bala. A long trip, through the Trompia-Sabbia-Rendena-Meledrio-Sole valleys, led the participants to reach Cles in Val di Non (Trento region), late in the evening.

24 September (Speakers: Bargossi, Neri, Nicora, Radrizzani)

The Tregiovo Basin area was the aim of the early part of this geological excursion. Some researchers from Bologna University, the CARG project and the Geological Office of the autonomous province of Bolzano, who are involved in publication of the new local map at a scale of 1:50.000, illustrated the stratigraphy and petrography of some volcanic bodies underlying the lacustrine Tregiovo Fm., within the so-called Monte Luco stratigraphic sequence. After some brief stops, another followed near the base of the aforementioned sedimentary unit. The outcrops on the side of the road to Lauregno allowed the participants to collect a number of plant-bearing samples from the local fine blackish pelites.

The importance of the Tregiovo Fm. is due to the fossiliferous content, and consequently some knowledge of the presumed age. Investigations on the macroflora, microflora and tetrapod footprints have recently led researchers to relate the unit to the late Early Permian (Kungurian) and early Late Permian (Ufimian, perhaps up to Kazanian) times. However, this attribution deserves further paleontological and radiometric research, and correlation, for a general agreement. As the Tregiovo Fm. is overlain by the last volcanic products ("upper rhyolitic ignimbrites"), the age may approximate to the end of the volcanic activity in this sector of the Southern Alps; moreover, the chronostratigraphical classification would lead us to refine the boundary between the Permian Cycles 1 and 2, and indirectly to evaluate the time gap better.

The subsequent stop was near the Mendola Pass. The participants appreciated a panoramic view of the Athesian volcanics and, towards the east, the magnificent Dolomite Alps. A small guide for this stop was distributed.

The final stop of the day was on the well-exposed Tesero section, in Val di Fiemme (Western Dolomites), which crops out along a road near the village (Fig. 8). During the second half of the last century, this section became famous worldwide after the discovery of Permian-type, unworked brachiopods and foraminifers (= "mixed fauna") located about 1.5-2.0 m above the marine Bellerophon-Werfen lithostrati-

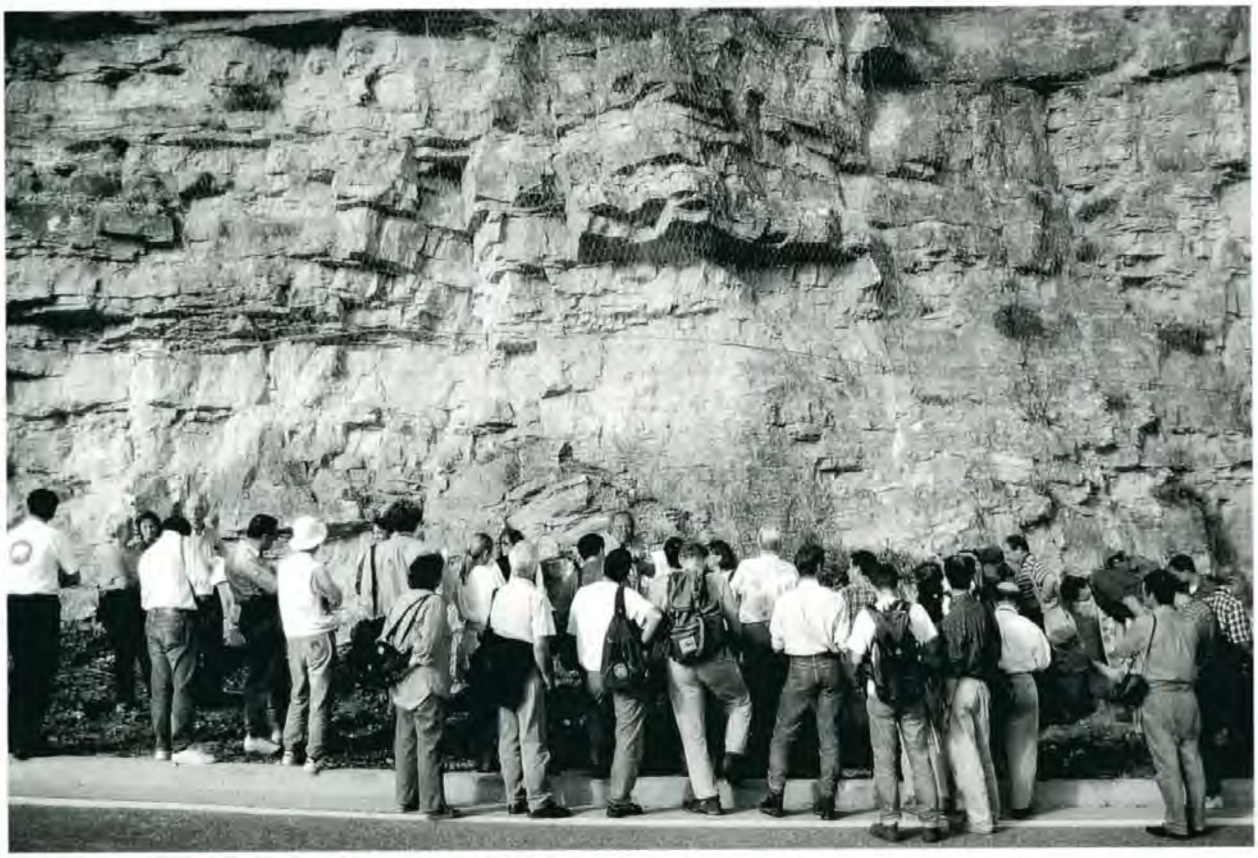

Fig. 8 - Participants observing the P/T boundary along the famous Tesero section, western Dolomites.

Fig. 9 – At the Bletterbach-Butterbloch waterfall, western Dolomites.

graphic boundary. This finding allowed a re-consideration of the position of the P/T boundary, traditionally placed at the base of the Tesero Member of the Werfen Fm. Paleontological, sedimentological and geochemical studies were intensively performed by Italian and foreign researchers. Recently, the discovery of conodont faunas led to a refinement of the debated chronological marker. The co-occurrence (35-40 cm above the onset of the Tesero Member) of *Hindeodus praeparvus* (Morphotypes M1 and M2), *Hi.* sp. *A* and a few ramiforms indicates the *praeparvus* Zone, considered the base of the Triassic by Orchard & Krystyn (1998). Upward (about 1.3 m above), *Hi. changxingiensis* also occurs along with *Hi. praeparvus*. The next conodont fauna is found in the lower Mazzin Member, 11 m above the base of the Werfen Fm.; it is characterised by *Hindeodus parvus parvus* associated with *Hi. praeparvus* M1. The former species represents the first conodont biozone of the Induan. In the Tesero section the FO of such a fossil is at least 8 m above the disappearance of the Permian-like components of the already recorded "mixed fauna".

Later, in the evening, the participants reached Cavalese, in Val di Fiemme.

25 September (Fontana, Massari, Kerp, Nicosia, Pittau)

The last day of this post-Congress excursion was once again dedicated to visiting the Bletterbach Gorge near Redagno (Radein), in Bolzano province. The local section has long been known for its spectacular outcrops and abundance of plant remains and tetrapod footprints, repeatedly highlighted in a very large number of old and current studies. The Upper Permian succession, which was the focus of attention, displays an overall transgressive trend, and may be divided into a number of depositional sequences (the first five and the lower part of a sixth sequence have been identified). It is made up of two sedimentary units represented, from base to top, by the continental fluvial red beds of the Val Gardena Sandstone, and by the evaporites to marine sediments of the Bellerophon Fm. Upwards the succession is followed by the Lower Triassic Werfen Fm., which initially includes the so-called "Tesero Horizon" *Auct.*

Four stops were made climbing the Bletterbach Gorge. They generally included, at different heights, stratigraphical, sedimentological, paleontological (on the fossil macroflora, microflora and the tetrapod footprints) and petrographical explanations. The first stop was at the famous Bletterbach waterfall (Fig. 9), the second one slightly upstream, the third at the most important ichnological site of the section, and the last stop at the head of the valley, where the upper part of the Permian succession and the overlying Lower Triassic and Anisian units are spectacularly exposed.

The excursion closed in Redagno, with a short visit to the new, small Natural Science Museum, which includes some geological drawings and samples of this area of the South-Alpine Permian, along with a number of fossils, mainly represented by tetrapod footprints and vegetal remains. Later, on behalf of the local community, Mr. Sepp Perwanger welcomed the participants. He also stressed that the geology of the investigated Bletterbach area has always attracted the interest of many Italian and foreign researchers. The organisers of the meeting thanked Mr. Perwanger and the other representatives for their warm hospitality. Finally, after eleven eventful days of enlightening debate and wonderful scenery, in a very friendly atmosphere, a hearty toast concluded the Congress.

ACKNOWLEDGEMENTS

The Brescia Conference and the related field trips to Sardinia and the central-eastern Southern Alps could not have taken place without the generous help of certain institutions, to which we are indebted:

- the Council of Brescia, in particular the Mayor, Prof. Paolo Corsini, and his Vice, Avv. Giuseppe Onofri, for their financial support in organising this meeting;

- the Civic Museum of Natural Sciences, which hosted the Congress; the Director, Prof. Marco Tonon, along with the Earth Sciences Curator, Dr. Paolo Schirolli, who overcame organisational difficulties, and ensured the publication of these Proceedings in a special volume ("Monograph") of the Museum;

- the Ministry of University and Scientific and Technological Research (MURST), Co-fin. 1998, for the financial support of a two-year study programme on the Late Paleozoic stratigraphical and structural evolution of the Alpine and Apennine sectors, compared with Western Mediterranean areas;

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- the Earth Science Departments of the Universities of Bologna, Cagliari, Ferrara, Genova, Milano, Modena, Padova, Pavia, Perugia and Roma, along with the Sardinian Region, the PROGEMISA S.p.A. of Cagliari, the CARG Project and the Geological Office of the autonomous province of Bolzano, for their active contribution in the preparation of this meeting. Furthermore, we sincerely thank Jean Broutin (Paris), Hans Kerp (Münster), Slavcho Yanev (Sofia), Christoph Breitkreuz (Potsdam) and Peter Brack (Zürich) for their help. In particular, we are very grateful to Prof. Peter A. Ziegler (Basel), who gave the opening lecture, and also Prof. Gérard M. Stampfli (Lausanne) for his co-operation.

- The IUGS Subcommission on Permian Stratigraphy (SPS) and the Italian Geological Society (SGI). The former was represented at the Congress by the Chair, Dr. Bruce R. Wardlaw, and the Secretary, Prof. Claude Spinosa; the latter, which sponsored the meeting, was represented by the *ad interim* President, Prof. Francesco A. Decandia, and a Board Member, Prof. Mario Vanossi, who is also the Co-ordinator of the "Gruppo Alpi" at Pavia University.

We would particularly like to mention: Sergio Ambrogio, Ausonio Ronchi, Giuseppe Santi and Paolo Schirolli for their hardwork and enthusiasm in the successful preparation and realisation of the meeting.

Samantha Jones and Pamela Sutherland for the revision of the English texts, Kathy Riklin for administration, and Irene Bresciani in organising the participants' programme in Brescia, are acknowledged. We are also grateful to Eva Frick Galliera for the evening organ concert in the S. Gaetano Church.

Finally, our thanks go to all those who took part in the excursions and the scientific sessions in Brescia. Their cordiality, the stimulating discussions on the Upper Paleozoic to Lower Triassic successions, and the many impressive aspects of the landscapes constantly enlivened the atmosphere of the Field Conference, and will certainly arouse pleasant memories.

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