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Newsletter 3

IGCP Project 276

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COVER: Paleozoic formations of the Little Kabyle Massif in Algeria visited by the participants of the 3rd fieldmeeting of the IGCP Project No 276 (September, 1990)

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Litotectonic unit of correlation in the internal zones of Maghrebides; the example of Tahrict unit (Ouled Asker), Western Little Kabylia, Algeria

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The massif of Little Kabylia is one of internal elements in the southern chain of the Western Mediterranean. The basement unit of this massif is essentially composed of basement gneisses followed by a schistous and weakly metamorphosed unit. Between both these large units a formation called alternations was distinguishable at the summit of which a mineralized horizon metamorphosed in grades limited by the greenschist and amphibolite facies (Afalfiz, 1990).

For certain authors the structure of the Little Kabylian massif started in the Variscan with a blastomylonitization followed in the Alpine period by decompression and a distensive phase. For others the structural evolution of the massif is older (Precambrian) with a granulitization followed by mylonitization in the Variscan and later in the Alpine stage (Djellit, 1987).

Litotectonic profile of the Tahrict unit

Only elements of the schistous assemblage and some parts of the alternations crop out in the area of Tahrict (Fig. 1). Investigations of this unit allowed to distinguish the following formations from the base to the top (Fig. 2).

Formation of alternations

The thickness is variable (60 - 80 m) and the formation clearly occurs at the base of the schistous assemblage. At the base, it rests with a tectonic contact, over the gneissic complex of the basement of Little Kabylia. The formation displays portions with very differing character from which the most important ones are the mica-schists with biotite, garnet and some felspar, the quartzites, the cippolinos associated with small bodies of actinoliths, and the graphitic fine-grained mica-schists. The formation terminates upwards by a carbonate layer with intercalations of amphibolitic rocks in which a base-metal ore is localized. This level of mineralized marble is situated (at the summit of the formation of alternations (Fig. 2) and was the target of a detailed geological and metallogenetic study (Afalfiz, 1990).

From the results of the geological investigations of the Tahrict unit and namely of the mineralized body present, as well as in the results of paragenetic and geochemical observations, it is possible to assume the principal features of the base-metal ores. The comparison of these main features with similar worldwide deposits, allows (the classification) of the type of ore as hydrothermal-synsedimentary and metamorphosed.

Data on rare element concentrations analyzed from the mineralized carbonate beds gave similar profiles to the mineralized bodies at Usua in West Greenland (banded iron ore formation; Appel, 1980). According to this author, this type of ore is common for Precambrian formations.

Isotopic data of lead from galenite in this ore, according to Touahri (1987) shows the average evolution curves indicating a Lower Paleozoic age. It is possible to show such parts where these concentrations have not escaped the eventual recrystallizations during tectono-metamorphic events which the ores underwent later.

Formation of biotite schists

This formation overlies, without a tectonic discontinuity, the formation of alternations. Towards the summit, the formation is terminated by a key horizon of porphyroids. A lenticular layer of marble is associated with the porphyroids.

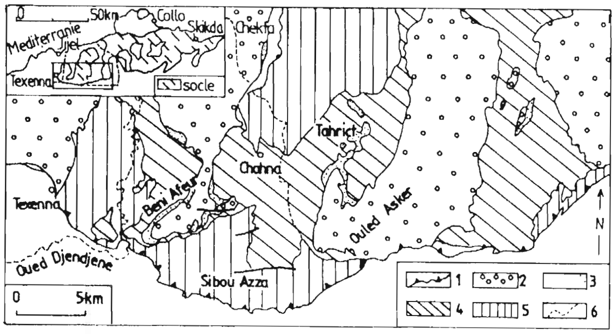

Fig. 1. Localisation of outcrops of dated or supposed Paleozoic formations in the west of Little Kabylia. 1 - front of the Alpine overthrust over the basement of Little Kabylia, 2 - sedimentary formations and nappes of Cenozoic age, 3 - dated or supposed Lower Paleozoic, 4 - phyllites (the schistous assemblage), 5 - gneisses and associated rocks (the gneissic assemblage), 6 - hydrographic network (after Ehrmann, 1928; Durand-Delga, 1955; Bouillin, 1981).

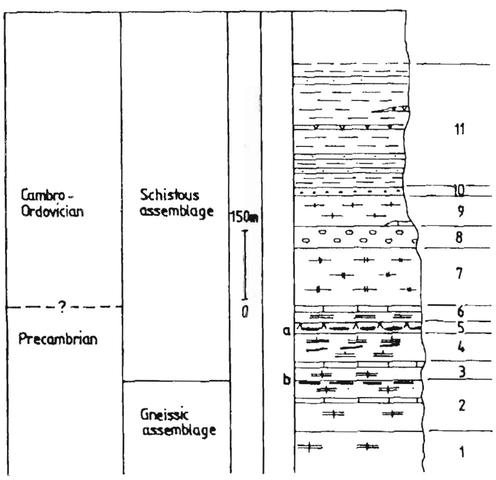

Fig. 2. Litotectonic unit of Tahrict. a - cataclastic contact, b - dductile contact; 1- fine grained gneisses, 2 - fine sandstone with marble intercalations, 3 - felspar-bearing mica-schist + marble, 4 - graphitic mica-schist, 5 - base-metal ore (Fe, Pb, Zn, Ba), 6 - mica-schist and marble , (3-6 - Formation of alternations), 7 - biotite schists, 8 - key horizon of porphyroids, 9 - muskovite - chlorite schists, 10 - key horizon of quartzite, 11 - sericite-chlorite schists with intercalations of dolerite, sandstone, arcose and quartzite.

Formation of muscovite-sericite-chlorite schists

This formation follows with normal contact over the biotite schist formation. Upwards the formation is divided by a quartzite layer of 10 m thickness. This quartzitic element creates a second key horizon in the Little Kabylia massif.

Formation of chlorite-sericite schists

This is an epimetamorphic schist with intercalations of quartzitic schists, dolerite, sandstone and arcose. Near the lower boundary of the formation, a gradual transition from quartzite at the base into schistous levels immediately above is clearly visible.

The dated Paleozoic

Paleozoic beds overlying the basement have been dated firstly at Beni Afeur (SE from Jijal) in 1922 by E. Ehrmann. Later the age was determined precisely by Durand-Delga (1955) and Alberti (1980). This Paleozoic is represented by arenaceous and pelitic levels of Silurian and Lower Devonian age.

Beneath these Paleozoic levels, Bouillin et al. (1981) mapped in the area of Tahrict (Ouled Asker) and found the fossiliferous assemblages characteristic for Tremadocian. According to these authors the arenaceous levels are resting without tectonic or metamorphic discontinuity over a 200 m thick unit of coarse sandstone and arcose with volcanodetritic intercalations.

Actually, this layer (200 m) represents, according to our investigations, the uppermost part of the schistous assemblage of the Little Kabylian basement. This summital part is composed of quartzite, sandstone, arcose and porphyroids intercalated in the formation of muscovite-chlorite and sericite schists.

Litotectonic consequences

The existence of arenaceous levels of Tremadocian age in the Tahrict area passing downward without any discontinuity into the schistous assemblage indicates the Cambro-Ordovician age of the upermost part in the Tahrict unit. What concerns the basal part, where the base-metal ore occurs, the age may represent the Precambrian/Cambrian boundary, according to the data on rare elements and from the isotopic concentrations in the base-metal ore.

Conclusion

The litotectonic unit of Tahrict has a Lower Paleozoic age and created by rare key horizons, it represents a correlative element in the Maghrebide chain. In the set of elements, at least three levels are distinguishable, namely:

- the layer of quartzites situated at the base of sericite-chlorite schists,

- the layer of porphyroids associated with a lenticular layer of marble always forming the base of muscovite-chlorite schists,

- the carbonatic level with intercalations of tuffites with magnetite, baryte and the iron, copper, lead, zinc and silver ores. The mineralized horizon occupies the summit of the formation of alternations which compose the base of the schistous unit of Tahrict.

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Present knowledge of tectogenesis of Veporicum (West Carpathians)

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Abstract

The variegated structure of Veporicum and its complex structures may be explained by the combined effect of two long orogenic stages - the Hercynian and the Alpine, including the incorpotation of pre-Hercynian fragments in the structure. The reconstruction of the Hercynian fragments is best throvngh gradual elimination of the effect of each successive superimposed tectonic process (paleo-alpine, neo-alpine). The author offers brief information about the present approach to the results of all tectonic processes.

The author informs briefly on the present knowledge of the tectogenesis of the Veporicum, based on the latest authors' s mapping reflected in the enclosed tectonic scheme. The results of the mapping and of the structural research enable new interpretations of kinematics and succession of tectonic processes in the Veporicum. The article only contains general conclusions and the respective factographic material may be found in quoted publications.

The structure of the Veporicum - an alpine tectonic unit-comprises several crystalline complexes, pre-Late Carboniferous metamorphic rocks and granitoids, as well as their Upper Paleozoic - Mesozoic cover (including displaced Mesozoic nappes).

The results of recent investigations enable the division of the metamorphic rocks of Veporicum into several complexes of different lithology, age and degree of metamorphism and granitoids. Prominent differences in composition and structure between the basic longitudinal (NE) zones of the Veporicum in the sense of Zoubek (1957) are due to combined effects of several tectonic alpine and pre-alpine stages (Plašienka, 1983; Bezák, 1988). We shall now deal with the analysis of present knowledge of tectonic situation in the part of the Veporicum, building the western part of the Slovenské rudohorie Mts. (Fig. 1).

In the northern part (the Krakľová zone) the high grade metamorphic rock complexes predominate (various types of gneisses, partly migmatitized, strongly diaphthorized into mica schists and phyllonites). Low grade metamorphic rocks of Lower Paleozoic age are, in some places, folded within the metamorphites. Granitoid rocks are mainly in the central (Kráľova hoľa zone) and eastern parts of the Kohút zone. Granitoids of basic types (the Sihla type, hybrid, porfyric and leucocrate granitoid) are in various proportions present in isolated segments, separated by NE and NW faults. The tectonic scheme (Fig. 1) shows that the composition of the southern zone (Kohút) is most variegated, comprising granitoids and several metamorphic complexes of Lower Carboniferous-Lower Paleozoic-Precambrian (?).

The tectonic superposition of the major part of the granitoid-migmatite complexes over the metamorphites (Klinec, 1966) and the assignment of some low grade metamorphic complexes to the Lower Paleozoic (Klinec, Planderová & Miko, 1975) are most significant results for the structure of Veporicum. Futher investigations, based on geological mapping, structural-tectonic analysis, petrology of metamorphic rocks, palynology, geochronology and geophysics concluded that a lot of elements in the structure of the Veporicum cannot be explained only as the result of alpine tectonic processes. The remains of pre-Alpine structures are locally preserved in Alpine tectonic units or are strongly modified by alpine structures (Siegl, 1982; Bezák, 1988). Strike-slip faults seem much more significant, but their duration and extent are discussed (Pospíšil et al., 1989).

When analysing the tectonic development, we should start from the latest movements along faults

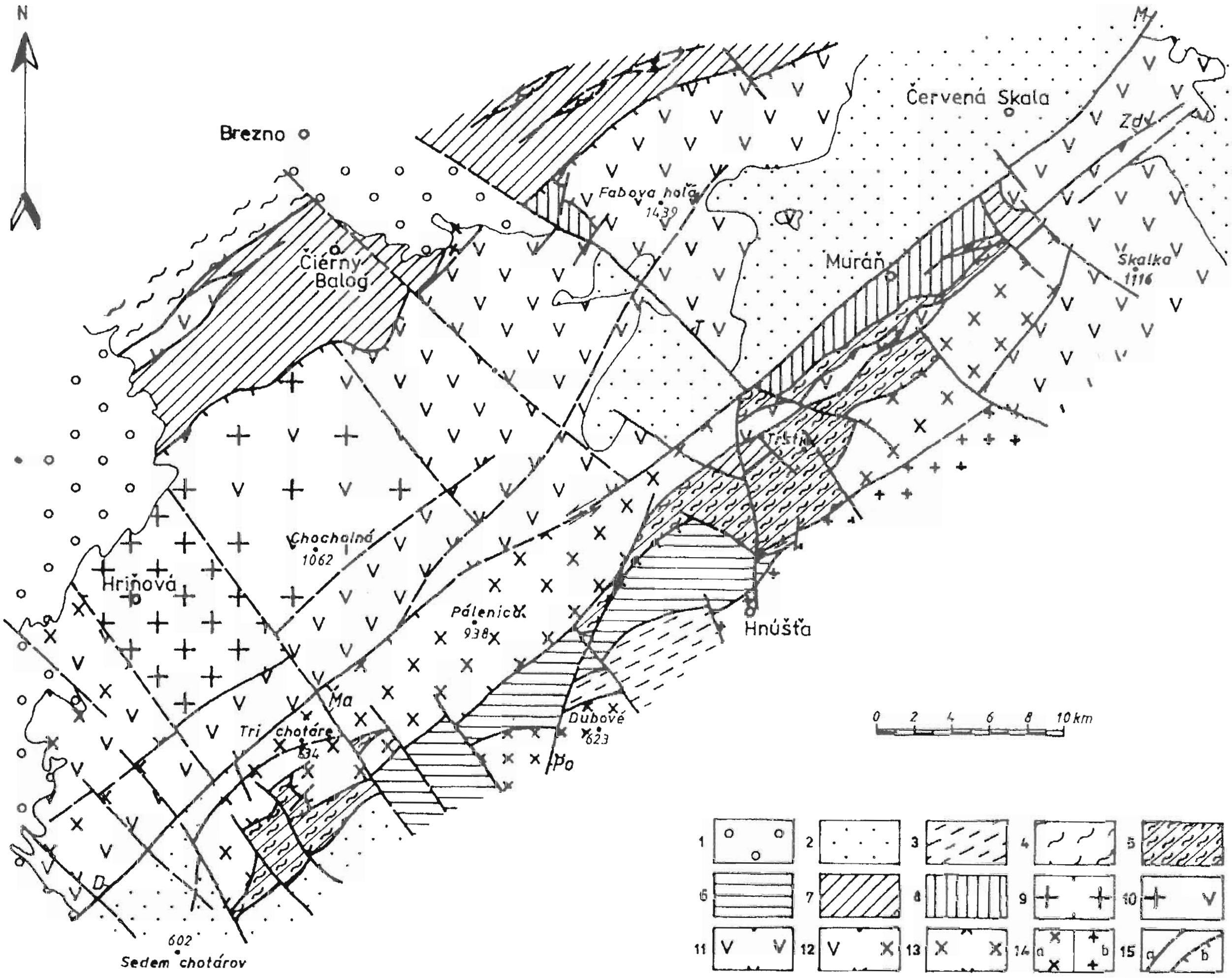

Fig. 1. Tectonic scheme of western part of the Slovenské rudohoric Mts. 1 - Tertiary; 2 - Upper Paleozoic and Mesozoic, undivided (including the Gemeric and Silica nappe); 3 - schists containing magnesites (Sinec complex, Lower Carboniferous ?); 4 - diaphthorized gneisses and phyllonites, including folded phyllites; 5 - garnet mica schists (Ostrá complex, Lower Paleozoic); 6 - albitized gneisses (Klenovec complex, Lower Paleozoic); 7 - gneisses containing metabasic intercalations, partly migmatitized (lower Paleozoic-Proterozoic ?); 8 - light-coloured quartz-feldspar gneisses (Muráň orthogneisses) including metabasic intercalations; 9 - granitoids of Sihla type; 10 - alternation of Sihla- and porphyric types of granitoids; 11 - dominant porphyric granitoids; 12 - hybrid (?) granitoids containing porphyric granitoid bodies; 13 - hybrid granitoids; 14 - leucocrate granitoids in southern belt a) - two mica granites, b) - biotite granodiorites-tonalites; 15 - tectonic lines - a) mostly steep faults with predominant combined movements, - b) remains of overthrust structures, undivided (of all generations). M - Muráň f., Zd - Zdychava f., D - Divín f., T - Tisovec f., Po - Poltár f., Ma - Málinec f.

associated with the origin of Neogene basins. These are mostly vertical movements on NW faults whose function in some cases (e. g. the subsidence of the Brezno basin and of the block with the Mesozoic W of the Tisovec fault) may be controlled by strike-slip, mainly upon the Muráň system. Some Neogene faults (e. g. Malinec) extending over crystalline complexes separate various complexes and so their pre-disposition may be older. A comparison between complexes of different ages shows that oscillation proceeded on some complexes (e. g. the subsidence of the Mesozoic and the uplift of mica schists on the Divin fault).

The system of the steep NE faults in the southern part is the most significant structure (Fig. 1). It mainly concerns the Muráň tectonic system consisting of several genetically related and mutually linked faults which make it difficult to denote and define each of them, as for example, the problem concerning the SW extension of the classical segment of the Muráň line (Bezák, 1980). Its continuation to the SW of Tisovec was placed on the conspicuous boundary of granitoids and metamorphites. From the present viewpoint the following most prominent branches of the Muráň system may be defined: the classical segment of the Muráň fault NE of Tisovec, the Divín fault in the segment Tisovec-Divín, a parallel fault N of the Divín fault, dividing granitoids of various types and extending to the northern margin of the Muráň Mesozoic, the Zdychava system S of the Muráň fault system and their possible extension SW of Tisovec on the contact of granitoids and metamorphites S of the Divín fault to the southern margin of the Tuhár Mesozoic as well as the fault zones, bordering on the S the mica schists of the Ostrá complex, and combined with the NNE - NS lines (the Poltár fault). The thron of individual faults is not great but only the total thron is recorded. The thron of geological markers, is a problem although in the future perhaps some complexes may be taken into consideration (orthogneisses with basic rocks near Muráň and Pohronská Polhora, porphyric granitoids, mica schists below and above granitoids). The strike-slip faults modifying the paleoalpine overthrust structure were evidently active in the northern part as well (Hók & Hraško, 1990). Generally, the Kinematic picture is a system of southern blocks moving from the southwest to the northeast through a series of parallel and subparallel faults. The faults mechanism probably affects the entire Carpathians and the formation of the Carpathian arc. Fragments of the Cretaceous orogeny should be approached also from this viewpoint because they may have been in a quite different position in the past.

The N-vergent overthrust tectonics was dominant in the paleoalpine stage. Further compression resulted in reversefault zones, sometimes superimposed on the overthrust planes (Plašienka, 1983). With respect to the character of the Mesozoic cover the convergence of the blocks bearing the Mesozoic of the South-Veporic and North-Veporic type is the most significant tectonic convergence in the Veporicum. The area of proximation may be identical with the zone where granitoids overthrust upon the North-Veporic metamorphites. The overthrust structures are evident both W and E of the transversal Tisovec fault. They have also been indicated geophysically (Tomek et al., 1989). Yet the zone is not identical with the steep Pohorelá fault in the sense of Zoubek (1957) or with the Rimava overthrust in the sense of Klinec (1971). With respect to the existence of two significant tectonic units in Veporicum, it is interesting that there are also lithological and metamorphic differences in the North-Veporic and South-Veporic crystalline complexes (different metamorphic development, different type of Lower Paleozoic complexes, presence of Lower Carboniferous with magnesites).

The most part faults affecting the granitoids may be attributed to the Hercynian period. Some data (position in a half-window near Cinobaňa, decreasing magnetic anomalies, to the north locally preserved pre-alpine structures) indicate a reverse-southward vergency. Metamorphic differences in crystalline complexes with some Mesozoic cover, mylonite structures fixed by granitization, extensive deformation of porphyric granitoids under the conditions of higher P-T than in the Mesozoic cover, NW-structures in the granitized complex a. o. are the most prominent traces of the Hercynian tectonics. The structure of the Hercynian orogeny was different and at present its fragments are incorporated in new Alpine tectonic units.

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Carboniferous preflysch sediments in the Alpine-Mediterranean belts

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Abstract

In most parts of the Circummediterranean the Variscan preflysch tectofacies terminates within the uppermost Tournaisian - late Viséan. Only within the Paleozoic of Graz and the Bükk-unit do autochthonous preflysch sediments reach the Namurian/?Westfalian A. Within the Carboniferous preflysch tectoenvironment three distinct facies can be recognized:

1. ± continuous pelagic facies of pelites, lydites (sometimes with phosphorites) and nodular limestones of very little thickness: Western-Mediterranean; and parts of the Graz Paleozoic, Bükk-unit and Southern Alps; Calabria/Sicily; W Kraistids; Istanbul Paleozoic.

2. Sequences of pelagic limestones dominated by erosional gaps (subaerial or submarine) at the Devonian/Carboniferous boundary: Parts of Pyrenees, Eastern - and Southern Alps; Szendrö-Mts.

3. Shallow water preflysch sediments in autochthonous occurrences, only known from the Eastern and Southern Carpathians. But all over the area allochthonous blocks of shallow water materials (Lower Carboniferous - Bashkirian) are included within the flysch superposing the preflysch sediments.

Introduction

Circummediterranean Carboniferous sediments reflect the intensity of the Variscan orogeny in different tectofacies (preflysch - flysch - molasse - foredeep - stable shelf) which varies in sedimentary facies, regional distribution, and stratigraphic age.

During the last decade, a systematic documentation of the prealpidic Circummediterranean realms was performed within IGCP Project No. 5 and published in IGCP No. 5 Newsletters Vol. 1-7. This research provides a good data base for paleogeographic restorations, which should be carried out within the scope of IGCP Project No. 276.

The considered area covers from the West Mediterranean area (Alboran Balearic Trough with Internal Rif, Kabylean Massif, Menorca, Betic Cordillera; Catalonian Coastal Ranges, Pyrenees, Massif of Mouthoumet and Mt. Noire) to the Eastern- and Southern Alps, Sardinia, Tuscany, Calabria/Sicily, the Hungarian Paleozoic of the Bükk unit (Szendrö-, Uppony- and Bükk-Mts.), Carpathians, Dinarids, Moesian Platform, Dobrogea, Hellenids, Pontids, Taurids and the Greater Caucasus.

Papers covering the total Carboniferous of this region are lacking. However a few regional or thematic summary papers are available, in press, or in preparation (Ebner, 1978; Vai, in press; Ebner, 1991 , Ebner et al., 1991).

The presented paper is a rough record of those sedimentary units which take in a proved preflysch setting in accordance with their continuous superposition by flysch sediments. Variscan flysch above Carboniferous pelagic preflysch sediments is only missing in the Noric Nappe of the Grauwacken-zone and the Paleozoic of Graz.

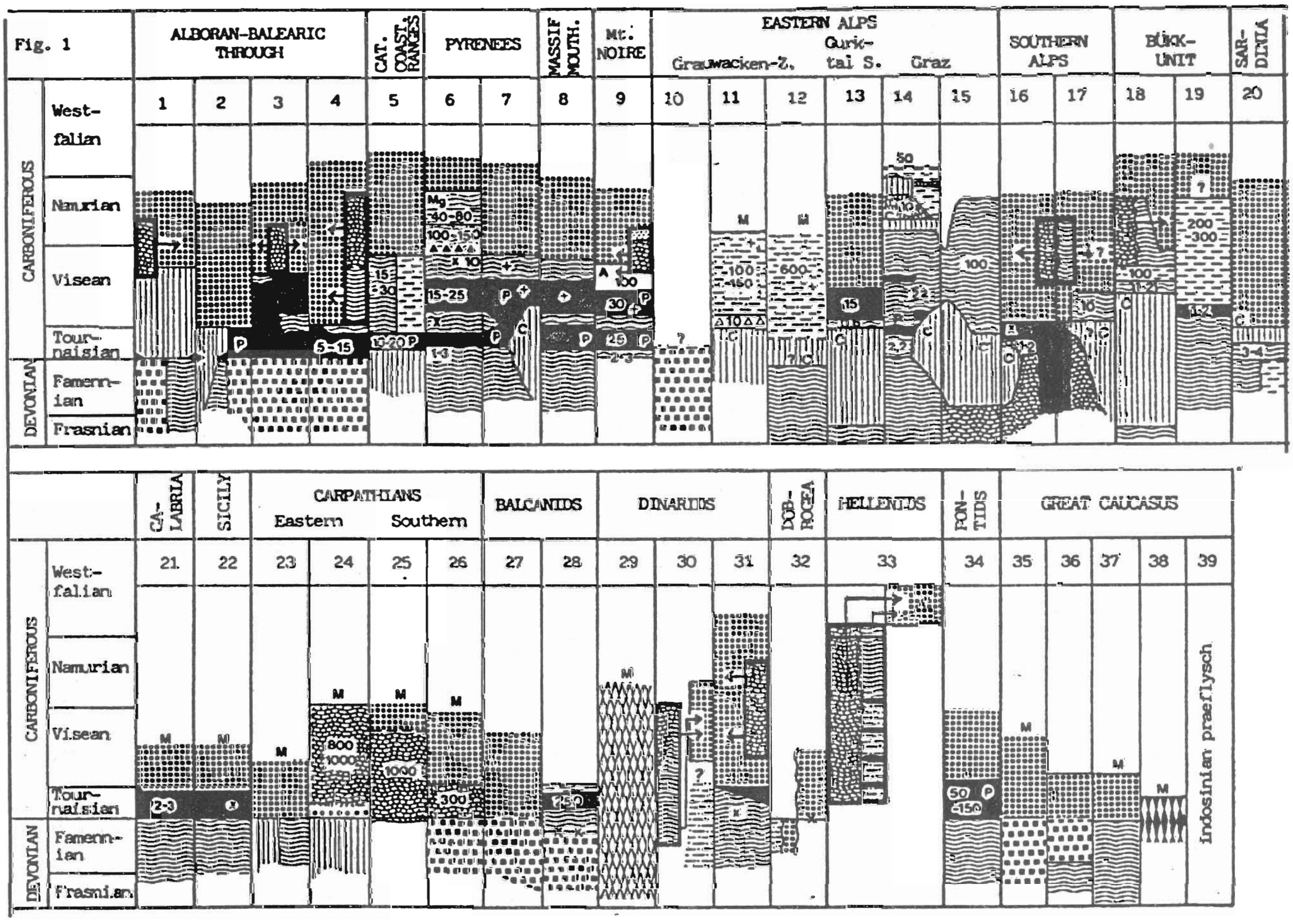
Other Carboniferous tectofacies

The climax of the Variscan orogeny within the Carboniferous together with regional differences in the orogenetic intensity gave rise to a complicated pattern of Carboniferous tectofacies which was heavily disturbed by later alpidic movements. Besides the described preflysch environments, the following other tectofacies occur within the Circummediterranean Carboniferous:

Flysch environments

Ebner (1990) recognized the following flysch stages:

- Presudetic flysch within the Eastern Alps (Gurktal Thrust System, ? Western Grauwacken-zone), Balcanids, Eastern and Southern Carpathians, Dobrogea/Macin-unit, Sardinia, Calabria/Sicily, Caucasian Fore Range-zone).



LEGEND to FIG. 1

FLYSCH-TECTOFACIES:
- base of Carboniferous flysch
- top of Devonian flysch

OCEANIC ENVIRONMENTS:
- parts of oceanic environments in general
- ophiolitic sequence

PREFLYSCH-TECTOFACIES
- ? praeflysch: pelitic/psammitic
- pelites
- lydites, dark pelites
- P phosphoritic nodules
- pelagic limestones
- lydite breccias
- limestone breccias
- A allodapic limestones
- Mg magnesite
- \+ volcanic intercalations
- shallow water carbonates
- conglomerates
- terrestrial clastics

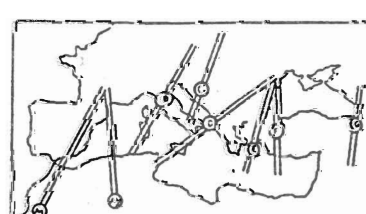
- stratigraphic gaps
- C conodont mixed faunas
- 5 thickness in m
- 1 - 10 m
- M metamorphic sequence (below)
- orogenetic phase
- age and facies of allochthonous blocks
- tectonic boundary

List of documented areas:

1 Internal Rif (Aakaili-Kondiat Tiziane-Beni Hozmar-Talembote nappe)
2 Kabylean Massif (Little K., Great K., Massif of Chenoua)
3 Menorca
4 Betic Cordillera
5 Catalonian Coastal Ranges
6,7 Western and Eastern Pyrenees
8 Massif of Mouthoumet
9 Mt. Noire
10-12 Grauwackenzone
  10 W Zell/See
  11 Noric nappe (Eisenerz area)
  12 Noric nappe (Veitsch area)
13 Gurktal Thrust System (Stolzalpe nappe)
14,15 Paleozoic of Graz
  14 Rannach Group
  15 Hochlantsch Group
16 Carnic Alps
17 Karawanken Mts.
18 Szendrö Mts.
19 Uppony Mts.
20 Sardinia
21 Calabria
22 Sicily
23,24 Eastern Carpathians
  23 Infrabuccovina nappe
  24 Subbuccovina nappe
25,26 Southern Carpathians
  25 Supragethic nappe
  26 Danubian unit
27,28 Balcanids
  27 Stara Planina, E Kraistids, Kucaj-Zone
  28 W-Kraistids
29-31 Dinarids
  29 Vardar Zone
  30 Jadar Paleozoic
  31 Central and SE-Bosnia
32 Dobrogea
33 Hellenids/Chios
34 Pontids/Istanbul Zone
35-37 Great Caucasus
  35 Precaucasian Zone
  36 Fore Range Zone
  37 Main Range Zone
38 Root Zone
39 Southern Slope Zone

POSITION OF CONSIDERED AREAS IN IGCP No. 5 GEOTRAVERSES:

A: loc. 1-9
B. 10-17, 20
C: 18, 19
D: 20, 21, 23-31
E: 32-34
F: 35-39



- Intravisean - Westfalian flysch in the Western Mediterranean, Southern Alps, Jadar-Paleozoic, ? Outer Dinarids.
- Upper Carboniferous - ? Permian flysch in the Hellenids.
- Flysch of the "filling up" - type (=conformable superposition by shallow water sediments) in the Bükk-unit (Namurian - Westfalian), ? Jadar Paleozoic (Lower Carboniferous - Namurian) and Pontids (Middle Visean).
- Flysch is only missing in those areas in which the Variscan orogeny was only weak or non effective (Taurids, Caucasian Southern Slope-zone, Moesian Platform, Predobrogea Downwarp, Macedonia, Velebit) or where shallow water sediments of Veitsch/Nötsch/Ochtina-type were deposited.

Real oceanic environments

These are only tentatively assumed for the Vardar Trough (Grubic & Ercegavac, 1974; Krstic et al., 1988) and in the Caucasian Root-zone (Belov et al., 1978; Adamia et al., 1980, 1982).

? Bretonic marine molasse environments

In some areas (Eastern Alps: Grauwackenzone/Veitschnappe, Nötsch; N-Gemeric Dobšiná-Group; Transdanubian Szabadbattyan-Fm., Transcaucasian Shori-Molasse) carbonatic/clastic, marine, shallow water sediments take in a position (? tectonic or sedimentary) above Prevіséan metamorphic rocks.

These sediments are in part interpreted as marine molasse, similar to sediments related to a Previséan orogenetic (? Bretonic) phase.

Continental molasse environments

In the Pyrenees, S-France, parts of the Alps, Tuscany, Sardinia, Eastern and Southern Carpathians, Balcanids, Dobrogea, Great Caucasus.

Marine Upper Carboniferous *molasse* of Auernig-type within the Southern Alps or shallow marine sediments of the "filling up" - type in Bükk-Mts., ? Jadar Paleozoic and the Istanbul-zone of the Pontids.

Fore deep environments:

In Tuscany (Carpineta-Fm., Farma-Fm., St. Antonio-Lmst.; Cocozza et al., 1987) , Velebit Mts. (Ramovš et al., 1984), Predobrogea Downwarp (Belov et al., in prep.), parts of the Moesian platform (Vai in press; Cocozza et al., 1987), N-Pontids/Zonguldak area (Tokai, 1981).

Stable shelf environments in the Taurus area (Demirtasli, 1981).

Preflysch sediments

The Variscan preflysch sedimentation usually reaches the Lower Carboniferous. Only within parts of the Blacanids (Spassov et al., 1978; Spassov, 1983; Maslarevic & Krstic, 1987), the Caucasian Fore Range- and Precaucasian-zone (Adamia et al., 1980, 1982; Belov, 1978) and in most parts of the Variscan belt which were affected by metamorphism, does the preflysch stage terminate within the Precarboniferous or close to the Devonian/Carboniferous boundary.

A special situation is represented by the Alboran-Balearic Trough, where a Devonian flysch is followed by a second Intravisean - Namurian flysch stage (Chalouan, 1987; Buchroithner et al., 1980a, b; Bourrouilh, 1973, 1982; Bouillin & Bourrouilh, 1986; Leprivier & Bourrouilh, 1986; Herbig & Stattegger, 1987; Ebner, 1991).

On the other hand, the youngest typical Carboniferous preflysch sediments reach the Namurian (Paleozoic of Graz/Ebner, 1977, 1978; Bükk-unit/Kovacs & Perc, 1983; Kovacs et al., 1983; Western Pyrenees/Bourrouilh, 1983).

Pelagic environments

The most prominent sedimentary preflysch-facies consists of pelagic lyditic/carbonatic sequences which show the following characteristics:

- Lithological dominance of flaser- (nodular, "griotte") limestones, dark pelites and lydites, often associated with phosphoritic nodules (the last two especially within the Uppermost Tournaisian and Lower Visean).
- The Tournaisian even of continuous environments is always reduced to a few meters.
- Stratigraphic gaps with conodont mixed faunas especially covering the Devonian/Carboniferous boundary, relate to karstification, subaerial and submarine erosion or stratigraphic condensation (Ebner, 1989).

- The pelagic Carboniferous preflysch environments follow mostly Upper Devonian cephalopod- and conodont-bearing flaserkalk (griotte) environments. Only in parts of the Graz Paleozoic (Zier, 1983; Gollner & Zier, 1982, 1985) and the Southern Alps (Tessensohn, 1974, 1975; Bandel, 1972; Pohler, 1982) they do overlie Frasnian shallow water environments.

- The preflysch is generally followed within the Visean by flysch sediments (for exceptions see above).

- Lower Carboniferous volcanics are only known from the Pyrenees, the Massif of Mouthoumet and Mt. Noire ("cinerites").

- Conodonts are the best guide fossils of this facies.

- Conodont mixed faunas are widely spread within this facies and provide an important tool for paleogeographic and geodynamic restorations (Ebner, 1989).

It is worth noting, that not only conodonts can be separated from carbonatic materials. By means of HF it is even possible to extract them from siliceous rocks like lydites (Herzog, 1983).

At the Devonian/Carboniferous boundary there was an important change within some fossil groups (e. g. conodonts, ammonoidees, trilobites).

This, the above mentioned gaps and lithological changes (in the Pyrenees, the Mt. Noire as well as the Carnic Alps the boundary is often marked by a band of shales: "Boundary Shale", Hangenberg-Schiefer) make it more difficult to fix the boundary by biostratigraphic methods.

Nevertheless, beside the recently most favored sections of the Mt. Noire (Feist & Flajs, 1988), some other sections of the Paleozoic of Graz (Ebner, 1980; Sandberg et al., 1983; Ziegler & Sandberg, 1984) or of the Carnic Alps (Schönlaub et al., 1988) were prominent candidates for the Devonian/Carboniferous Boundary stratotype. On the basis of running geochemical, mineralogical, and facial investigations extraterrestrial events are very unlikely to have caused the change in sedimentation and fauna (Schönlaub in Ebner et al., 1991).

Within the pelagic preflysch facies two types of environments can be recognized:

Either the Tournaisian is represented by a very condensed carbonatic/lyditic succession (type 1) or falls for the most part within an erosional gap covering the Devonian/Carboniferous boundary (type 2).

Type 1 is best represented in the West-Mediterranean region (often called "Carbone basale non detritique" by local workers).

There it is characterized by only a few meters of Tournaisian limestone, often a band of shales within the Devonian/Carboniferous boundary level and lydites, frequently associated with phosporitic nodules within the Late Tournaisian and Lower Visean. Possible stratigraphic gaps are of submarine nature; conodont mixed faunas are caused, more probably by stratigraphic condensation. The most important fossils are: conodonts, radiolarians, and cephalopods. In the Pyrenees, Massif of Mouthoumet and Mt. Noire the following regional markers occur (in the top):

- Calcaires postjaspes (cu III, thickness of a few decameter)

- Fm. des Jaspes: cu II - base cu III (lydites, black shales, often with phosphoritic nodules and Mn-rich sediments. Locally a separation of a Fm. Jaspes inferieur and a Fm. Jaspes superieur by a few m of the cephalopod-bearing Calcaire intercalaire is possible).

- Calcaire supragriotte (often with a band of shales marking the Devonian/Carboniferous boundary, the thickness of cu I-cu II is only up to 3 m!).

Within the Pyrenees, Massif of Mouthoumet and Mt. Noire, Crillat (1983) differentiated the Tournaisian of type 1 into three subtypes with different regional distribution. However the proof of an erosional superposition of the Calcaire intercalaire above Lower Devonian in the Central Pyrenees and parts of the Mt. Noire indicate the proximity of both ± complete (type 1) and incompelete (type 2) successions.

At the top of this preflysch environment there are the Visean Calcaire postjaspes which are developed in the Mt. Noire as allodapic limestones with reworked shallow water fossils (Calcaire a Colonne cu IIIb, Feist, 1978; Feist & Flaljs, 1987).

Further thin layers of tuffitic volcanics ("cinerits") are known from some levels of the Jaspes superieure resp. Calcaires postjaspes (Wennekers, 1968; Boersma, 1973; Barrouquere et al., 1983).

In the Eastern Pyrenees the preflysch stage goes up to the Late Namurian with a sequence of intraformational lyditic breccias (100-150 m) and carbonatic rocks including the magnesite of Eugui and a band of shales with cephalopods of Namurian B age (Pilger, 1973; Bourrouilh, 1983).

Modified type 1 is also found in parts of the Alboral Balearic Trough (Massif of Cheloua, Menorca, Betic Cordillera) Sardinia, Calabria, Sicily, Carnic Alps, parts of the Graz Paleozoic, Balcanian W-Kraistids, the Uppony Mts., ? Bosnia and the Istanbul-zone of Pontids.

In the Carnic Alps the ± complete successions of the Kronhofkalk with its reduced thickness, local shale bands and lydites (but no phosphorites) resemble type 1 (Grüne Scheid: Müller, 1959; Gedik, 1974; Schönlaub, et al., 1988, Kronhofgraben: Schönlaub, 1969 and Elferspitz: Ebner, 1973a, b). But here the preflysch is terminated by lydites within cu II (Herzog, 1983, 1988). The termination of the preflysch by lydites and the lack of thicker limestones in highest preflysch positions was also reported from the Kabylean Massif (Bourrouilh, 1986), Calabria (Majeste-Menjoulais et al., 1984; Spalletta & Vai, in press), Sicily (Spalletta & Vai, in press), and the Istanbul-zone of the Pontids also with phosphorites (Kaya, 1969, 1978a, b).

In Uppony Mts. (Kovacs & Pero, 1983) and parts of the Rannach Group (Paleozoic of Graz; Ebner, 1978; Nössing, 1974, 1975) pelagic, carbonatic sequences reach the Namurian. The very reduced thickness of the continuous Tournaisian and thin Upper Tournaisian - Lower Visean lyditic niveaus (with phosphorites in the Paleozoic of Graz) suggest an environment similar to type 1.

Type 2 is represented within the Eastern Alps (Noric Nappe of Grauwacken-zone, Schönlaub et al., 1980; Schönlaub, 1972; Stolzalpe nappe of Gurktal thrust system Neubauer & Pistotnik, 1984; Neubauer & Herzog, 1985; parts of the Paleozoic of Graz (Ebner, 1978) parts of the Carnic Alps (Gedik, 1974), Karawanken Mountains (Tessensohn, 1974, 1975), Szendrő Mts. (Kovacs & Pero, 1983), ? Sardinia and in a few sections of the Pyrenees (=type C sensu Crilat, 1983).

Type 2 is characterized by erosional stratigraphic gaps of different continuity covering the Devonian Carboniferous boundary. The main lithology of the transgressive Lower Carboniferous (Upper Tournaisian - Lower Visean) is composed of micritic pelagic limestones often hardly distinguishable from the Devonian limestones below. At some levels, intercalations of lydites (without phosphorites) are seen.

Reworked carbonatic materials (microbreccias - limestone conglomerates/breccias) occur in a thickness of a few cm up to 10 m at the base of the Lower Carboniferous transgression.

The stratigraphic gaps at the base of this unit are due to:

- Local emergence and karstification (indicated by conodont mixed faunas in which faunal elements representing the interval of the gap are totally missing).
- Erosion and resedimentation of continuous carbonatic pelagic sequences (indicated by conodont mixed faunas with faunal elements representing the interval of the gap).
- Filling of submarine fissures with pelagic sediments (indicated by autochthonous pelagic micritic sediments with conodonts representing the interval of the gap; represented only in the Szendrö Mts.).

The thickness of this environment which starts within the Scaliognathus anchoralis zone of latest Tournaisian fluctuates from a few meters up to 100 m in the Hochlantsch Group of the Graz Paleozoic (Zier, 1983; Gollner & Zier, 1982, 1985).

In the Carnic Alps, carbonatic resediments with do/cu conodont mixed faunas (Plöckenpaß area Gedik, 1974; Schönwipfel Herzog, 1988), and the erosional contact of carbonatic Lower Carboniferous above Frasnian shallow water limestone (Bandel, 1972; Pohler, 1982) also make an erosional phase obvious in parts of the Carnic Alps.

Further lydites at the end of typ 2 preflysch environments are known from the Gurktal thrust system (Schönlaub, 1971; Neubauer & Pistotnik, 1984; Herzog & Neubauer, 1985).

Only in some areas (Hochlantsch Group of Graz Paleozoic, Szendrö Mts.) this pelagic environment reaches up to Namurian B. In the Szendrő Mts. these youngest preflysch sediments interfinger with both carbonatic platform sediments and the basinal siliciclastic Szendrö-phyllite Fm. In the Rannach Group of the Graz Paleozoic the pelagic environment is superposed after an erosional break within the Namurian A by the marine shallow water environment of the Dult-Fm. (Ebner, 1978; Kovacs & Pero, 1983; Gollner & Zier, 1982, 1985; Zier, 1983).

The example of the Paleozoic of Graz, Carnic Alps, and Pyrenees in which both types of preflysch environments exist in a short distance from each other demonstrates that the sedimentation of the carbonatic-lyditic-phosphoritic environment of type 1 did not happen in a deep sea environment. Further to this, Nössing (1974, 1975) pointed out that this sedimentation took place below the euphotic zone, between 60 and 300 m depth. This interpretation is additionally supported by bathymetric interpretations of geochemical (Mn) data (Buchroithner et al., 1979; Ebner & Prochaska, 1989). Parts of this shelf area emerged in Upper Devonian - Lower Carboniferous times and were affected by a transgression within the Late Lower Carboniferous. In these parts, preflysch sediments of type 2 were deposited, which are characterized by subaerial erosional gaps at their base.

The subaerial gaps in the interval of the Devonian/Carboniferous boundary and the maximum of Carbonatic pelagic sedimentation within the Uppermost Tournaisian-Visean are caused by two factors:

- a global trend with worldwide regression within the Upper Devonian culminating at the Devonian/Carboniferous boundary followed by a transgression during the Lower Carboniferous (Johnson et al., 1985, 1986; Vail et al., 1977; Sandberg et al., 1983, 1986, 1988; Veevers & Powell, 1987);
- local synsedimentary tectonics e. g. in the Paleozoic of Graz (Gollner & Zier, 1984, 1985; Ebner & Prochaska, 1989).

In Menorca (Bourrouilh, 1973; Buchroithner et al., 1980a) and the Little Kabyles (Bouillin & Bourrouilh, 1986) type 1 preflysch sediments are intercalated between a Devonian and a Visean/Namurian flysch environment.

The situation in the Outer Dinarids is difficult to reconstruct. Some reports on pelagic limestones, lydites and Tournaisian conodonts (see Ramovs et al., 1984) from SE and Central Bosnia point to a type 1 environment or to an intercalation of these lithologies within a (volcani-)clastic sequence.

Beside these Lower Carboniferous of typ 1 and 2 clastic (lyditic) environments with some intercalations of limestones are also possible within parts of the Southern Alps, the Noric Nappe of the Grauwacken-zone in the Eastern Alps (Eisenerz-Fm.; Schönlaub et al., 1982; Schönlaub, 1982; sequence near Veitsch described by Nievoll, 1985, 1987), the Jadar Paleozoic, Montenegro and the Chios autochthonous.

On account of the poor biostratigraphic and sedimentological data a reconstruction of these Lower Carboniferous clastic environments is problematic as well as their delimitation to the following flysch-environments.

Shallow water environments

Besides the above mentioned shallow water sediments closing the pelagic preflysch stage of the Graz Paleozoic and Szendrö Mts., thick autochthonous shallow water sediments in a preflysch setting are restricted to the Eastern and Southern Carpathians. There, these sediments show a minor Sudetic metamorphic overprint (low pressure greenschist facies, Kräutner, 1987).

The 800 - 1 000 m thick Tibau-Fm. of the East Carpathian Subbuccovina nappe covers directly the Prevariscan basement. It is built up of dolomites and limestones interfingering with metaconglomerates, sericitic shales, and phyllites (Bercia et al., 1976). Possible equivalents of the Tibau-Fm. are also found within parts of the Supragetic North Poiana Rusca unit (Kräutner, 1983, 1987). These were assigned to the Lower Carboniferous on account of palynological data (Iliescu & Kräutner, 1975; Kräutner, 1983, 1987).

Biostratigraphically proved Tournaisian is represented in the South Carpathian Danubian-units by the 300 m thick detrital, sparry Ideg-Lmst. with brachiopod, coral, trilobite and conodont faunas (Cordacea et al., 1960; Mirauta, 1964; Nastaseanu & Kräutner, 1983; Kräutner, 1987).

Further allochthonous blocks and reworked materials of Lower Carboniferous to Upper Carboniferous (Namurian - Bashkirian) shallow water sediments are interbedded within the Carboniferous flysch all over the area. Therefore during the late preflysch stage and/or the early flysch stage shallow water shelf areas must have been connected with the deeper environments.

For the Alboran-Balearic Trough it is assumed, that these shallow water materials were transported from the Paleo-North/African shelf area to the more northern flysch basins. The age of the biodetrital limestones (partly with foraminifers and algae) ranges from the Upper Visean - Lower Bashkirian (Milliard, 1959; Bourrouilh, 1977; Bourrouilh & Lys, 1976; Buchroithner et al., 1980a, b; Herbig, 1984; Chalouan, 1987; Ebner, 1991).

In the Mt. Noire area the allodapic materials (with Visean 3b shallow water fossils) of the Clacaire colonne and olistolithic blocks (with corals, foraminifers, gastropods and productids indicating the total Uppermost Visean) of the Mt. Noire flysch point to a shallow water shelf situated north of the flysch basin (Mamet, 1968; Engel et al., 1978; Engel, 1982; Feist, 1978; Feist & Flajs, 1987).

Reworked Lower Carboniferous materials and slabs of shallow water carbonates indicating a close relation of both environments are also known from the Hochwipfel-flysch of the Southern Alps (Mambrini, 1975; Schönlaub & Flügel (in prep.), Spalletta & Venturini, 1988). Further the position of the Devonian - Carboniferous "Felsen" of the Seeberg-Karawanken area which are interbedded in a flysch matrix may possibly be seen in the light of gravitational mass transported sediments (see also Tessensohn, 1971, 1974, 1975).

In the Jadar Paleozoic the Drucetic limestone was originally also partly sedimented in a shallow water area, on the evidence of sheet cracks and birds-eyes (Scharfe, 1977).

The Praca-Beds in SE Bosnia have recently been interpreted as olistolitic blocks of Lower-Middle Carboniferous fossiliferous (crinoids, foraminifers, bryozoans, algae, ostracods) carbonatic materials (Krstic et al., 1988).

Regional aspects

Pelagic sediments dominated by lydites (type 1) are the typical Carboniferous preflysch facies of the West-Mediterranean area. This continuous facies is best represented within the Pyrenees, Massif of Mouthoumet, Mt. Noire, and Catalonian Coastal Ranges, where it was replaced by the flysch within the latest Lower Carboniferous.

Within the Alboran Balearic Trough synorogenetic flysch sediments are developed within the Devonian and Visean-Namurian. In this area the proof of Lower Carboniferous lyditic preflysch sediments is missing where an orogenetic event was recognized within the Devonian/Carboniferous boundary level (Chalouan, 1987; Boullin & Perret, 1982).

The pelagic environment of type 1 is characterized by lydites, ± phosporites, cephalopod-bearing nodular limestones, and a very low rate of sedimentation (3-5 Bubnoff units = m/10^6 years) with any terrestrial input. Calculations on the depositional depth range between 60 and 300 m for the Graz Paleozoic (Nössing, 1974, 1975) and 300-500 m for the Mt. Noire (Wendt & Aigner, 1985).

This environment together with transgressive sea level fluctuations, which produced a better circulation and mixture of the sea water, were responsible for the "phosphoritic-event" characteristic of the Upper Tournaisian and Lower Visean of the West-Mediterranean.

This "phosphoritic-event" is also found in parts of the Graz Paleozoic and the Istanbul-zone. It is clearly contemporaneous with the maximum of Lower Carboniferous carbonatic sedimentation in areas where the Devonian/Carboniferous boundary is covered by stratigraphic gaps.

Modified pelagic lyditic facies (type 1) is also represented in parts of the Eastern- and Southern Alps, Uppony Mts., Calabria/Sicily, Western Kraistids, ? Bosnia and the Istanbul Paleozoic.

Pelagic carbonatic preflysch sequences dominated by gaps (type 2) are characteristic of the Alpine area, the Szendrö Mts. and Sardinia. Most of these gaps are the result of a subaerial or submarine (Szendrö) erosional event before the Uppermost Tournaisian. Besides sea level fluctuations more probably synsedimentary tectonics were also responsible for these gaps.

In the Eastalpine Gurktal Thrust System, the Southern Alps and Sardinia the preflysch stage of type 2 was terminated within the Visean by flysch. Within the Grauwacken-zone Lower Carboniferous carbonates are followed by the non flyschoid clastic environment of the Eisenerz-Fm. (Schönlaub, 1982). Variscan flysch is also missing within the Paleozoic of Graz. Here the Variscan sequence is terminated by shales and shallow water limestones of the Namurian- ? Westfalian Dult-Fm. Limestones of similar lithological character and age are also known from the top of the Szendrö and Uppony-preflysch.

In the low grade metamorphic Eastern and Southern Carpathians, the Carboniferous begins with very thick marine shallow water sediments, which locally directly overlie the Previsean basement (Kräutner, 1987). These are the only Lower Carboniferous preflysch shallow water sediments of the total area, which occur in an autochthonous position. They are superposed by volcaniclastic sediments, which are tentatively interpreted as presudetic flysch environments (Ebner, 1991).

All other indications of shallow water preflysch sediments go back to allochthonous materials intercalated inside the flysch. This suggests that over the entire area the late preflysch- and/or flysch basin was in a close relation to a shelf area with shallow water sedimentation.

In the Balcanids the Lower Carboniferous as well as the Late Devonian is represented by flysch sediments. The single exception is the lyditic-pelitic-carbonatic Berende-Fm. in the Western Kraistids (Spassov et al., 1978; Spassov, 1983; Maslarevic & Krstic, 1987).

Further to the east and also inside the Dinarids it is difficult to reconstruct the Carboniferous preflysch stage on account of the scarce present data base. But nevertheless the analysis of allochthonous flysch-blocks points to the existence of shallow water and pelagic preflysch environments up to the Bashkirian and a younger flysch in the Hellenids. On the other hand, the Thracian flysch of the Istanbul-zone indicates a Visean age, whereas the total Variscan preflysch of the Great Caucasus falls within the Precarboniferous.

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Pre-Alpine metamorphic events in Gemericum

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Abstract

The Early Paleozoic sequences in Gemericum are prevailingly metamorphosed under chlorite zone of greenschist facies conditions. Apart from the contact aureoles near granitoid bodies no high-temperature zone with gradual transition to greenschist facies rocks was distinguished. In the peripheral part of this unit slices and fragments occur and they are usually overthrust on the low-grade rocks. Beside the metamorphic conditions estimated according to mineral assemblages in different lithological type of rocks, the sequences of metamorphic processes are discused here.

Introduction

The Early Paleozoic of Gemericum is prevailingly metamorphosed under chlorite zone of greenschist facies conditions (Zoubek, 1953; Kamenický, 1967, etc.). Only along the northern and eastern boundary of this unit a few occurrences of amphibolite facies metamorphites were found (Rozložník, 1965; Dianiška & Grecula, 1979; Hovorka et al., 1979; Faryad, 1988). The low-grade sequences mainly in the central part are overprinted by contact metamorphism of granitoid magmatism. Several Rb-Sr but also some K-Ar data indicate Permian age of this granitoid magmatism. Varga (1973) distinguished biotite and K-feldspar zones around granite bodies and due to large extent of these zones he interpreted them as zones of regional metamorphism. The presence of amphibolite facies metamorphites distributed on the boundary of Gemericum led Grecula (1982) to suggest a possible correlation with granitoid magmatism. According to this interpretation the Hercynian metamorphism resulted from high-temperature zonation and anatexis in the central part of Gemericum

Geological setting

Gemericum (eastern part of the Slovenské rudohorie Mts.) is built predominantly by the Early Paleozoic rock complexes (as well as accompanying young Variscan granitoids) and Late Paleozoic and Mesozoic sequences in its peripheral parts (Fig. 1). With relation to the nearby units of Veporicum Gemericum is in nappe position and boundary between them is represented by the Lubeník-Margecany scale zone (line). The Early Paleozoic of Gemericum is divided by the majority of authors (Andrusov, 1958; Bajaník et al., 1984; etc.) into two superposed groups: the older Gelnica and the younger Rakovec Group (including the Čermeľ and Ochtiná Formations; Fig. 1). The Gelnica Group is prevailingly represented by grey-green and dark schists and phyllites, sialic to intermediate, occasionally also basic volcanic rocks. Black schists and phyllites in the central part usually contain lenses of marbles and lydites. Beside grey-green phyllites of the Rakovec Group also contain metamorphosed mafic rocks. According to Grecula (1982) the lithostratigraphical development of the Gelnica and Rakovec sequences is common, they differ only in the type of volcanism in the upper part. Along the Rakovec Group slices and fragments of amphibolite facies rocks so called the Gneiss-amphibolite Complex of Gemericum are distributed (Faryad, 1990). The Early Paleozoic sequences in the peripheral parts of Gemericum are overlaid or overthrust by the Upper Carboniferous, Permian and Mesozoic Formations.

Greenschist facies metamorphites

Metasedimentary rocks

They are represented by chlorite-sericite quartz phyllites, black schists and phyllites containing

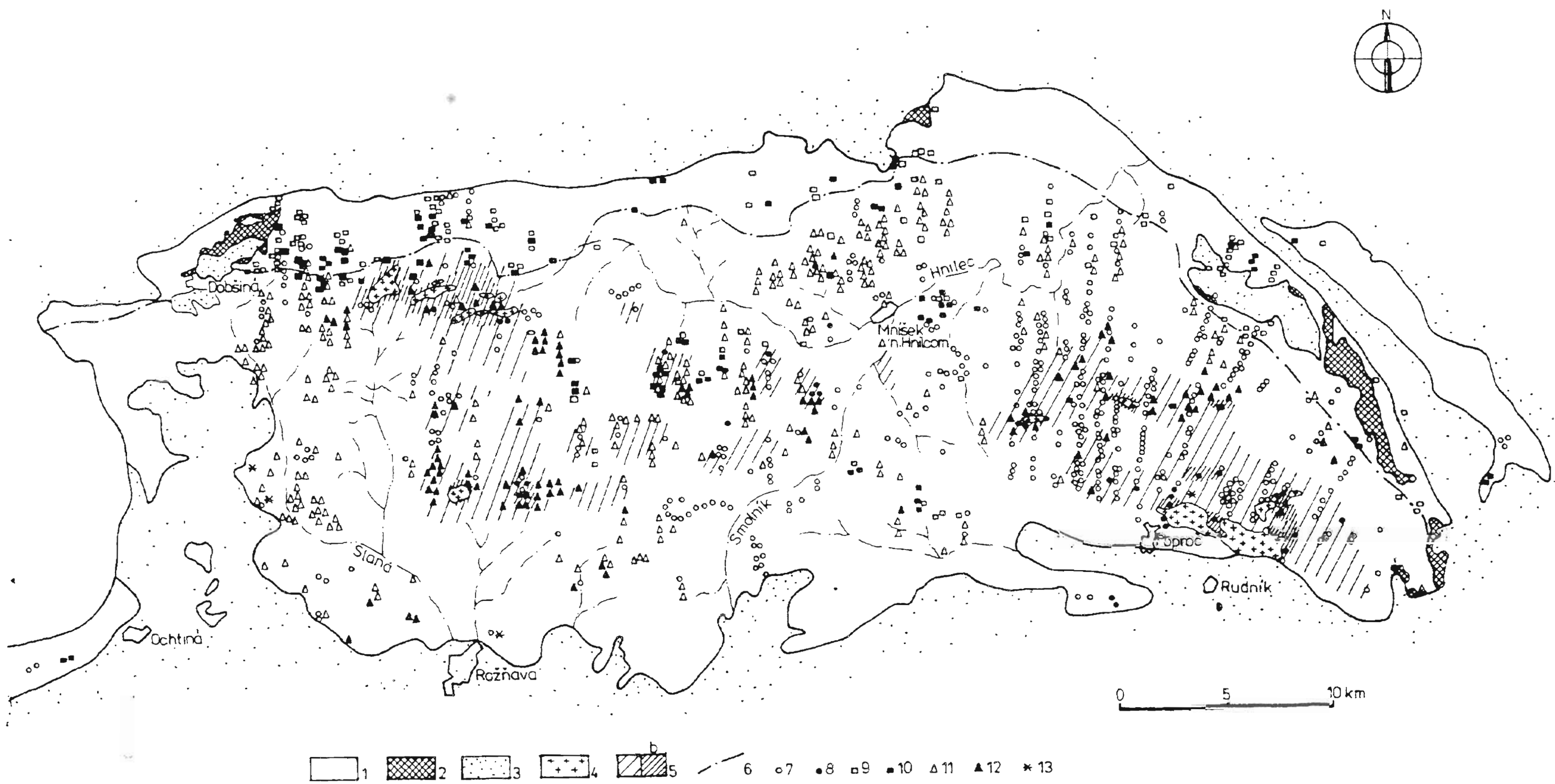

Fig. 1. Distribution of metamorphic mineral assemblages in the Early Paleozoic rocks of Gemericum. 1 - greenschist facies sequences, 2 - Gneiss-amphibolite Complex, 3 - Late Paleozoic and Mesozoic Formations, 4 - granitoids, 5 - zones of contact metamorphism (chlorite zone, b - biotite zone), 6 - boundary of the Rakovec (to the North) and Gelnica Group (to the South), 7-8 - metasedimentary rocks containing quartz, chlorite, white mica and (8) biotite, 9-10 - mafic rocks comprising albite, chlorite, epidote ± carbonate and (10) actinolite, 11-12 - salic to intermediate volcanic rocks consisted of quartz, albite, white mica, chlorite ± microcline ± epidote and biotite, 13 - sedimentary and salic volcanic rocks containing chloritoid.

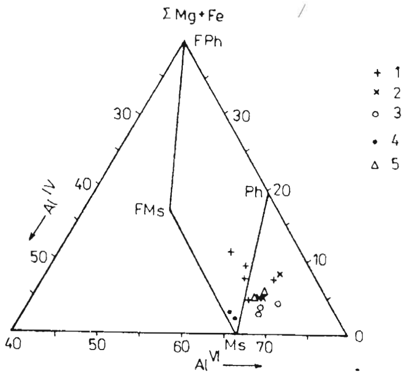

Fig. 2. Mg+Fe: $Al^{IV}Al^{VI}$ diagram for white mica from the Early Paleozoic regional metamorphosed pelitic and pyroclastic rocks. 1 - metapelites, Locality Grajnár and Nálepkovo (Ms + Chl + Qtz), 2 - black metapelites, Locality Úhorná (Ms + Chl + Qtz + All), 3 - metapelites Locality Rožňava (Ms + Cld + Qtz), 4 - metapyroclastic rocks, Locality N. Slaná (Ms + Cld + Qtz), 5 - phyllites of biotite zone from contact with Gneiss-amphibolites Complex, Locality Rudňany.

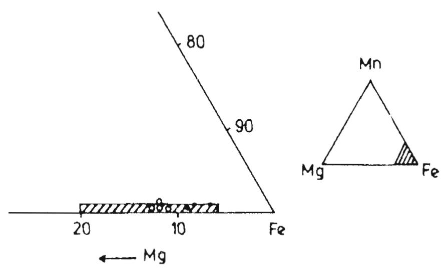

Fig. 3. Chloritoid composition from metapelites and metapsamites (open circles) and metapyroclastic rocks (filled circle). Dashed- chloritoid fields from chlorite zone (Ashworth & Evirgen, 1984).

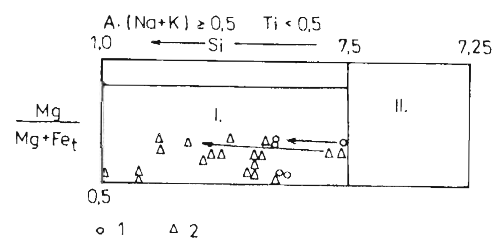

Fig. 4. Amphibole composition from regional metamorphosed mafic rocks. 1 - Gelnica Group, 2 - Rakovec Group.

lenses of carbonatic rocks. Schistosity in these rocks usually coincides with bedding and it is conspicuous mainly in peletic varieties. Sedimentary microstructures are present mainly in psamitic rocks, clasts of quartz and feldspars where are preserved. Outside the zones affected by contact metamorphism, sedimentary rocks contain quartz, phengite-rich white mica (Fig. 2), chlorite, albite (mainly in metagraywackes), rutile, occasionally also carbonates (calcite and siderite). In the western and southern part of Gemericum also iron-rich chloritoid is developed (Fig. 3). In relation to schistosity seems to be postkinematic mineral but it is older than crenulation cleavage. The last foliation has regional extent and usually crosses the schistosity.

In the volcanosedimentary rocks overprinted by contact metamorphism near Zlatá Idka (the central part of Gemericum) we have found also kyanite (Faryad and Dianiška, in print). It occurs with andalusite, muscovite and quartz. Both kyanite and andalusite are usually replaced by muscovite. Almandine garnet as accessory mineral was found only in three samples in the zone affected by contact metamorphism near Poproč (the eastern part of Gemericum). Pseudomorphs of chlorite and carbonates after garnet and biotite were reported also from the Rakovec phyllite near tectonic contact of overthrusting the amphibolite facies rocks (Faryad, 1990).

Carbonatic rocks contain mainly calcite, siderite, magnesite and occasionally also ankerite, quartz, muscovite, chlorite and talc. In some Mn-rich carbonatic rocks spessartine-rich garnet, rhodonite, pyroxmangite, manganocalcite, rarely muscovite and chlorite have been found.

Mafic rocks

Metabasites are widespread mainly in the Rakovec Group; only sporadically they occur in the Gelnica Group. They usually comprise albite, chlorite, epidote, actinolite sphene, calcite or siderite, occasionally stilpnomelane and biotite. Actinolite occurs namely in metabasalt, metadolerite and metagabro. In the last rocks there is actinolite content up to 20 vol. % and zoned actinolite grains are also present (Fig. 4). In metabasite from Ochtina actinolite reaches content up to 60 vol. %. Beside actinolitic hornblende also hornblende in one sample has been found.

In some doleritic varieties from the Gelnica and Rakovec Groups relicts of magmatic pyroxene (augite-diopside) are rarely preserved. Stilpnomelane with actinolite and biotite were found only in three localities in the northern and central part of Gemericum. A paragenesis of blue amphibole (barroisite to Na-hornblende) with actinolite, Ca-rich garnet, albite and chlorite was reported from the Rakovec locality (Hovorka et al., 1988).

Sialic to intermediate metavolcanite

The Early Paleozoic rhyolitic to dacitic metavolcanic rocks are aboundant mainly in the Gelnica Group. Recently we have distinguished also subvolcanic members of these rocks (Faryad, 1990). Metarhyolite and their subvolcanic members usually display a poorly developed and disoutinuous schistosity. The most frequent mineral components are quartz, white mica, chlorite, albite and microcline. Occasionaly calcite, rutile and biotite are present there. Apart from quartz and feldspars magmatic biotite and hornblende are observed.

In the western part of Gemericum iron-rich chloritoid of similar composition as in phyllitic rocks is developed. A paragenesis of actinolite with albite, epidote, chlorite, sphene and calcite was found in mylonitized dacite from Mníšek n/Hnilcom. In one of these samples also almandine-rich garnet was found, associating with grunerite, hematite, magnetite and accessory biotite. These mineral assemblages were probably formed after iron-rich quartzite occurring as small lenses in dacitic tuff.

Conditions of metamorphism

According to mineral parageneses in metapelites, metagraywackes and metasandstones of the Gelnica and Rakovec Groups, temperatures did not exceed greenschist facies conditions of chlorite zone. Supposing the Variscan age for chloritoid formation, metamorphic conditions in consistency with data from Ashworth and Evirgen (1984) correspond to chloritoid-bearing chlorite zone of greenschist facies (300 - 380 °C at about 300 MPa). Biotite occurrence in K-rich sialic metavolcanite does not represent typical biotite zone usually defined in metapelite. After phase equilibrium (Korikovsky, 1979) biotite in such rocks appears at 330 °C. He supposes such temperature also for chloritoid formation (Fig. 5). Using the pyroxmangite-rhodonite geothermometer (Schultz-Guttler & Peters, 1986) we have obtained temperature 370 ± 15 °C from Mn-rich carbonatic rocks.

According to the new experimental data of Liou et al. (1987) metamorphic temperatures did not exceed 400 - 430 °C in mafic rocks of the Rakovec and Gelnica Groups. Only in metabasites from Ochtina containing actinolitic hornblende (±hornblende) indicate temperature up to 450 - 460 °C, Faryad, 1991a). The age of metamorphism of the last rocks is according to radiometric date from amphibole (Kantor & Ďurkovičová, 1980) 340 mil. y. The origin of blue amphibole from Rakovec is (by Hovorka et al., 1988) interpreted as the result of medium- to high pressure and low-temperature metamorphism, during subduction of the Rakovec Group rocks. They suggest the subduction process to be older than Variscan greenschist facies metamorphism in Gemericum.

Low-temperature boundary of the Variscan re-

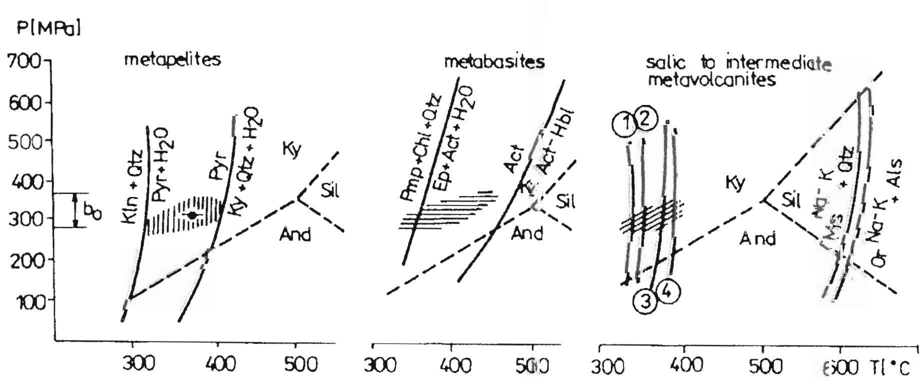

Fig. 5. P-T conditions of Variscan regional metamorphism in Gemericum (excepting the Gneiss-amphibolite Complex and the mafic rocks from Ochtina). Temperatures are estimated after mineral reaction in different rocks types. Filled circle in metapelite field corresponds to temperatures obtained using the rhodonite-pyroxmangite geothermometer (Schultz-Guttler & Peters, 1987). Pressure conditions are considered after b₀ values of muscovite (Sassi & Vozárová, 1987) and b₀ values calculated from chemical analyses of muscovite methoda after Guidotti et al., 1989). The pressure ascertained by the first method are supposed as minimum. 1-4 - mineral reactions: 1 - Stp + Chl + Mc = Bt + Ms +Qtz, 2 - Chl_{Fe} + Prl = Ctd_{Fe} + Qtz, 3 - Prl = Ky (And) + Qtz + H_2O, 4 - Chl_{Fe} + Qtz = Alm + H_2O

gional metamorphism is not well known in the Early Paleozoic rocks, since the prehnite-pumpellyite to greenschist facies conditions were reached also during Alpine metamorphism. Results of X-ray study from black schist carbonaceous matter show that greenschist facies conditions were not reached everywhere (Mola et al., 1986).

The pressure conditions of this greenschist facies metamorphism were due to the lack of index minerals not clear. Using the b_0 values of muscovite Sassi & Vozárová (1987) supposed low-pressure geothermal gradient (40 °C/km, 270 MPa at 370 - 400 °C) for Variscan regional metamorphism in Gemericum. Using the b_0 values calculated from microbrobe analyses of muscovite (methoda after Guidotti et al., 1989) we have obtained relative high-pressure conditions (270 - 380 MPa at 370 - 400 °C) (Fig. 6). The medium-pressure conditions preceding the low-pressure ones are indicated also by Si content in muscovite (3.12 - 3.34, methoda after Massonne and Schreyer, 1987). "Microprobe analyses of minerals are given in Faryad (1991a, b) and they were carried out on the electron analyser JEOL 733 Superprobe in Dionýz Štúr Geol. Inst. in Bratislava. In addition to actinolitic amphibolite from Ochtina, the Gneiss-amphibolite Complex (both of them are in tectonic contact with nearby rocks) and contact metamorphosed rocks, the Early Paleozoic rocks indicate no evident regional metamorphic zonation. Such metamorphic feature together with kyanite occurence indicate low-medium to medium-pressure condition of the metamorphism in Gemericum.

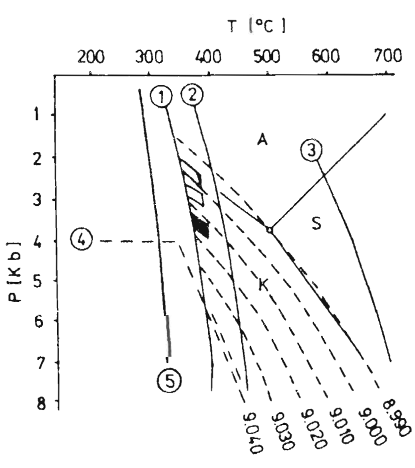

Fig. 6. Diagram for b_0 values of muscovite and position of some important reaction (Guidotti & Sassi, 1986): 1 - Kln + Qtz = Prl + H_2O, 2 - Prl = Als + Qtz + H_2O, 3 - St + Qtz = Als + Bt + H_2O, 4 - maximal low-pressure stability field of glaucophane (Maresch, 1977), 5 - Kln + Qtz = Prl + H_2O (Frey, 1987). Opne rhomus - conditions of metamorphis ascertained after b_0 values muscovite (Sassi & Vozárová, 1987; Mazzoli & Vozárová, 1989), filled rhombus - pressure conditions after b_0 values calculated of microprobe analyses of muscovite (methoda after Guidotti et al., 1989).

The Gneiss-aphibolite Complex of Gemericum

Amphibolite facies metamorphites occur as tectonic slices and fragments along the Rakovec Group on the northern and eastern boundary and among Mesozoic rocks in the southern part of Gemericum. In the northern part these rocks together with the Rakovec Group are overlaid the Upper Carboniferous conglomerates, containing their pebbles. The Gneiss-amphibolite Complex is represented by amphibolites, gneisses, rarely serpentinites and carbonatic rocks. In contrast to the previous works (Dianiška & Grecula, 1979; Hovorka et al., 1985) supposing sedimentary protolithes for gneisses, recent petrological data indicate, they are derived from calc-alkaline volcanic (andesite and dacite) rocks of island-arc affinity and they are comagmatic with associating mafic rocks metamorphosed into amphibolites (Faryad, 1990).

Amphibolites consist of hornblende, plagioclase, epidote, rarely garnet, quartz, actinolite, chlorite and sphene. At least four genetic types of amphibole, formed during progressive, retrogressive processes and diaphtoresis, can be distinguished in these rocks. The most common one is magnesiohornblende formed during amphibolite facies conditions. Two types of garnet-bearing amphibolite distinguished according to garnet composition were reported by Hovorka & Spišiak (1985). Hornblende and garnet occur also in carbonatic rocks forming lenses in amphibolite.

Gneisses contain plagioclase, quartz, hornblende muscovite, biotite, rarely garnet and cummingtonite. Garnet is usually reversely zoned and some grains are rounded by a narrow garnet rim of different composition (Faryad, 1990b). Serpentinites occurring in some localities comprise antigorite, tremolite, antophyllite and talc. According to Hovorka et al. (1984) serpentinites occur in the lower part of tectonic slices of amphibolite facies metamorphism.

Contitions of metamorphism

P-T conditions of amphibolite facies metamorphism were estimated using the garnet-biotite, hornblende-biotite and hornblende-garnet geothermometers. The calculated P-T values were scattered in greater intervals. Using the nornblende-garnet thermometer (Graham and Powel, 1984) and garnet-hornblende-plagioclase-quartz geobarometer (Kohn and Spear, 1989, only after P_{Fe1} and P_{Fe2} moduls) we have obtained 500 - 640 °C and 450 - 600 MPa. The average P-T values calculated after P_{Fe1}, P_{Mg1}, P_{Fe2}, P_{Mg2} moduls (l. c.) correspond to 570 - 700 °C and 700 - 900 MPa (Fig. 7). The high P-T values were obtained from the centre of a regressively zoned garnet. The high-pressure and temperature values due to the presence of mineral assemblages seem not to be real; also the low Na_2O contents in hornblende (max. 1.45 w. %) do not indicate such high-pressure conditions (Laird, 1982).

Several K-Ar data from hornblende (except one date giving 448± 23 m. y.) indicate Variscan age of amphibolite facies metamorphism (Cambel et al., 1980; Kantor et al., 1981). In some localities strongly foliated gneisse and amphibolite were found containing mineral parageneses of epidote-amphibolite facies conditions. Together with biotite and garnet-bearing phyllite of the Rakovec Group and serpentinites they were probably formed during obduction and overthrusting of the Gneiss-amphibolite Complex on the Rakovec Group rocks

Contact metamorphism of the Early Paleozoic sequences

We have distinguished three zones in the contact aureoles of gemeric granitoids: chlorite zone (zone I) developed in distance of 1 km from granitoid, biotite zone (zone II) up to 500 m and andalusite zone (zone III) up to several meters around granitoid bodies (Faryad, 1990). The chlorite zone differs from its regional metamorphosed equivalents only in metamorphic fabrics; e. g. the newly-formed muscovite cross schistosity in metapelites and metapsamites and quartz-beariny varieties display considerable recrystallization. In the high-temperature part of this zone also chlorite spots appear. In metavolcanic rocks, occasionally also in metapelites biotite appears, too.

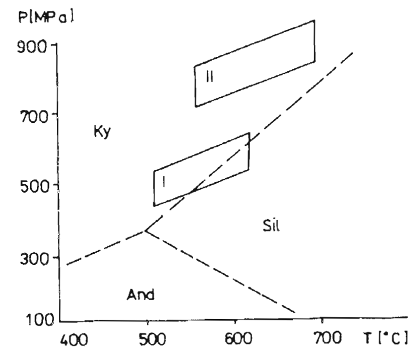

Fig. 7. P-T conditions of Gneiss-amphibolite Complex metamorphism calculated using the garnet-hornblende-plagioclase-quartz mineral association (Graham & Powell, 1984; Kohn & Spear, 1989): I - P-T values calculated using only the moduls P_{Fe1} and P_{Fe2} and II - moduls P_{Fe1}, P_{Mg1}, P_{Fe2} and P_{Mg2} (l. c.).

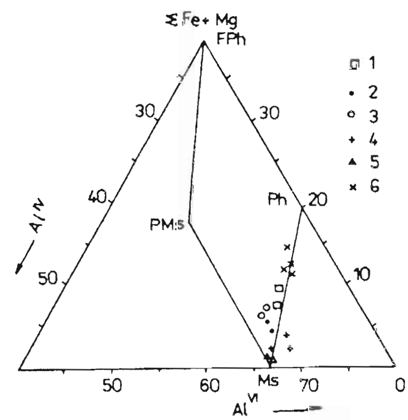

Fig. 8. Muscovite composition from contact metamorphosed sedimentary rocks of Gemericum. Parageneses: 1 - Ms + Chl + Qtz, 2-3 - Ms + Chl + Bt + Qtz, 4 - Ms + Bt + Qtz + Grt, 5 - Ms + Bt + And (1-5 - metapelites), 6 - Ms + Chl + Ab + Ep (metagreywackes).

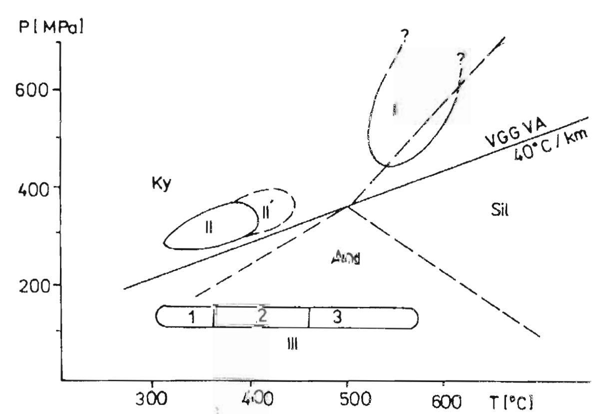

Fig. 9. Likely P-T conditions of Variscan regional and contact metamorphism in Gemericum: I - Gneiss-amphibolite Complex of Gemericum, II - Gelnica and Rakovec Groups, II - metabasites from Ochtina, III - contact metamorphism: 1 - chlorite zone, 2 - biotite zone and 3 - andalusite zone. VGG Va - Variscan geothermal gradient of Eastern Alps (Bogel et al., 1979).

The occurence of single zone is in good agreement with granitoid depth determinated by borehole or estimated by gravimetric methods. It was found in several thin-sections that contact metamorphism is younger than schistosity and cleavage. Contact among granite and country rocks is usually tectonized. In several localities granitoid bodies are overthrust on phyllites.

Metasedimentary rocks

The biotite zone is represented by the occurrence of biotite, occasionally also andalusite or garnet. In carbonaceous rocks of andalusite andradite-rich garnet, hedenbergite-rich pyroxen and hornblende were found (Faryad & Peterec, 1987).

Towards granitoids in the country metasediments, white mica become rich in muscovite component (Fig. 9). Spots of finegrained chloritoid associating with almandine (relict ?) and oxichlorite were found in two samples of chlorite zone near the Poproč granite. Similarly as in regional metamorphosed sedimentary rocks chloritoid is rich in iron. In biotite zone also normally zoned spessartine-rich garnet was found. Biotite associating with chlorite contains more iron (F=Fe/Fe+Mg = 62-65 percent) than biotite from andalusite zone (F=59-60 percent).

Metabasites

Basic rocks overprinted by contact metamorphism are rarely present in Gemericum. In biotite zone actinolitic hornblende occurs, some time hornblende and occasionally also Ca-rich garnet and biotite. Some actinolite grains are normally zoned. In contrast to the greenschist facies metamorphites, schistosity features disappear and epidote, actinolite and albite veinlets appear resulting from striped structure of mafic rocks.

Sialic to intermadiate metavolcanites

Rhyolitic rocks of chlorite zone are distinguished from biotite zone according to the intensity of recrystallization and size of mineral grains. In both zones similar mineral paragenesis are present consisting of quartz, muscovite, albite, microcline, and biotite. Comparing with phyllites of chlorite zone white mica from sialic volcanic rocks is rich in muscovite component. In andalusite zone also apatite appears occasionally containing inclusion of rutile. In felsic rocks of this zone towards granite K_2O, SiO_2, Rb and TiO_2 contents increase (Faryad, 1989).

In rhyodacites and dacites of chlorite zone quartz, muscovite, biotite, albite, epidote, is present rarely actinolite, sphene and microcline. In andalusite zone hornblende can also appear.

Conditions of metamorphism

In consistence with index minerals, temperatures for chlorite, biotite and andalusite zone can be estimated 300 - 350° and 450 - 550 °C respectively. Temperature 550 °C was also calculated using the thermodynamic parameter for skarns (Faryad - Peterec, 1987). According to the mineral parageneses in contact aureoles and the lack of staurolite in pelitic rocks we suppose the gemeric granitoids were intruded at shallow level (3,5 - 6 km and 100 - 150 MPa, Fig. 8).

Conclusion

The Upper Carboniferous conglomerates containing pebbles from the underlying Rakovec Group and the Gneiss-amphibolite Complex indicate that metamorphism of the Early Paleozoic rocks and overthrusting of the Gneiss-amphibolite Complex on them are older than Upper Carboniferous sedimentation. Mineral parageneses in sedimentary, mafic, intermediate and felsic rocks indicate that the Gelnica and Rakovec Groups together with the Čermeľ and Ochtiná Formation were metamorphosed prevailingly under chlorite zone of greenschist facies conditions. Occurrence of some metamorophosed basic rocks containing a large amount of actinolite, rarely also actinolitic hornblende (in the west part of Gemericum) seems to be the result of uprising and overthrusting of these rocks from depth in Gemericum.

Consistent to the lack of metamorphic zonation and kyanite occurrence in the low-grade rocks the low-medium to medium-pressure conditions can be supposed for this metamorphism. Medium-pressure conditions indicate also geobarometric data form the Gneiss-amphibolite Complex of Gemericum. Mineral compositions in the last rocks assign regressive trend of metamorphism. With respect to the geochronological data the metamorphism of the Gneiss-amphibolite Complex is Variscan but according to the age value 448±23 mil.y. this metamorphism could have began in the end of Caledonian Orogen.

In contrast with other West Carpathian Variscan granitoids the Gemeric granitoids are younger (Permian in age) and accompanied by tin mineralization. They were intruded into greenschist facies sequences at shallow level (3,5 - 6 km). Wide high-temperature contact aureoles were not formed around granitoid occurrences because of small size of granitoid bodies on surface.

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Critical thermic isograde in metamorphic-hydrothermal model of vein mineralization on the background of the Variscan events; Gemeric unit, Western Carpathians

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Abstract

Ore vein mineralization in the Gemericum belt has a very close relationship to metamorphic processes. Siderite and siderite-sulfidic veins occur only within low grade metamorphosed Paleozoic sequences while antimonite veins are in the zones of amphibolite facies of metamorphism. Geological and petrological studies indicate that during the amphibolite facies of metamorphism about 4.4 wt% of water was liberated from metapelites and these ones were depleted of some elements (e. g. the content of Cu in black shales was reduced from the original content of 25 ppm to 14 ppm). Concentrations of metals in metamorphic fluids were calculated (in ppm): Cu 100, Pb 27, Zn 16.7, Ni 64, Co 22, Ba 1 375, Sn 18, Ag 2.5, As 64, Sb 6, Mo 4 and Fe 2 - 10 %. The isotopic ratio of $^{87}Sr/^{86}Sr$ from vein barite (ranging from 0.71042 to 0.71541) points that this is derived from crustal rocks during regional metamorphism. The largest concentration of fluids were generated at the metamorphic peak of about 500 °C. The position of the 500 °C temperature isograde to the source rock sequences is more important for the different metal concentrations in fluids and for mineral composition of ore veins, too.

Introduction

The Gemericum represents the tectonic unit of the West Carpathians with the largest concentration of ore deposits in Slovakia. Ore veins made of siderite and sulphides are the most numerous. Antimonite veins compose independent belts within the Gemeric unit. Strata-bound base-metal mineralization occurs in two lithostratigraphic levels in Lower Paleozoic rock. Also, siderite bodies in the carbonate horizons with lydites in black shale sequences of Early Paleozoic age are significant. Metasomatic magnesite bodies in Carboniferous lithologies are among the largest deposits in the world. The relation between the replacement of ore mineralization and metamorphic zones is very close. Siderite-sulphidic veins as a rule are connected with low-grade metamorphosed sequences while Sb-veins were discovered only in the zones of amphibolite facies metamorphism. Metasomatic siderite bodies are present in the upper part of the greenschist facies only.

The origin of ore veins and metasomatic bodies was hitherto either related with granitoid pluton (Varček, 1961, a. o.) or with hypothetic ophiolitic complex occuring in 10-15 km dept beneath Alpine nappes of Gemericum (Rozložník, 1989). In recent years a direct genetic connection is appearing between the ore veins and metamorphic processes. A new genetic model was elaborated assuming relations between metamorphic-hydrothermal processes and ore veins (Grecula, 1982; Grecula et al., 1989) which has subsequently been completed by data on mobilization of metals from source rocks during the metamorphism between the chlorite and the cummingtonite-horblende-albite zones (Radvanec, 1987, 1988) on the O, C, O, Sr isotopic data (Radvanec et al., 1990; Žák et al., 1991) as well as by data on the structure and chemical composition of minerals within zones of ore element mobilization in rocks and lower parts of ore veins (Bartalský, 1991).

Geological background of mineralization

The Gemericum consists mainly of metamorphosed vulcano-sedimentary sequences of Silurian and Devonian age, only locally covered by discordant Carboniferous and Permian formations. The existence of a riftogenous basin is presumed during the formation of Lower Paleozoic sequences, with oceanic crust opiolites in the central rift zone and continental crust with basalt(andesite)-rhyolitic volcanism in the other sections (Grecula, 1982).

The lithological filling of the Lower Paleozoic sedimentary basin is spatially very variegated. The lower part is detritic as it is formed by a series of black laminated metapelites, and its top part is composed of pelitic black shales with lenses of carbonates and lydites. The middle part is formed by a variety of metapelites. A variegated diabase-keratophyre formation lies upon the base and in the top part of this lithological sequence. The upper part of Lower Paleozoic unit consists of volcanic rocks. There is a very close relation with geotectonic evolution. The basic rocks coresponding to ophiolitic suites with MORB type volcanism are situated mainly in the northern part of the Gemeric belt and the rhyolitic sequences represent the continental passive margin.

Lower Paleozoic rock sequences were metamorphosed by prograde and retrograde Variscan and Alpine diaphtoritic events. We have evidence about metamorphism between the chlorite and garnet-horblende zone in the basic schist in the Variscan time (Dianiška & Grecula, 1979; Hovorka & Spišiak, 1981; Faryad, 1991b). The metapelites and semipelites had been overworked by the similar metamorphic conditions in the same period as well (Faryad, 1991a, c, d). During Variscan subduction-collision events low-pressure-higher-temperature metamorphic overworking occured locally leading up to the granitization and formation of small granitic bodies of S-type (Rb-Sr ages 240-280 Ma, 87Sr/86Sr initial ratios between 0.711 and 0.734; Kovách et al., 1986). At the same time mineralization processes took place and Variscan overthrust-nappe structure was formed after mineralization events. The existence of Variscan nappes is supported by evidence in geology, geochronology etc. (Grecula, 1982). Variscan tectonic units of Gemericum are overworked by Alpine nappe and shear tectonics. The beginning of the activity of shear and transpresion zones is based on geological and radiometric data for the first stage of Alpine events and for significant movements completed on the same structures in Neogene (Grecula et al., 1990). The Alpine shear zones already destroyed the Variscan ore veins as in the Variscan nappes.

The types of vein mineralization are spatially different. Single belts of veins occur in Gemeric complexes metamorphosed to various degrees. E. g. the antimony ore veins which are in rocks near the boundary between green-schist and amphibolite facies. Similarly , the nickel and cobalt bearing minerals in siderite ore veins occur mainly within the gneis-amphibolite complexes. However, in the higher metamorphic environments there are no ore veins, but in the belts where the intensity of metamorphism has been lower the ore veins are rather

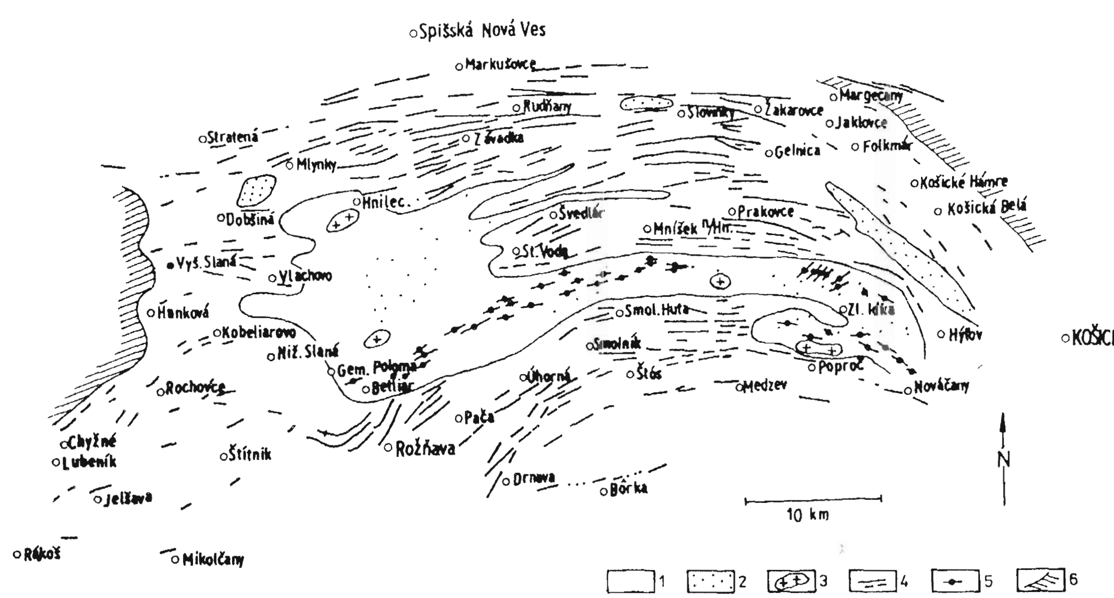

Fig. 1. Map of the ore vein deposits in the Gemericum related to the metamorphic zones. 1 - rock sequences metamorphosed in the condition of greenschist facies, 2 - rock sequences metamorphosed in the condition of lower amphibolite facies, 3 - granitic bodies, 4 - siderite-sulfide ore veins, 5 - quartz-antimonite veins, 6 - crystalline rocks of the Veporicum unit.

completely

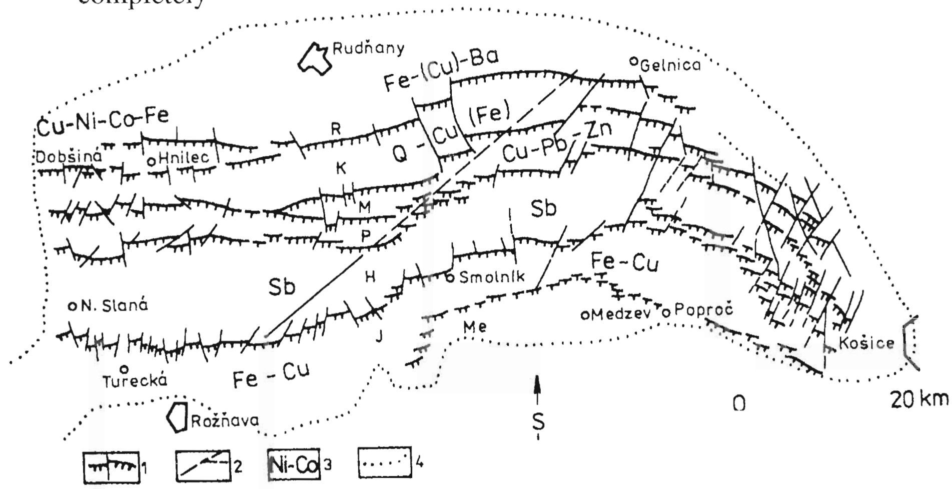

Fig. 2. The dominant element associations of the vein mineralization in relation to the Variscan nappes of the Gemericum. Variscan nappes: R - Rakovec, K - Kojšov, M - Mníšek nad Hnilcom, P - Prakovce, H - Humel, J - Jedľovec, Me - Medzev. 1 - nappe line, 2 - faults, 3 - dominant element association of the ore veins, 4 - the boundary of the Gemeric unit.

developed. In the low-grade metamorphites the number of siderite-sulphide ore veins is the highest and their depthward extension exceeds 1 km. The lowermost parts of the siderite ore veins are situated within rock environments roughly representing the lower part of the amphibolite facies. On the other hand in the high crystalline environment of Gemericum (gneiss, migmatite and granitoids) the ore veins composing the siderite-sulphide ore formation are already lacking (except for small occurrences bound to mylonite belts). To the contrary, antimony ore veins (already mentioned) and tin-tungsten-molybdenium-gold ores related to the crystalline complexes with granitoids are present there.

There are also very close relations between the mineral associacion of the veins and the rock filling of the Variscan nappes of Gemericum. Different nappes consist of different rock sequences based on their original position in the sedimentary basin and were metamorphosed to a variable degree and contain vein mineralization of different types. Siderite veins, which besides Cu, locally contain Ni, Co and Cr are bound to areas of ophiolite suite, quartz-sulphidic veins there are in rock sections of transition crust type. Siderite-sulphidic veins are found in nappes containing rocks originating in the continental crust.

Metal mobilization during metamorphic events

In areas of higher metamorphism original and stratiform deposits, are strongly modified but also the mobilization of metallic elements during metamorphic events is common.

Regional geochemical and petrological studies (including lithogeochemical research over an area 1 500 km^2) indicate that during the regional metamorphism (chlorite and cummingtonite-horblende-biotite zone) about 4.4 wght % of water was liberated from metapetites. We have evidence - mineral assemblage (Cal + Qtz + Chl + Ttn) that the metamorphic system had been opened for fluid phase and "free" water passed into a fluid phase (Radvanec et al., 1990). During the above mentioned processes the rocks was depleted of some elements. For example, the content of Cu in black shales was reduced from (from chlorite to amphibolite facies) the original content of the average 25 ppm to 14 ppm; the content of Cu in the strata-bound ore bearing horizon was reduced from 78 ppm to 56 ppm. The content of Fe in black shales was reduced 4.22 wght % to 1.7 wght %, and represents a 41 % decrease of Fe. The decrease of Fe from the ore-bearing horizon was 58 % (from 5.6 % to 3.2 %). Moreover, the metamorphism of black shales (more as a 1 km thick complex) and the carbonate

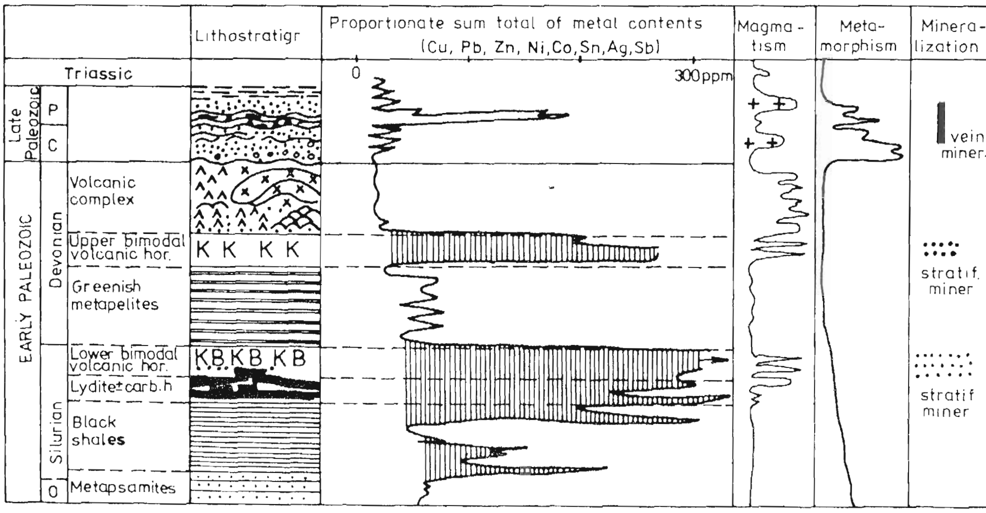

Fig. 3. Proportionate sum of metal contents of the Early Paleozoic units in the Gemericum (used more than 12 000 analyses of rock samples) and magmatic, metamorphic and mineralization events.

rocks produced significant amounts of CO_2 and CH_4. The main content of metals in the fluid was derived from black shales, lydite horizon, ore-bearing stratiform layers and variegated volcanic complexes. Model calculations indicate the following concentrations of metals (in ppm) in metamorphic fluids: Cu 100, Pb 27, Zn 16.7, Ni 64, Co 22, Ba 1 375, B 107, Sn 18, Fe 2 - 10 %, Ag 2.5, As 64, Sb 6, and Mo 4 (Radvanec, 1988). The main portion of these metals were transported in metamorphic fluids like HCO_3^- complexes.

Isotope evidence

Generaly, positive $\delta^{34}S$ (+1 ‰ to +17 ‰) from stratiform polymetalic deposits (Smolník and Mníšek nad Hnilcom) (Kantor & Rybár, 1970; Hurný, 1980) and disseminated sulphides mineralization agree well with the idea of the formation of these deposits by submarine exhalative processes and dominantly inorganic reduction of seawater sulfate (Ohmoto, 1986). Bacterial reduction of seawater sulfate seems to be more important only locally, especially in the units with higher amounts of organic matter ($\delta^{34}S$ = 0 to -7 ‰ , Žák et al., 1991). As a whole, the rocks of Gemericum are a reservoir of sulfur with predominantly positive $\delta^{34}S$ values.

A majority of siderite-quartz-sulphidic vein deposits were formed from fluids with positive sulfur isotopic composition of hydrothermal fluid ($\delta^{34}S$ of source sulfur is in the range from +2 to +10 ‰ , Žák et al., 1991). The more complex $\delta^{34}S$ pattern of vein deposits with mineral association with siderite-quartz-sulfides-barite ($\delta^{34}S_{tet}$ = -10 to +1 ‰ $\delta^{34}S_{py}$ = 0 to +7 ‰ , $\delta^{34}S_{bar}$ = +10 to +19 ‰ , Cambel and Jarkovský, 1985) in Rudňany requires a complex explanation taking into account mixing of two different fluids. There is oxidation of fluids of deeper origin and the influx of sulfates mobilized from overlying Permian evaporites (Ohmoto & Lasaga, 1982). Nevertheless, the rocks of the Lower Paleozoic sedimentary basin of Gemericum are the main source of sulfur in the vein.

The value of carbon isotopes in the Lower Paleozoic carbonate rock, which are in some part changed metasomaticaly to siderite, rarely) in magnesite, are close to -5 ‰ , while in kerogen and/or graphite there are between -22 and -26 ‰ (in $\delta^{13}C$ PDB).

An extensive set of $\delta^{13}C$ determinations on hydrothermal carbonates (mostly siderite) from vein deposits indicate that carbon isotopic composition of hydrothermal fluids was shifted outsite the "deep - seated" range (Valley et al., 1986) to more negatives values ($\delta^{13}C$ of $CO_{2\ fluid}$ = from -8 to -11 ‰). It is calculated, according fractionation, tobe between H_2CO_3 (or CO_2) and siderite (Carothers et al., 1988; Ohmoto, 1986; Golyšev et al., 1981). Mixing of deep seated carbon and carbon derived from organic matter and carbonate are possible.

The calculated $\delta^{18}O_{fluid}$ is in the range of +5 to +10 ‰ SMOW throughout the main (siderite) stages at all studied vein deposits. They are in the range typical for fluids of magmatic or metamorphic origin and/or any origin which have undergone high temperature isotopic exchange with magmatic and metamorphic rocks (Žák et al., 1991).

The relatively high and variable $^{87}Sr/^{86}Sr$ ratio of vein barites in the range from 0.71042 to 0.71541 generally indicate crustal sources of hydrothermal fluids (Radvanec et al., 1990). This ratio is higher than those of Paleozoic marine evaporites and limestones, suggesting that influence of this source is only of minor importance (Burke et al., 1982; Kump, 1989; Taylor & McLennan, 1986). Correlations exist between the $^{87}Sr/^{86}Sr$ barite and $^{87}Sr/^{86}Sr$ initial ratios of Variscan S-type granitoids in this area (Kovách et al., 1986).

Mineralogical evidence

Evidence of metamorphogenous CO_2 bearing fluids are confirmed by the results of the study of the inclusions in the "roof" zone vein mineralization. The fluid inclusions was a study of the quartz from the siderite veins and from small secretion veins or lenses and eyes in the deeper parts below the zones of vein formation. The fluid inclusions in all samples were very similar, and were created in the same geological processes. There are two principal kinds of primary fluid inclusion: the first one with low salinity (below 10 wgt. % ekv. NaCl) with solid solution and fluid CO_2 and/or CH_4, T_H =260 °C. This is similar to the first generation of CO_2 rich fluids inclusions containing the fluid released from prograde metamorphic reactions in pelitic and carbonate rocks (Sisson & Hollister, 1990). The second one with high salinity about 20 wght % ekv. NaCl (composition of solution NaCl, H_2O, Ca^{2+} calculated according Roedder, 1984), T_H = 160 °C, T_E= from -43 to -55 °C (Radvancová, 1991). Inclusions with Ca^{2+}, NaCl are typical for processes of regional metamorphism (Roedder, 1984). This data represents two unmixible solutions and describe the evolution of mineralogical processes in time.

The metamorphogenous fluids attack the older minerals in the rock and there is an equilibrium between rock and fluid phase with crystallization of new minerals. These processes are very closely tied to the generation of vein mineralization. It is confirmed by fluid inclusions in the quartz, very similar to the ore veins and the rocks. The content of Mn in fluids has an important role, because its higher content is very typical for gemericum siderite veins. The demonstration of rock and mineral alterations by metamorphogenous fluids was observed in the older minerals (e. g. rim pyrite with Mn = 0.98 wght %, replaced the older clear rutile to the Mn-ilmenite). The siderite as a younger mineral is situated on the rim of minerals (pyrite, magnetite) and in the cracks of minerals. Mobilization of REE in the allanite is found in these processes. Simultaneously the new minerals of the ilmenite and Ti-magnetite with Mn (MnO = 0,19 - 5,28 % MnO) have precipitated in the rock. These new minerals are typical for metamorphogenous association in the Gemericum and they have had very close relation with the generation of vein mineralization.

Model of critical thermic isograde

Metamorphic model of vein mineralization (Grecula, 1982) was completed and made more precise by isotopical, geochemical, mineralogical and other data. The recent examination of this model in prospects of ore deposits also gave positive results. Up till now this model has not beeen recognized implanly the real (although necessary generalized) geological conditions of the lower paleozoic rock sequences metamorphosed from the lower part of greenschist facies up to amphibolite facies with the manifestation of anatexis. The origin for the creation of the model of critical (metamorphic-thermic isograde - CTI) with 450 - 550 °C temperatures of the best fluid and metal leaching from rocks is the application of the metamorphic model for various rock environments. The determining factors for CTI model are temperature and pressure of metamorphic peak and rock environments in which the metamorphic processes with this temperature peak are taking part. The CTI model cannot be reconstructed exactly within the framework of one metamorphic process in the Variscan subduction-collision terrains (Gemericum also belongs here) due to very strong overworking by overthrust and nappe tectonics. The CTI model cannot be determined in high-grade crystalline complexes because metamorphites corresponding P-T conditions of CTI were high above those shown in the present erosion level and therefore were eroded also with accompanied ore mineralization. This is the reason why siderite-sulfidic vein deposits cannot be found in high-grade metamorphic complexes. If there are small indications of this type of mineralization in crystalline complexes, they are always connected with diaphtoritic or mylonitic zones, that is, with conditions of lower metamorphic degree.

This is why vein siderite mineralization occurs only in those regions where predominantly low-grade metamorphites are present and where the higher-grade metamorphites as source rocks are in the deeper part of the crust. From this point of view low-grade metamorphites make up the environment where ore veins are deposited. If overthrust-nappe tectonic style is characterized for mentioned regions, then the higher as well as lower-grade metamorphites with various types of mineralization are located in tectonic positions in respect to the present surface. For example veins of siderite formation are not present in the environment of higher metamorphites, but, on the other hand, Sb-veins are present or such environments remain without vein mineralization.

The regions with the above mentioned conditions of genesis and preservation of siderite vein formation are rare in the frame of Alpine belts. Gemericum is not only the region with the largest concentration of siderite veins but the region where such vein deposits are exploited at present. This enabled their detailed and continual study in mines to the depths of about 800 - 1 000 m (that is from the upper parts of veins down to the depth where fluids were generated and able to mobilize into cracks and into the origin of ore veins. Moreover, such "depth profile" enabled the study of the grade of metamorphism at the depth where fluids were generated. Various isotopic and mineralogical studies paved the ground for a genetic metamorphic model of vein mineralization and a critical thermic isograde model with emphasis on lithology of rock complexes and P-T conditions in the peak of metamorphism. Not only the mineral association but also concentration of ore minerals depends on temperature (for example 400 °C, 500 °C or 600 °C). In the case where the metamorphic peak did not

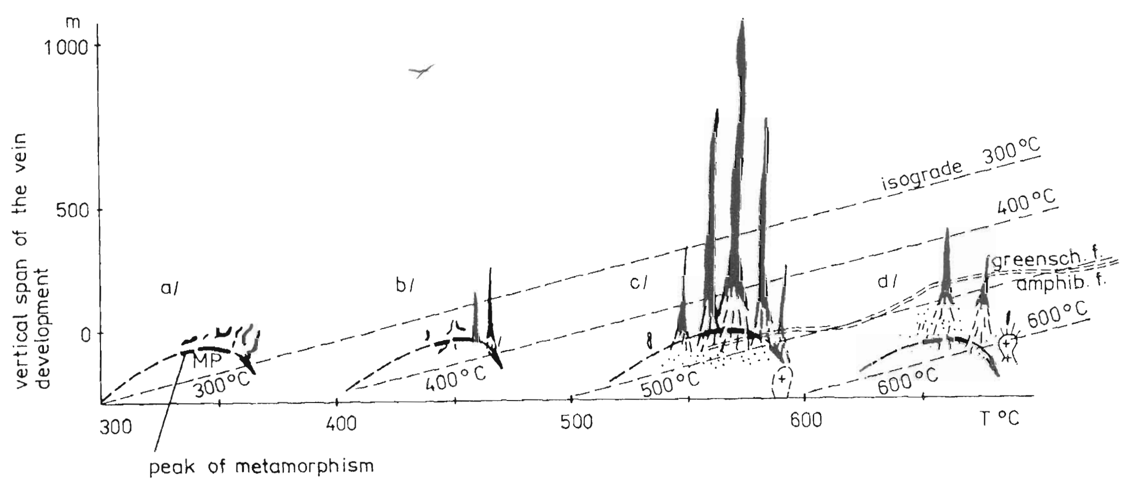

Fig. 4. Relations between metamorphic fluid generation and thermal peak of metamorphism expressed by the plot of temperature and vertical span of ore veins. Value of the metamorphic peak: a - 300 °C with generation of metamorphic secretion veinlets and lenses without ore content, b - 400 °C with small veins mostly without significant ore concentration, c - 500 °C with the main phase of ore-bearing fluid generation and with the generation of ore veins in lower temperature metamorphic levels (siderite and sulphides), d - 600 °C with subsequent phase generating ore veins of smaller vertical extent (siderite and Fe, Cu, Sb, Co, Ni, As sulphides) and antimonite veins with Sb-sulphosalts ±Au, Ag; granitization and melting processes with local intrusions and with showings of Sn - W ores and rare elements.

exceed 400 °C (the respective mineral association is chlorite-albite-phengite) more significiant ore vein bodies are absent, except of veinlets and lenses composed of quartz-albite±calcite. The largest concentrations of fluids have been generated at a metamorphic peak of about 500 °C (biotite to horblende subfacies) and their subsequent transport was responsible for the development of significiant ore veins. Under metamorphic conditions, reaching ot their peak about 600 °C, the ore veins already generated, but very rarely, and their lower parts reached into amphibolites and gneisses. In places where these veins occur in metabasites (amphibolites), even cobalt and nickel-bearing sulphides are present in the veins. In cases where the thermal peak of metamorphism was even higher than 600 °C, then ore veins of siderite-sulphide formation are lacking and the related metamorphic processes are manifested by anatexis and local intrusions of granites giving occurences of W-Mo-Sn-Au ores.

The indicated observations peak in favour of deductions that the thermal maximum of metamorphism is located between 500 °C and 550 °C together with the presence of an incipient retrograde path which wich induced metamorphic boiling is the favourable condition for the maximum fluid release, i. e. the time of the main ore producing phase. This also proves why the amount of ore vein is so strictly varied in the metamorphic complexes.

The second important condition for ore vein generation is that the metamorphic fluids become carriers of metals mainly by HCO_3^- complexes. These metals are being released from source rocks during metamorphism. Therefore this second condition for the generation of ore veins is important also; mainly that the metamorphic events related with peak of metamorphism should reach the main metalliferic source horizon. In the Gemeric unit, such horizon was located by Radvanec (1988) in the thick complex of black shales with lydites and carbonates as

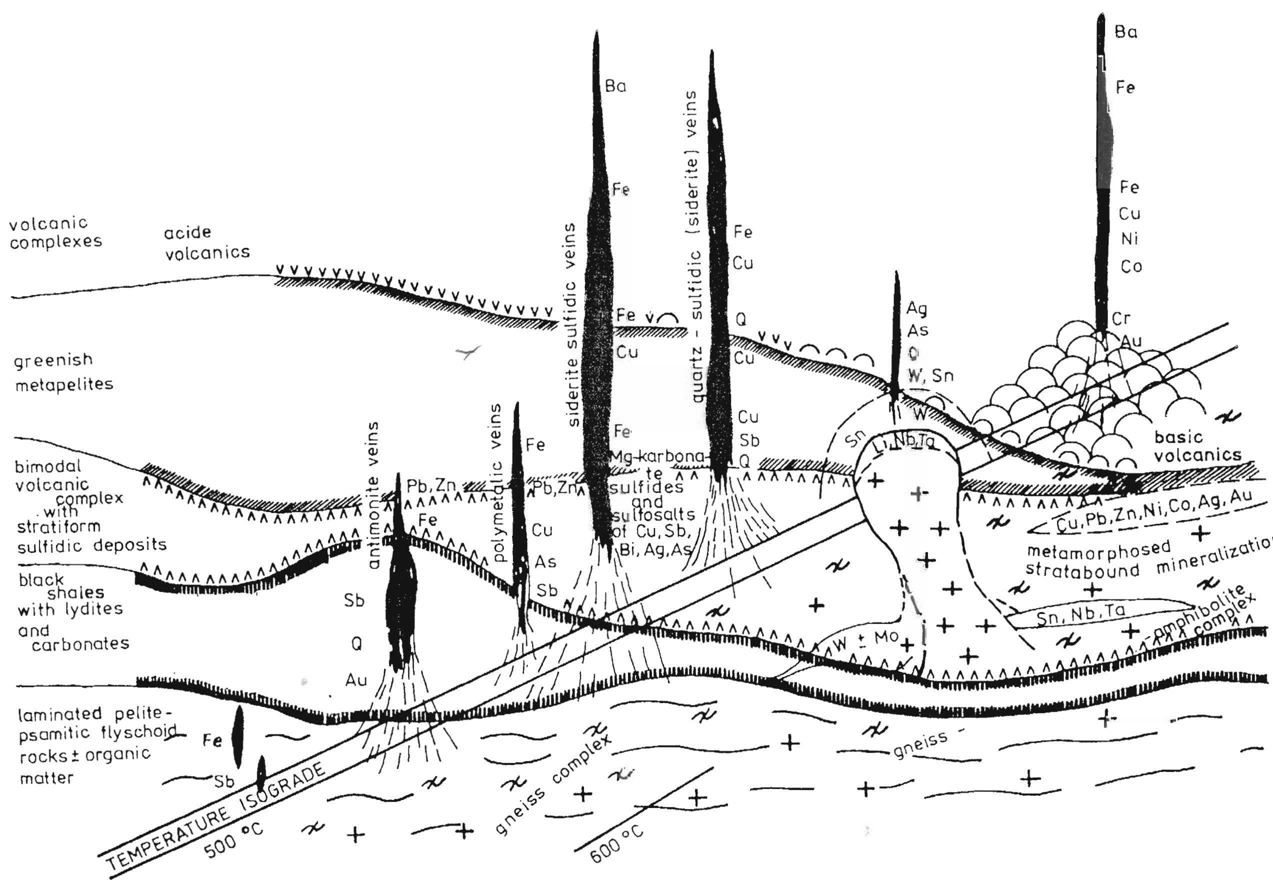

Fig. 5. Model of critical temperature isograde.

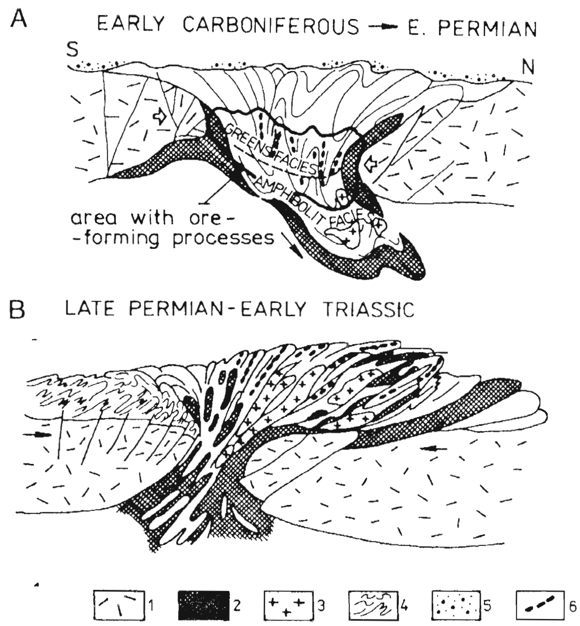

Fig. 6. Metallogenetic processes on the background of the Variscan tectonic and metamorphic events. A. Early Carboniferous - Early Permian stages: collision events with metamorphic and mineralization processes. B. Late Permian - Early Triassic stages with formation of the nappe structures in the accretionary terrane. Mineralized structures were overthrusted together with rock sequences. 1 - continental crust, 2 - oceanic crust, 3 - orogenic granitoides, 4 - sedimentary and volcanic sequences of the Early Paleozoic basin, 5 - Middle Carboniferous remnant basins with clastic sediments, 6 - ore veins and mineralized structures.

well as in the variegated volcanic complex with the presence of strata-bound ores (in spite of their low concentrations). Such rock complexes are carriers of 10 - 100 times more metals if compared with the background value. Large differences in metal content occurs also between the indicated source lithologies and other lithologic complexes of the Paleozoic age in the Gemeric unit. Because it was proved that black shales and the variegated volcanic complexes represent the main source of metals for ore-bearing fluids, the relation between the location of the indicated critical thermic isograde between 500 and 550 °C and the source horizons mentioned is significant not only for the generation of ore veins but also for their characteristic mineral assemblages. These relations determine whether the ore veins will be mostly antimonite ores or siderite-sulphide ores with numerous further varieties what is determined by the character of source lithological complexes with their respective mineral assemblages in the ore veins is shown in figs. 4, 5.

Conclusion

All the geological, geochemical, and isotopic criteria, indicate a close genetic relationship of vein mineralization in the Gemericum to the processes of metamorphism. Vein mineralization alone is located in zones of lower grade metamorphism, but fluid generation is presumed in the amphibolite-facies zones. Vein mineralizations were formed during prograde phases of metamorphism of about 500 °C. Silicate minerale assemblage indicate temperature between 400 - 550 °C in zones of fluid generation; in zones of fluid concentration the temperatures were 350 - 450 °C (pyrite-arsenopyrite and mica-feldspar thermometers) and the deposition of vein hydrothermal minerals took place at temperatures between 130 and 250° according to fluid inclusion and isotopic data. Nevertheless, fluids responsible for vein deposits comprise not only metamorphic fluids but also deep circulating convective fluids.

Not only the grade of metamorphism but also the lithological character of rock sequences as sources of liberated metals is important for the genesis of new ore veins and their mineral association. For example - Sb mineralization was derived mainly from black shales, siderite-sulfidic mineralization from variegated volcano-sedimentary sequences. Mineralization processes took place when the metamorphic peak was achieved during the Variscan subduction-collision events. Later, during the end of Variscan tectonic events large overthrusts and nappes were created and the vein mineralization was displace. The final tectonic destruction of ore veins was caused by Alpine deformation processes.

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The metamorphic Mn-ore deposit of Razoare (Romania) and its geological setting

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Abstract

The geological setting and metamorphic formations of the Preluca Mts., belonging to the North Transylvanian Crystalline "Islands" are described with emphasis on the Mn-ore deposits of Razoare. K/Ar datations of muscovites and biotites gave Alpine ages (~100 Ma) for the metamorphic country rock.

Introduction

The first describer of this area was the Transylvanian polyhistor Benkö (1793) who gave account of the Macskamezö (Razoare) micas. Pošepny (1862) first described the crystalline formations: gneiss and micaschist units and crystalline dolomites. Hauer & Stache (1863) mentioned the Preluca crystalline "Inselgebirge" in their famous work "Geologie Siebenürgens". Detailed maps were published by Hofmann (1887) & Koch (1898, 1902).

Kossmath & John (1905) were the first to describe the Macskamezö (Razoare) Mn-ore deposit. They gave a list of the oreforming minerals and the results of 25 chemical analyses. Quiring (1919) published a detailed mineralogical description of the ore and the deposit, also describing the secondary, oxidized weathering zone.

Kräutner (1937) gave a revision of the crystalline schists of Preluca. He considered that carbonate rocks formed some synclines and the age of metamorphic rocks was Hercynian.

In 1944 Pantó made a study on the ores, supplying new data on the chemistry of the Mn-ore.

After World War Two, new research started in the field of mineralogy, petrology and geophysics: Gherasi (1953), Gherasi & Sandu (1953), Gherasi & Bodin (1954). In 1957 Stanciu wrote contributions to the tectonic framework of Preluca Mts., and Dimitrescu (1963) described the western part of this massif. Detailed maps and mineralogical-petrographic data were supplied by Kalmár (1966, 1972, 1973).

The systematic exploration of the Mn-ore deposit and that of the pegmatites began in 1975 with detailed mapping, drilling and mining activities.

Regional setting

The Preluca Mts., along with six other metamorphic massifs, constitute the so-called North Transylvanian Crystalline "Islands" (Insule Cristaline din NV Transilvaniei"; Szádeczky, 1927) (Fig. 1). These massifs represent uplifted parts of the metamorphic basement of the north-eastern part of the Pannonian and Salaj Basins, as proved by deep drillings. They also represent the most elevated blocks of the so-called Somes platform which separates the Transylvanian basin from the Pannonian Basin. These elevated blocks are surrounded by Paleogene and Neogene continental and epicontinental deposits intruded - in some places - by small igneous bodies of Paleogene and Neogene age.

The Preluca Mts. is the most complete and interesting massif: it has the greatest extent (approx. 140 sq. km.) and contains metamorphic Mn-ore bodies of economic importance.

Geological outline of the Preluca Mts.

The structure of these mountains is the Salnita - Stoiceni asymmetrical anticlinorium. It is orinted SW-NE. The oldest unit occurs in its axial zone, that is, the Razoare Gneiss. This formation passes into the Magureni Carbonate Formation. The high-

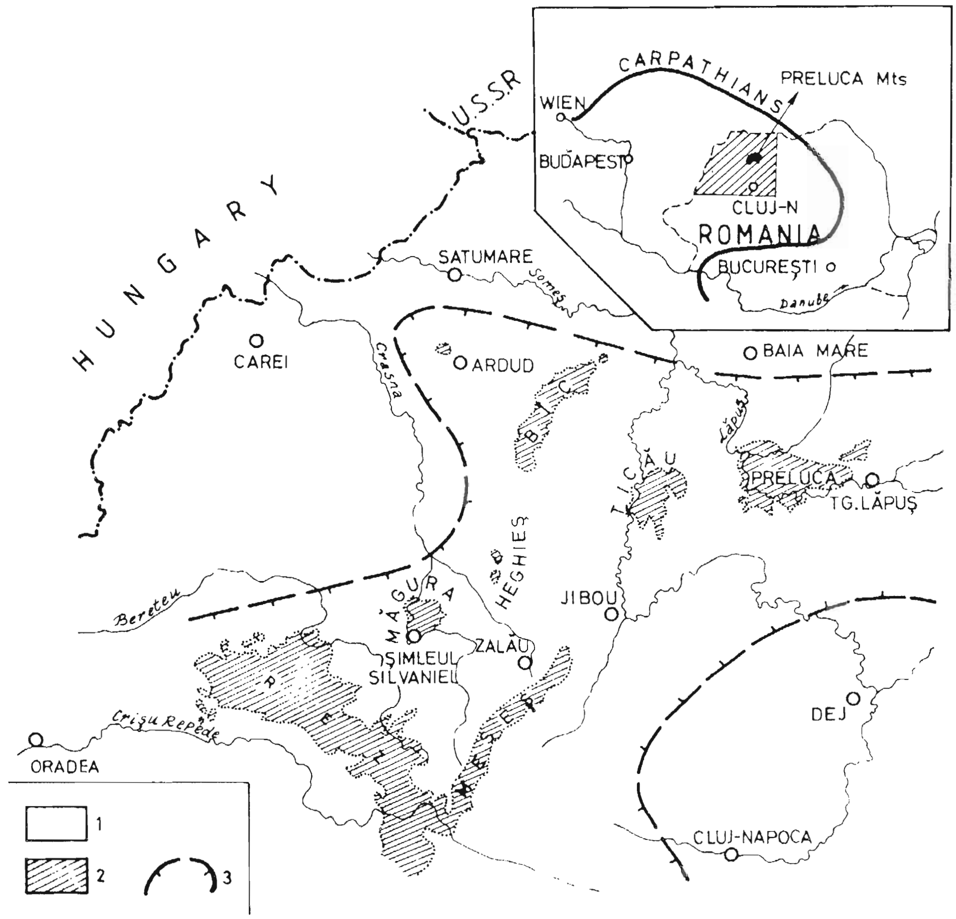

Fig. 1. The North Transylvanian Crystalline Islands and their position in the Carpathian area (see inlet). 1 - Tertiary and Quaternary deposits, 2 - Metamorphic rocks, 3 - The boundary of "Somes-platform" after geological and geophysical drilling data.

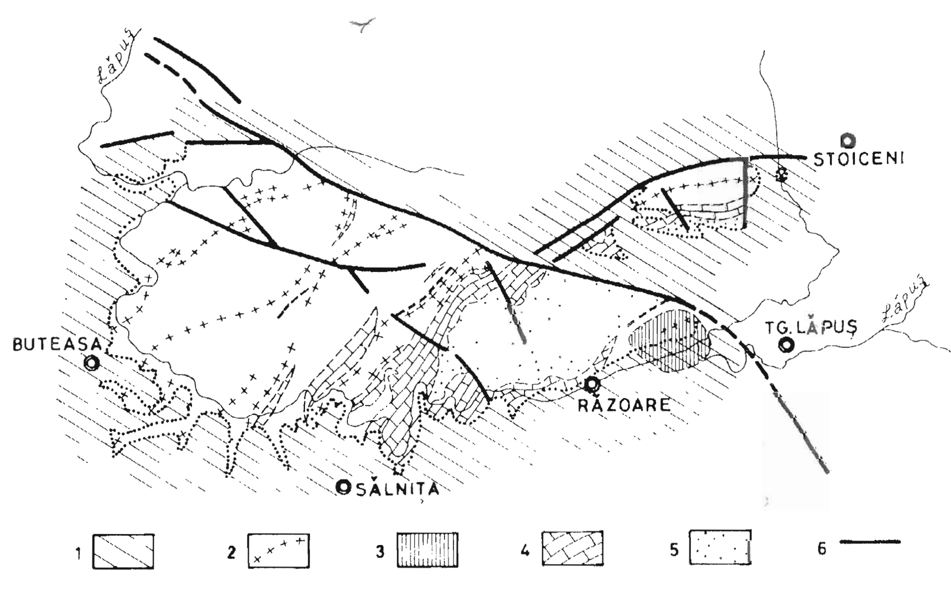

Fig. 2. Geological sketch of the Preluca Mts. 1 - Tertiary and Upper Cretaceous sediments, 2 - Preluca Formation, 3 - Area with Mn-ore lenses, 4 - Magureni Formation, 5 - Razoare Formation, 6 - Faults.

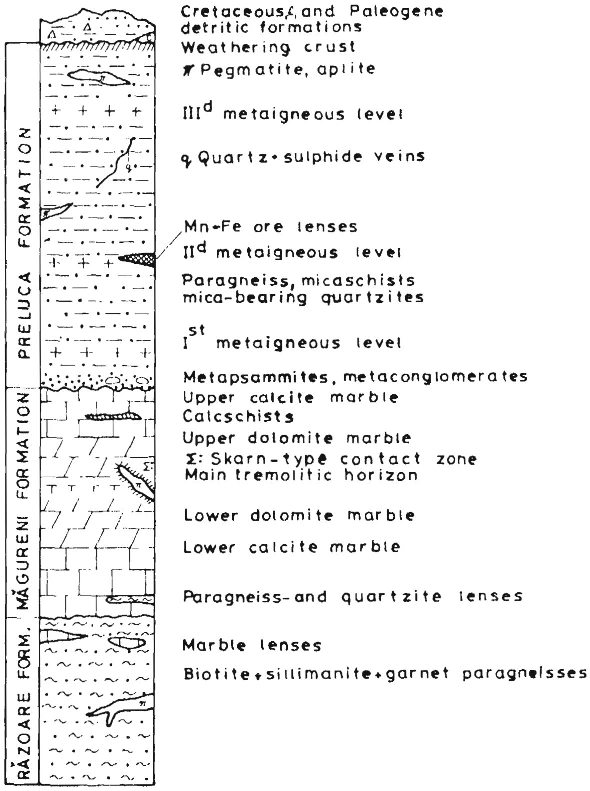

Fig. 3. Idealized lithostratigraphic column in Preluca Mts.

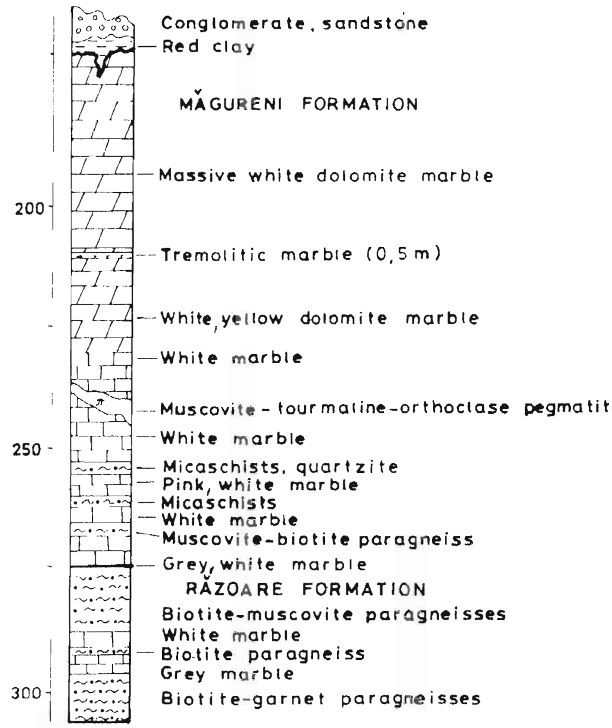

Fig. 4. Deep drilling No. 4, Dealul Crucii Hill, Magureni. It crosscuts the lowermost part of the Magureni Formation and its boundary with the Razoare Formation.

est unit is the Preluca Quartzite-Amphibolite Formation. It originally had an unconformity relationship with the underlying Magureni and Razoare Formations (Figs. 2 and 5).

The rock types

A short petrographic description of the three metamorphic formations is given here (see the lithostratigraphic column in Fig. 3).

The Razoare Gneiss Formation

The main rock type is a slightly folded, banded paragneiss showing an alternation of quartz + feldspar and mica-rich layers, 2 to 10 cm thick. White, folded quartz lenses are frequent. Massive intercalations, with a thicknes of tens of centimetres, of mica-poor gneiss can be observed in some places. Subordinate biotite-bearing quartzites occur in the whole thickness of the Razoare Formation. Calcitic marbles with a thickness of a few metres occur at its highest levels.

The following minerals occur in the paragneisses: quartz, plagioclase (An-27), potash feldspar, biotite, muscovite, garnet, and sillimanite. Accessorial components are as follows: tourmaline, zircon, apatite, titanite and opacitic minerals.

The thickness of the Razoare Gneiss Formation exceeds 1 000 m.

The Magureni Carbonate Formation

The Magureni Carbonate Formation consists of calcitic and dolomitic marbles, with a few intercalations of quartzites, gneisses, tremolitic marbles and biotite-muscovite marbles or carbonate schists. Five horizons can be distinguished therein which are as follows: (1) Lower calcitic marble with quartzite and gneiss intercalations; (2) Lower dolomitic marble; (3) Main tremolitic horizon: a 0.2 to 1.0 m thick tremolite-rich dolomitic marble of a regional extension; (4) Upper dolomitic marble here and there including small amphibolite lenses; (5) Upper calcitic marble with intercalations of muscovite and biotite marbles and calc-schists (Fig. 3).

The boundary between the Rázoare and Magure-

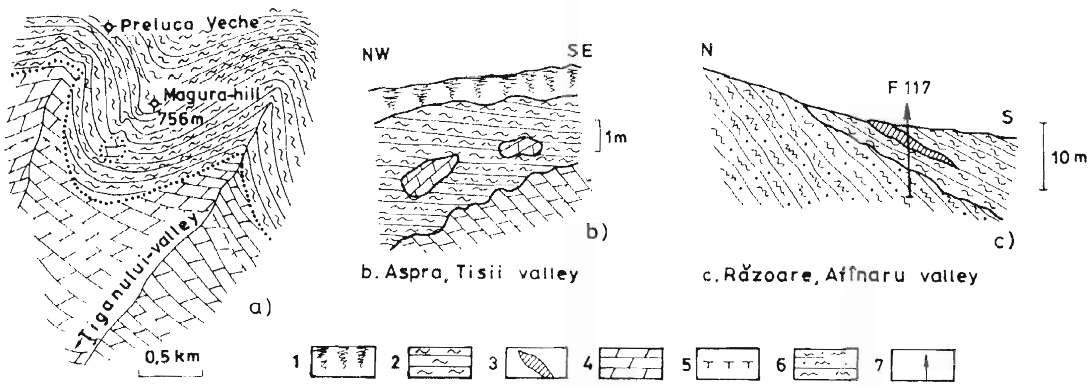

Fig. 5. Unconformity between Preluca and Magureni (a, b) and Preluca and Razoare (c) Formations, respectively. 1 - Soil, 2 and 3 - Preluca Formation: 2 - quartzite, micaschist, paragneiss, 3 - marble, 4 and 5 - Magureni Formation: 4 - dolomitic marble, 5 - "Main Tremolitic Bank", 6 - Razoare Formation: biotite+garnet+sillimanite gneisses, 7 - Borehole.

ni formations is transitional, with marble lenses appearing in the upper part of the Razoare Formation, and quartzitic intercalations in the lower part of the Magureni Formation (Fig. 4). Schistosity is generally parallel with the bedding and is conformable in both formations.

The mineralogical constituents of marbles are as follows: calcite, dolomite (up to 99 % of the rock), quartz, muscovite, phlogopite, tremolite, apatite and graphite.

The thickness of the Magureni Carbonate Formation is 700 to 800m.

Preluca Quartzite-Amphibolite Formation

The highest unit - the Preluca Quartzite-Amphibolite Formation - consists of an alternation of mica-bearing quartzites, micaschists and paragneisses (i.e. terrigenous rocks) containing three levels of metamorphic igneous rocks and some marbles.

The characters and directions of schistosity in this formation differ from that of the underlying formations. It contains, at some places, dispersed or abundant pieces of the underlying formations strongly varying in size (0.2 to 6 m). Their schistosity do not conform with that of the enclosing metamorphic rocks.

The main rock types of terrigenous origin in the Preluca Formation are: quartzites, mica-bearing quartzites, paragneisses, muscovite and muscovite-biotite micaschists, garnet-bearing micaschists. Their rhytmic appearance is caused by the alternation of mica and quartz (+/- feldspar) layers. The mineralogical constituents are: quartz, plagioclase (An 20 - 30(, potash feldspar, muscovite, biotite, garnet, staurolite, kyanite, sillimanite in the higher part of the sequence; and quartz, albite, muscovite, biotite, chlorite, garnet, in the lower part of the sequence (Buteasa village).

The metaigneous rocks show a great variety: serpentinites, amphibolitic rocks, hornblende or ferrotremolite felses, actinolitites, pyroxene and acidic orthogneisses (biotite-K-feldspar gneisses, muscovite-plagioclase-microcline "aplitic" gneisses).

All of the above mentioned rocks of magmatic origin are associated with garnet - bearing micaschists, graphitic quartzites and pure calcitic marbles.

The mineral components of the amphibolitic rocks are: amphibole (hornblende , ferrotremolite, hastingsite, actinolite), pyroxene (augite), plagioclase (An 25-40), biotite, garnet, zoisite, quartz, titanite, rutile, ilmenite, magnetite.

In the southwestern part of Preluca Mts. (Prislop and Buteasa villages) epidote and chlorite appear, and the amphibole is only represented by ferrotremolite.

As concerns the acidic orthogneisses, the following minerals occure in them: quartz, K-feldspar, plagioclase (frequently chessboard albite), biotite, muscovite, tourmaline, zircon.

Graphitic quartzites consist of elongated toothlike quartz crystals showing a preferred orientation, dispersed muscovite and zoisite crystals, graphitic pigment and pyrite idioblasts. These graphitic quartzites contain Mn-ore lenses in the eastern part of the area.

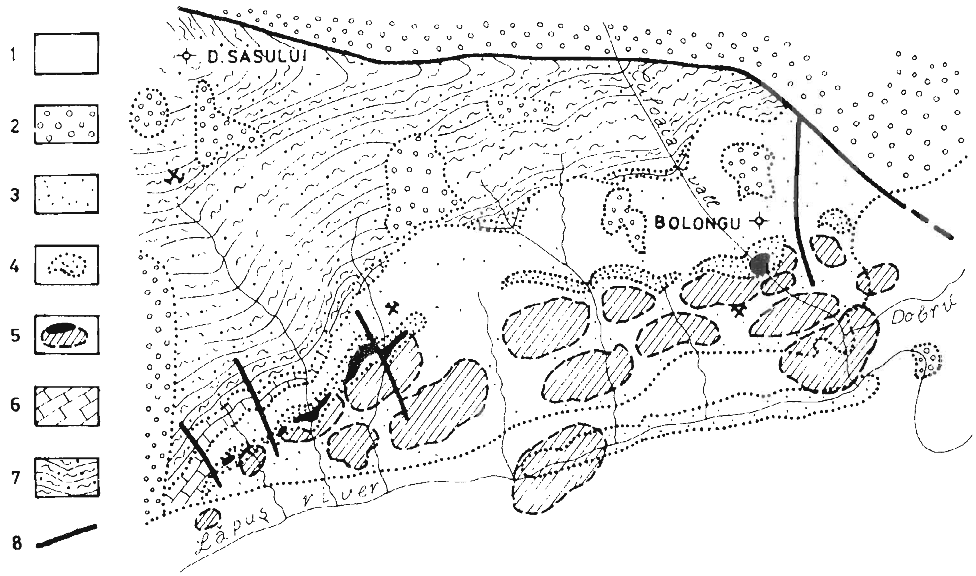

Fig. 6. Geological sketch of the Razoare Mn-ore deposit. 1 - Quaternary sand and pebbles, 2 - Eocene: red breccia, clay and sandstone, 3 to 5 - Preluca Formation: 3 - alternation of micaschists, paragneisses and quartzites, 4 - graphitic quartzites, 5 - Mn-ore outcrops and their deep limit, 6 - Magureni Formation: marbles, 7 - Razoare Formation: biotite gneisses, 8 - faults.

Carbonate rocks are pure calcitic marbles (locally graphitic) and impure calcitic-sideritic marbles containing detritic quartz, and lithic fragments of quartzites and gneisses.

This series shows a parallel, locally microfolded schistosity.

The post-metamorphic processes

Postkinematic pegmatite and aplite bodies are known in all three formations. The following minerals make up these bodies: quartz, plagioclase, potash feldspar, muscovite (a 40x40 cm large plate was found in Ciungi mine, Razoare), biotite, tourmaline, garnet, apatite, berill. A skarn-type reactive rim occurs at the boundary between the pegmatite body and the host rocks.

The metamorphic and pegmatitic mineral assemblages of the Preluca Mts. underwent a later hydrothermal alteration and weathering during Paleogene.

Age data

As corncerns sedimentation age, no data are available while some data concerning the age of metamorphic processes are available.

Some Pb-Pb age data were supplied by Popescu (1981, unpublished data, C.N.C.I.N., Magurele, Bucharest): 1.3 ± ∅.3 Ga, obtained from the lower gneisses of the Mn ore, Stefan lens, Razoare. Some mica K-Ar data were obtained from gneisses, micaschists and amphibolites of the Preluca Formation and from pegmatites (Soroiu, unpublished data, C.N.C.I.N. Magurele, Bucharest, Romania): they are scattered in the range 90 to 100 Ma.

Biotite and muscovite K-Ar ages have recently been determined from the bottom gneiss of the Afinaru Mn ore lens, Rázoare (Balogh, pers. comm. 1990): biotites gave age values of 98.8 ± 3.7 Ma and 98.9 ± 3.7 Ma, whereas muscovite from the same rocks gave 100.4 ± 3.8 Ma.

The Mn-ore deposit

Mn-ore bodies are known in the eastern part of the Preluca Mts., near the locality of Razoare, Tirgu-Lapus, within the Preluca Formation. The Mn ore bodies are intercalated within a level of graphitic quartzites which always accompany, according to field and deep drilling data, to the middle (or II) metaigneous level of the preluca Formation (Fig. 6).

The Mn ore makes up lenses which are 2 to 45 m thick, 30 to 200 m wide and 50 to 300 m long. They occur either as single thick lenses or elongated bodies divided into two levels by graphitic

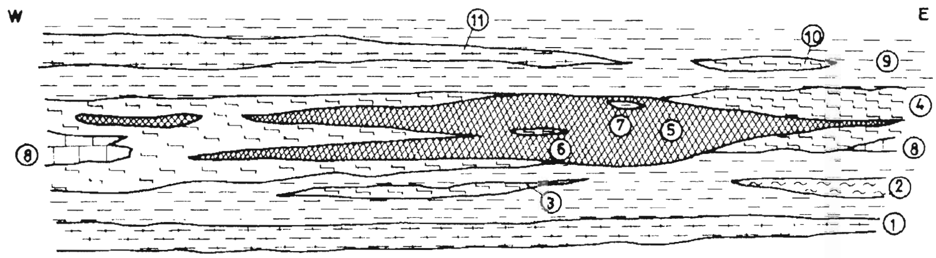

Fig. 7. Simplified cross-section along a Mn-ore lens. 1 - lower biotite gneiss, 2 - garnet-graphite micaschist, 3 - lower black quartzite (satellite lenses), 4 - main graphitic quartzite level, 5 - Mn-ore, 6 - quartzite intercalations, 7 - micaschist intercalations, 8 - marble, 9 - mica-bearing quartzite with micaschist and paragneiss intercalations, 10 - upper black quartzite (satellite lenses), 11 - upper biotite gneiss.

quartzites. The most important lenses containing 90 % of ore named Stefan, Borta, Zapodia, Arghir, Augustin, Hotarului, Afinaru, Cufoaia, Bolongu, Dobric, Pavel, Ursu and Lapus (Fig. 7).

Mineralogical constituens

Primary metamorphic minerals of the Mn-ore are as follows:

Oxides: magnetite, jacobsite, quartz;

Sulphides: pyrite, pyrrhotine, chalcopyrite;

Carbonates: calcite, rhodochrosite, kutnahorite;

Phosphates: apatite;

Silicates: knebellite, pyroxmangite, bustamite, rhodonite, spessartine, dannemorite, piemontite.

Secondary minerals of hydrothermal and weathering origin are as follows:

Oxides, oxi-hydroxides: pyrolusite, hematite, braunite, psilomelane, goethite, hydrohematite, waad, ochre, limonite gel, quartz, chalcedony, opale;

Sulphides: pyrite, marcasite, chalcopyrite, covelline, chalcosine, sphalerite, galenite, alabandine;

Carbonates: calcite, manganocalcite, oligonite, siderite;

Phosphates: vivianite;

Sulphates: gypsum, szmikite, chalcantite;

Silicates: chlorite (delessite, diabantite), serpantine-like minerals (neotocite, hissingerite), montmorillonite, sericite.

The distribution of minerals in the ore bodies follows some rules:

- silica-rich minerals and quartz preferentially appear at the bottom, top and extremities of the lenses, near the quartzitic host rock;
- on the contrary, carbonates can be found mainly in the middle part of the lenses, in association with silica-poor silicates (mainly knebellite), apatite and spinels;
- dannemorite is frequently associated to garnet, pyroxene and fine-grained green, pea-like apatite (so-called Bohnen-apatite, Kossmath & John, 1905);
- minerals of hydrothermal origin are localized along joints and breccia zones;
- oxidized manganese and iron minerals, amorphous and cryptocrystalline silica and clay minerals make up the Paleogene weathering crust.

Textural and structural features

The main textural types of the ore are:

- massive, coarse-grained: the crystals do not show any preferred orientation;
- banded: carbonate and oxide layers alternate with silicate layers; the thickness of these layers varies between 1 and 5 mm;
- schistose: it is defined by the parallel orientation of dannemorite prisms;
- breccia-like: primary Mn ore fragments are cemented by a secondary matrix (carbonates and chlorite).

Genetic considerations

The Mn ore bodies are located in a volcano-sedimentary sequence. Therefore, they are genetically related to ancient magmatic processes which took place during the sedimentation of the protoliths of the Preluca Formation.

The ores originated from hydrothermal solutions rising up along preferential way and precipitated at the sea bottom. The precipitates, mainly silica and temporarily Fe and Mn (oxides, carbonates, hydrosilicates) built up lenticular submarine accumulations. These accumulations covered by sediments evolved together with them forming the actual metamorphic ore bodies.

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Stratigraphic, metamorphic and deformational unconformity between the Variscan and Precambrian metamorphics of the Carpathians

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Abstract

A model is proposed for the metamorphic and structural relationships resulted in the case when a medium grade metamorphic basement (e. g. Bretila Group - Precambrian) and its sedimentary cover (Repedea Group, Rusaia Group - Silurian) have been affected by a younger (Variscan) low grade metamorphic event. Arguments are based on the detection of a primary stratigraphic unconformity, a corresponding metamorphic unconformity, and a deformational unconformity placed within the overprinted basement.

General Statements

In the European Alpine Belts the relationships between low grade and medium grade metamorphics are the subject of controversial opinions. Four main models have been proposed referring to prograde transition, retrograde overprint, stratigraphic and metamorphic unconformity and to tectonic unconformity (Fig. 1). In the relevant literature there are examples of convincing application for all four models. In the field, between the low grade and medium grade metamorphics, in most cases (except of fault contacts) a "transitional zone" is interposed, which may confuse the real geological relationships.

The growth of new medium grade minerals or the development of new low grade parageneses on relict higher grade minerals may be relevant for the transitional zones of models 1 and 2 (Fig. 1). However, this rule may not be conclusive in some cases. Thus, the same low grade minerals (chlorite, sericite, albite, zoisite, calcite, quartz) develop on medium grade minarals ("relicts") during the low grade overprint on as wellas "medium grade minerals" as clasts (metasandstones with relict biotite, garnet, staurolite, amphibole etc.) or as magmatic mineral components (relict biotite, amphibole, plagioclase in metagabbros, metadiorites, metadacites etc.). In such situations the preserved structural aspects of the protolith and/or data on its age may be relevant for a correst interpretation. Difficulties may appear also in distinguishing between models 3 and 4 (Fig. 1) as in both situations the "transitional zone" consists of retrogressive rocks with an overprinted schistosity.

Some relevant examples for model 3 of stratigraphic and metamorphic unconformity from the metamorphics of the East Carpathians will be discussed in the following.

Pre-Alpine Metamorphics in the East Carpathians: Geological Setting

The main part of the metamorphic rocks in the East Carpathians belong to pre-Alpine cycles (Variscan, Early Caledonian, Precambrian) (Fig. 2) (Kräutner, 1988). They are included in the "Central East Carpathian Nappes" (Median Dacides) and overthrust the flysch zone (External Dacides) (Sandulescu, 1984). Three main alpine nappe systems may be distinguished: Bucovinian, Sub-Bucovinian, Infrabucovinian. In the Bucovinian and Sub-Bucovinian nappes mainly Early Caledonian and Precambrian sequences are known (Fig. 2). In lower tectonic position and in a more external position of the Carpathian Orogen Precambrian medium grade metamorphics (Bretila Group) are overlain by Variscan low grade metamorphic sequences (Repedea Group, Rusaia Group) (Fig. 2).

The above mentioned alpine nappes include fragmentary Variscan nappe structures, as the main lithostratigraphic units of the metamorphic rocks

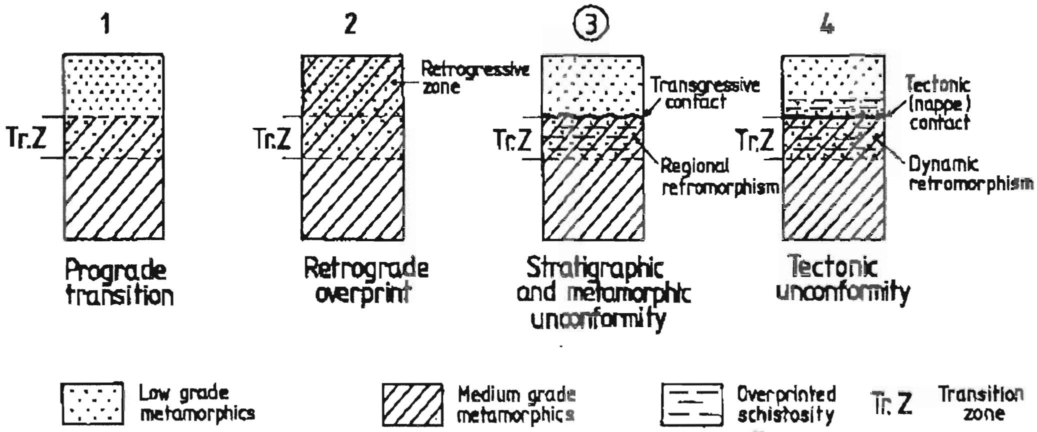


Fig. 1. Schematic representation of the relationships between low grade and medium grade metamorphics in the Carpathians.

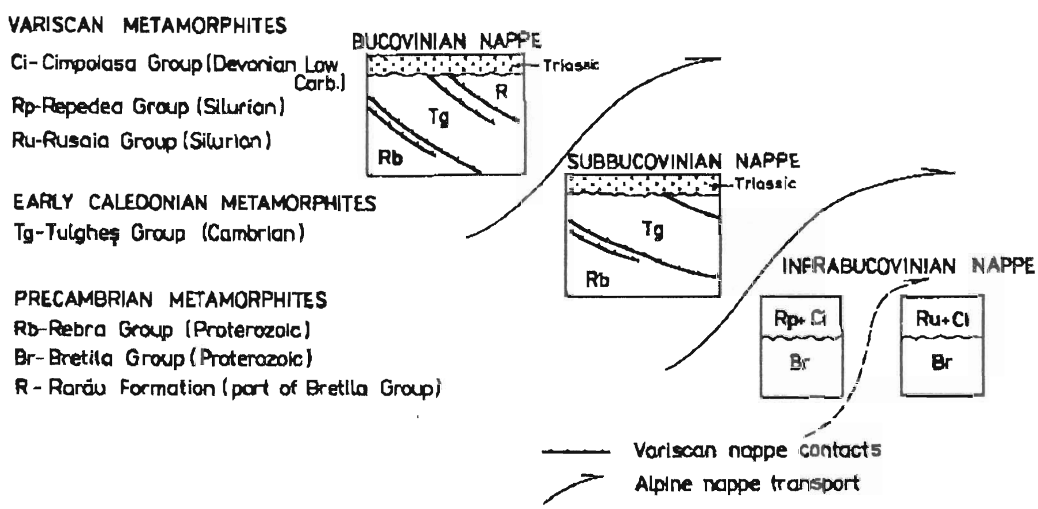


Fig. 2. Pre-Alpine metamorphics of the East Carpathians: the main lithostratigraphic units and their tectonic position.

are in nappe relations, prior to the Lower Triassic (Fig. 2) and to the intrusion of the Ditrau Massif (135 Ma). Tectonic nappe relationships (model 4 of figure 1) prevail among most of the metamorphic sequences in the East Carpathians.

Large zones of dynamic retromorphism are associated with both Variscan and Alpine overthrust surfaces extending about 10 - 200 m above and below the tectonic contacts. In some areas these dynamic retromorphisms overprint rocks with Variscan regional retromorphism leading to the formation of polyretrogressive rocks. In these schists a late kinking of the Alpine dynamic schistosity (lamination) may be observed.

"Transgressive" positions of low grade metamorphics on older medium grade metamorphics (model 3 of figure 1) are known mainly in the Infrabucovinian Nappes (Fig. 2). This stratigraphic unconformity is marked also by a metamorphic discontinuity as it will be discussed in the following.

The Primary Sedimentary Unconformity

In the Infrabucovinian Nappes Silurian sequences with Variscan low grade metamorphism (Repedea Group, Rusaia Group) overlie Precambrian medium grade metamorphics (Bretila Group). These relationships extend on regional scale and may well be observed at the Rodna half Window (Rodna Mts.) and at the Rusaia Window (Bistrita Mts.) (Kräutner et al., 1978, 1982, 1983, 1990).

The Silurian age of the Repedea and Rusaia Groups is well documented by Chitinozoa, acritarchs, spores, and the Devonian-Lower Carbonif-

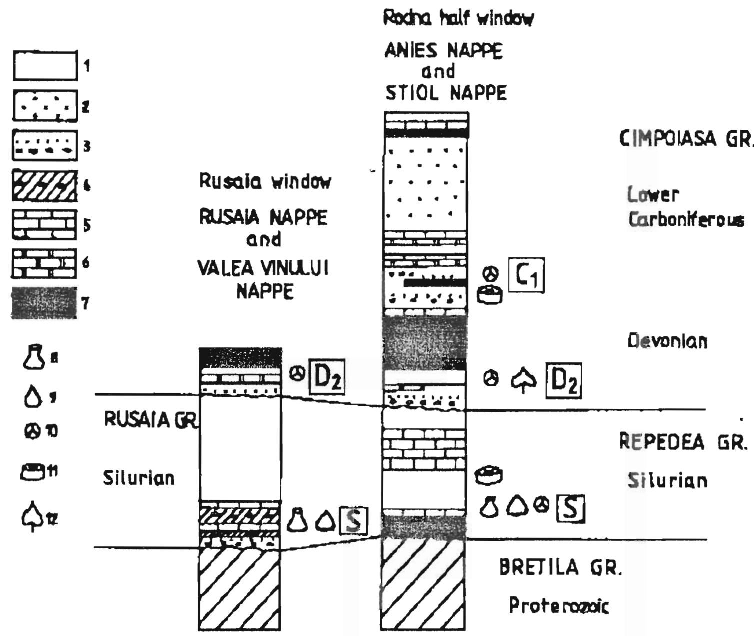


Fig. 3. Lithostratigraphic sequences of the Variscan low grade metamorhpic sequences in the Rodna half Window and the Rusaia Window. 1 - metapelites, 2 - metagreywacke, 3 - quartzites and metaconglomerates, 4 - quartzose metaconglomerates, graphitic schists with detrital muscovite, carbonatic schists, 5 - limestones, 6 - dolomites, 7 - greenschists (basic metavolcanics), 8 - Chitinozoa, 9 - acritarchs, 10 - spores, 11 - crinoids, 12 - plants.

erous age of the Cimpoiasa Group by spores (Iliescu et al., 1975, 1976, 1978; Kräutner & Mirauta, 1970). The lithostratigraphic sequences and the stratigraphic position of the palynological data are schematically shown in figure 3.

In the Bretila Group most of the radiometric ages are evidently rejuvenated by the Early Caledonian, Variscan and Alpine events: Rb-Sr whole rock isochrons of 529 ± 9 (Gorohov et al., 1977), 330 +- 35 Ma (Gorohov et al., 1967); U-Pb zircon ages of 300, 540 Ma; Boiko et al. (1975): K-Ar ages of 748, 552 - 126 Ma (Kräutner et al., 1988). This rejuvenation is proved by the fact that, according to the geological relationships, the Bretila Group is older than the Silurian and cannot be Cambrian in age because in the East Carpathians the Cambrian is lithologically definitely different (Tulghes Group) and proved by palynological data and radiometric ages (Iliescu et al., 1983). Therefore the Precambrian age of the Bretila Group may be inferred from both geological relationships and remanent radiometric ages (K-Ar model ages of 522, 552, 748 Ma, as well as by interpretation of K-Ar isochron ages of 850 ± Ma; Kräutner et al., 1976; Kräutner, 1988).

The primary relationships between the two mentioned sequences of Variscan and Precambrian metamorphics is believed to be of transgressive character. This stratigraphic unconformity is related to the beginning of the sedimentation in the Variscan cycle. The following arguments are in favour of these presumed relationships:

1. Age relationships. A Silurian sequence covers on regional scale a Precambrian basement after a stratigraphic gap including the Cambrian and the Ordovician.

2. Angular unconformity, observable on regional, outcrop and hand specimen scales.

2.1. On regional scale the basal horizons of the Silurian Repedea and Rusaia sequences cover different lithologic and lithostratigraphic units of the Precambrian Bretila basement (Fig. 4 A).

2.2. On the scale of outcrops the angular unconformity may be observed only in a few favourable places, due to the fact that the Precambrian rocks are usually strongly overprinted by a Variscan schistosity in parallel position to the unconformity (Fig. 4 B).

2.3. On the hand specimen scale the relationships are usually similar to those at 2.2, but in the Rusaia Valley (in galleries) a clear angular unconformity was observed in boudins without overprint

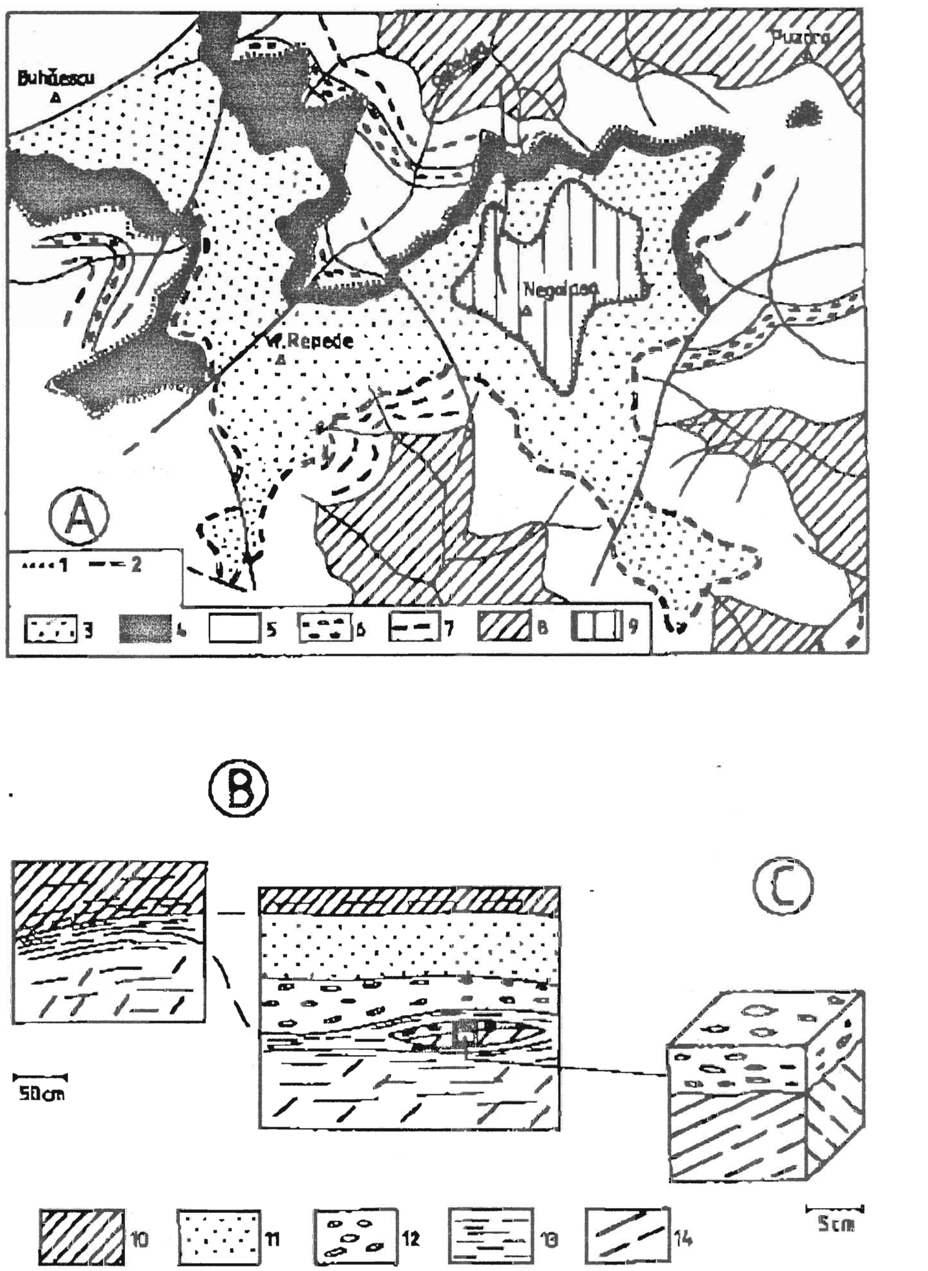


Fig. 4. The primary stratigraphic unconformity. Examples on regional scale from the Rodna Mountains (A), outcrop scale (B) and hand specimen scale (C). 1 - primary stratigraphic unconformity, 2 - primary contact disturbed by slipping of the cover, 3-4 - Repedea Group (Silurian): 3 - Fintina Formation - blastodetrital with carbonatic sequences, 4 - Stiol Formation, basic volcano-sedimentary, 5-9 - Bretila Group (Precambrian), 5-7 - Lespedea Gneissic Formation: 5 - retrogressive paragneisses, 6 - retrogressive augen gneisses, 7 - retrogressive leptynitic gneisses, 8 - Mireaja amphibolite-gneissic formation, 9 - Negoiasa gneisses, 10-12 - Rusaia Group (Silurian): 10 - graphitic schists with detrital muscovite, quartzose metaconglomerates, carbonatic schists, 11 - Stinisoara quartzites, 12 - basal metaconglomerate with carbonatic matrix, 13-14 - Bretila Group (Precambrian): 13 - Variscan foliation (schistosity) in retrogressive gneisses, 14 - Precambrian foliation in retrogressive gneisses.

(Fig. 4 C).

Arguments 1, 2.1 and 2.2 cannot be considered as conclusive by themselves because they are not incompatible with tectonic nappe relationships. Argument 2.3 is decisive, being the only case in which the primary relationships were not disturbed by the Variscan schistosity. However, such situations are rare, as the Variscan regional metamorphism was highly penetrative in the basement.

Due to the different tectonic competence between the cover sediments and the gneissic basement, the stratigraphic unconformity acted as a surface of preferential movement - decollement. This is proved by the intensive schistosity and lamination in the basement and cover rocks as well as by the local link of some basal terms of the Paleozoic sequences (Fig. 4 A, 4 B). This suggests that during the Variscan deformation and metamorphism the Paleozoic cover was locally displaced and slipped over its basement. In these conditions some lithons of high tectonic competence may be sheared and detached from their initial position and included as boudins in the laminated contact zone (case 2.3, Fig. 4 B). Such lamination zones and displacement of the metamorphic Paleozoic cover usually extend over large areas. This fact causes the main difficulties in a correct interpretation of transgressive or tectonic relationships between low and medium grade metamorphic sequences (models 3 or 4 of figure 1).

3. Metallogenetic processes related to stratigraphic unconformities. In the Rusaia Valley, in the basal conglomerates of the Rusaia Group, magnetite, hematite and iron jaspilitic ores occur (Rusaia iron ore deposit). Based on their geological environment, structural-textural aspects and mineralogical constitution, the iron ores were considered as metamorphosed stratiform ores, deposited from local submarine sources (Kräutner, 1967). This depositional model is proved by "feeding zones" (Fig. 5), observed during the mining of the ore. They consist of magnetite disseminations in retrogressive gneisses of the Bretila basement, and are situated under the stratiform lenses intercalated in the basal conglomerate. The mentioned magnetite disseminations form zones disposed parallel to the relict Precambrian schistosity of the gneis.

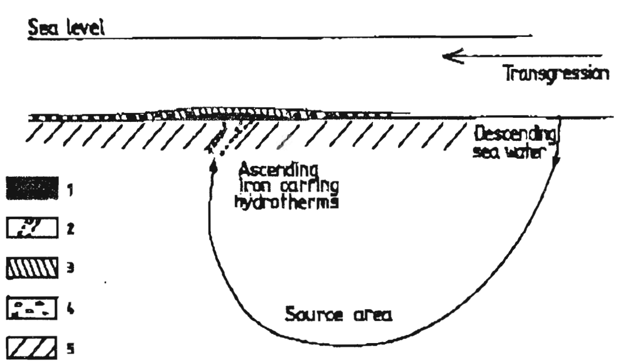


Fig. 5. Genetic model for the stratiform iron ores of the Rusaia deposit (prior to the Variscan metamorphism): A convectice cell with sea water started by the Silurian transgression. 1 - Silurian stratiform iron ore, 2 - disseminated magnetite ore (feeding zones), 3 - Silurian basal conglomerates with iron bearing carbonate matrix, 4 - Silurian basal conglomerates, 5 - basement - Precambrian gneisses of the Bretila Group.

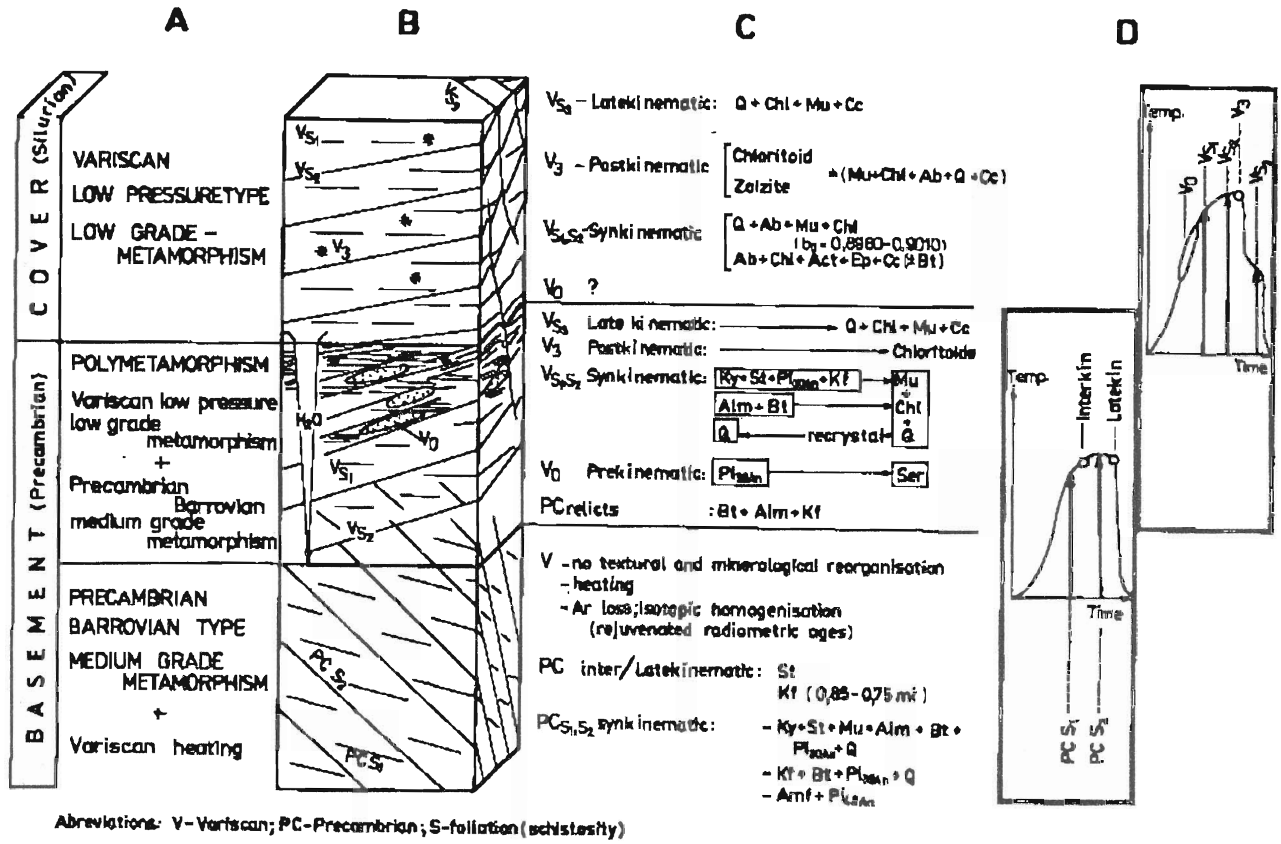
The proposed genetic model (Fig. 5) for the primary (premetamorphic) ores suggests a convective cell, fed by the sea water, and formed as a result of the Silurian transgression. The descending sea water is charged at depth by iron, carried to the surface as hydrotherms. Depending on the constitution of the hydrotherms and to the local conditions (Eh, pH) in the sedimentation area, massive oxidic deposition or iron-silicium gel deposition occurred round the submarine sources. A distant effect of this process is the iron content of the carbonatic matrix of the basal conglomerates.

The Metamorphic Unconformity

The described stratigraphic unconformity implies a corresponding metamorphic disconformity: in the cover a prograde low grade metamorphism, and in the basement a prograde Precambrian medium grade metamorphism, regressively overprinted by the metamorphism which affected also the Paleozoic cover. Therefore the basement has a polymetamorphic character.

In the Infrabucovinian units of the East Carpathians mineralogical and textural-structural reorganizations due to the mentioned overprint penetrate only in the upper part of the Precambrian basement to a depth of about 300 - 500 m below the unconformity surface. This restricted extent of the overprint may be explained by the depth to which, during the Variscan metamorphism, water from the sedimentary cover could penetrate in the Precambrian gneissic basement. Below this P_{H_2O} limit no plastic deformations and mineralogical changes are possible in the older medium grade metamorphics due to younger P-T conditions of the greenschist facies. Only brittle deformations and Ar loss or isotopic homogenizations due to heating may be observed.

The three mentioned "metamorphic zones" are clearly exposed in the pile of the Infrabucovinian metamorphics (Fig. 6). Details of the metamorphic



,Fig. 6. Schematic representation of the pre-Alpine metamorphic features in the Infrabucovinian nappes: A - sequence of metamorphisms, b - "metamorphic zones" in the pile of crystalline schists, C - mineral parageneses related to deformation phases, D - evolution of temperature, deformation and mineral growth during the metamorphic events.

features of each zone are indicated in figure 6. In the transitional zone no continuous gradual change of the metamorphic grade may be accepted because of the following facts: 1) the low grade parageneses grow over older medium grade minerals, preserved as relicts in the rocks (Fig. 6); 2) the medium grade metamorphism is of Barrovian type (with kyanite), while the low grade metamorphism is of low pressure type (b_0 of K-white micas in metapelites = 8.980 - 9.010 A; Kräutner et al., 1975).

Booth Precambrian and Variscan metamorphics were developed in multiphase process with successive mineral growths on more than one schistosities and also during interkinematic phases (details in Fig. 6). A peculiarity of this transitional zone is that the Variscan schistosities (S_{V1}, S_{V2}) are not thoroughly penetrative as they are for example in the Paleozoic cover. Thus restricted zones, lense-shaped or boudin-like, without Variscan schistosity, are preserved. In such zones the primary gneissic texture is preserved, and only restricted mineralogical reorganizations inside the mineral components are observable: A progressive to complete transformation of the plagioclase into a sericite (muscovite, paragonite) agglomerate with massive texture may be noticed, while all the other components such as biotite, almandine (except partly K-feldspar) are fresh, not affected by any change. Mineroscopical evidence indicates that these pseudomorphoses of sericite on plagioclase, in subsequent stages are stretched by the Variscan schistosities. Therefore, it seems that the Variscan metamorphism started by a static phase (V_o, Fig. 6) in conditions of heating and increasing P_{H_2O}. In the Paleozoic cover such relict stages may not be observed due to the complete and homogeneous penetration of the V_{S1} and V_{S2} schistosities. Mineralogical reorganizations on biotite, almandine and part-

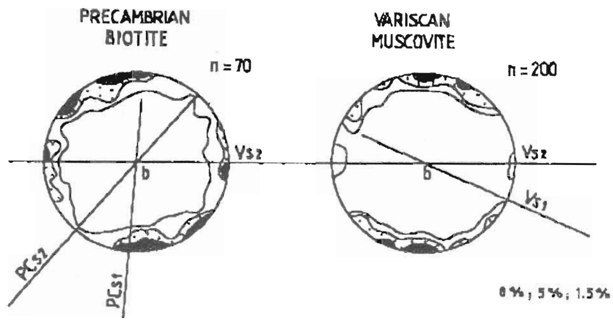


Fig. 7. Partial fabrics of Precambrian biotite and Variscan muscovite in a sample of retrogressive Bretila Gneisses (Rodna Mts.). Orientation of (001) poles.

ly K-feldspar appear only in the synkinematic stages V_{S1} and V_{S2}. Therefore, it seems that the retrogressive destruction of these minerals requires a deformational environment.

Statistical investigations on phyllosilicates have shown that in the retrogressive Bretila Gneisses relict biotite may preserve its Precambrian fabric, while in the same sample the Variscan muscovite (sericite) grows on the schistosities S_{V1} and S_{V2}(Fig. 7).

A late Variscan post-deformational mineral growth (V_3 in Fig. 6) is marked by an unoriented (rosette-like) development of chloritoid, both in the Paleozoic cover and in the retrogressive gneisses of the "transitional zone".

Macroscopically some retrogressive rocks of the "transitional zone" may be easily confused as Paleozoic low grade schists. This explains why, in the past, such rocks have been mapped as progressive low grade schists.

The Deformational Unconformity

According to the different evolution of metamorphism, significant differences in the synmetamorphic deformation may be expected in the cover and in the basement. In the Precambrian basement the Variscan deformation is restricted only to the zones of Variscan mineral reorganization. Therefore, it penetrates only in the volume of retrogressive overprint. In deeper zones of the basement the Precambrian structures were preserved undisturbed. In the overprinted zones the Variscan deformational overprint was so intensive that over large areas the Precambrian structures have been completely obliterated. This explains why in older maps these zones have been included in the progressive "epimetamorphic areas". Now it is evident that the expected deformational unconformity is placed inside the Precambrian basement, some hundreds of metres below the primary stratigraphic unconformity, near the lower limit of the retrogressive mineralogical reorganizations. This "deformational unconformity" represents in fact a larger zone of structural interferences that separates the mostly unaffected Precambrian deformations from the various structural aspects of the overprinted Variscan deformational zone.

Figure 8 proposes a model showing the main aspects of the pre-Alpine synmetamorphic deformations in the cover and in the overprinted basement. The following domains can be distinguished (the numbers correspond to those of Fig. 8:).

(1) Deformation of the Variscan cover. Three main deformational phases may be recognized, like in other Variscan low grade metamorphic piles (e. g. Poiana Rusca, Kräutner et al., 1981):

a) plastic folding concomitant with the development of the V_{S1} foliation in subparallel position to the main lithologic elements of the primary sedimentary sequence;

b) continuation and development of the deformation by gliding in the V_{S2} foliation in appropriate (but slight angular discordance) position with respect to the axial cleavage of the V_{S1}(B1) folds. V_{S2} cuts V_{S1} at different angles according to its intersection with the V_{S1} (B1) fold structure;

c) late kinking V_{S3} (pre-Permian) in oblique position to the former deformational elements.

(2) Deformation of the Precambrian (Bretila) basement. Two main phases may be observed, marked by two visible foliations: PCS_1, locally intensively folded, and PCS_2, the main schistosity involved in large scale folding.

(3-7) Deformations in the overprinted (transitional) zone are inhomogeneously developed, including parts with relict Precambrian structures as well as parts with complete Variscan obliteration. Five main types of deformational overprint have been selected in figure 8. In reality there are all forms of transition and mixing between these fundamental types.

(3) Schistose slipping zones between cover and

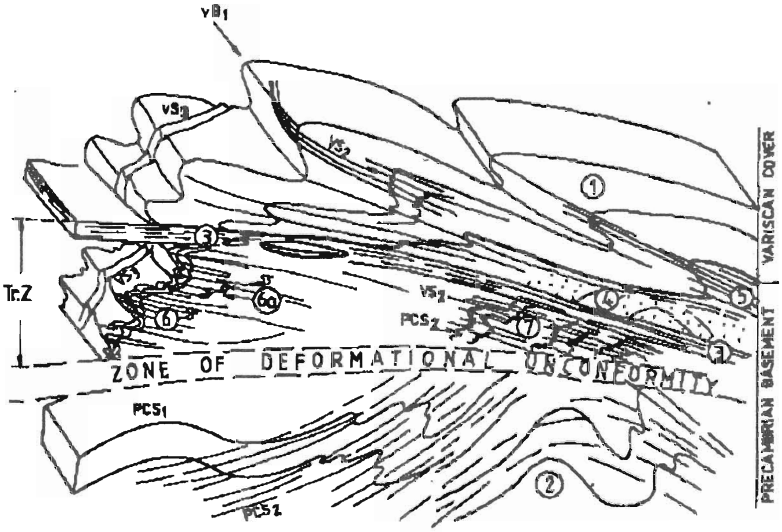


Fig. 8. Model for the successive synmetamorphic deformations in the pile of Infrabucovinian metamorphics. Different aspects of the overprint zone (Tr. Z.) are selected and distinctively represented. 1 - Variscan deformation of the Paleozoic cover (V_{S1}, V_{S1}, V_{S3} - successive foliation systems), 2 - Precambrian deformation of the Bretila basement (PCS_1, PCS_2 - successive foliation systems), 3-7 - Deformational types in the overprinted zone: 3 - Slip zone of the cover, including boudins from the basement and from the cover, 4 - Undeformed retrograde basement with Vo stage of the Variscan overprint, 5 - Total obliteration by V_{S1}, 6 - Complete mineralogical and structural reorganization by V_{S1} and V_{S2}. Only locally deformed relict lithons of the primary lithologic elements were preserved (6a), 7 - Intensive penetration and deformation of PCS_2 by the Variscan V_{S2} foliation.

basement or inside them. Local phyllonitization and boudinage (see also Fig. 4) suggest that this deformation developed late in the Variscan event.

(4) Well preserved Precambrian structures with Vo type retrograde overprint (Fig. 6) and scarce penetration of Variscan foliations.

(5) Intensive development of the V_{S2} schistosity with complete destruction of all the older structural elements and an entire mineralogical reorganization.

(6) Intensive development of booth folded V_{S1} foliation and V_{S2} schistosity in conditions of an entire mineralogical reorganization. Only locally deformed relict lithons of Precambrian gneisses have been preserved (6 a in Fig. 8). Macroscopically the rocks look like soma low grade metamorphics and only relict muscovite lamellas may indicate the origin from the Bretila Gneisses.

(7) Preserved relict Precambrian structures (mainly PCS_2) are deformed by prevailing V_{S2} foliations.

Concluding Remarks

Detailed studies of the "transitional zones" between low grade metamorphics and medium grade metamorphics seem to be essential for a correct interpretation of the geological relationships between the metamorphic sequences in the European Alpine Belt. Many of the published interpretations based on unilateral documentation proposed confusing and unconvincing conclusions about the evolution of metamorphic and deformational events. In the present study arguments are discussed for one of the models of possible relationships "hidden" inside the mentioned "transitional zones".

In the Infrabucovinian units unconformities regarding geometric (angular) disposition, metamorphic characters and deformational history suggest as primary geological relationships a Precambrian gneissic basement (Bretila Group) and a Silurian sedimentary cover (Repedea Group, Rusaia Group), both involved in the Variscan metamorphic event. In this case the "transitional zone" resulted from the Variscan low grade metamorphic overprint of the Precambrian basement. This model offers a reasonable interpretation for the rejuvenated radiometric ages.

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Further data on the phyllitic Paleozoic sequence with Mediterranean affinity from Bakony Mountains (Hungary)

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Abstract

Petrographic data concerning the phyllitic Paleozoic sequence of the Bakony Mts. are presented, also based on rock samples from boreholes. The P-T conditions of the Variscan metamorphism in these rocks are estimated from the mineral compatibilities, values of the "illite crystallinity" and muscovite b cell data. A summary of the b data from other phyllitic areas of West and South Carpathians, Hungary and Eastern Alps is also given.

In the ambit of IGCP Project No. 276 systematic research was developed, aimed at better understanding the Paleozoic lithostratigraphic sequences and evaluating the physical conditions of their metamorphism.

Several small outcrops of a very low to low-grade sequence occur in the southern part of the Bakony Mountains. Sedimentation ages range from Lower Ordovician (Upper Arenigian) to Lower Carboniferous (Viséan), as shown by fossil findings. The whole sequence displays a Mediterranean affinity (Lelkes et al., 1982). Unmetamorphosed Permian sediments cover this phyllitic sequence.

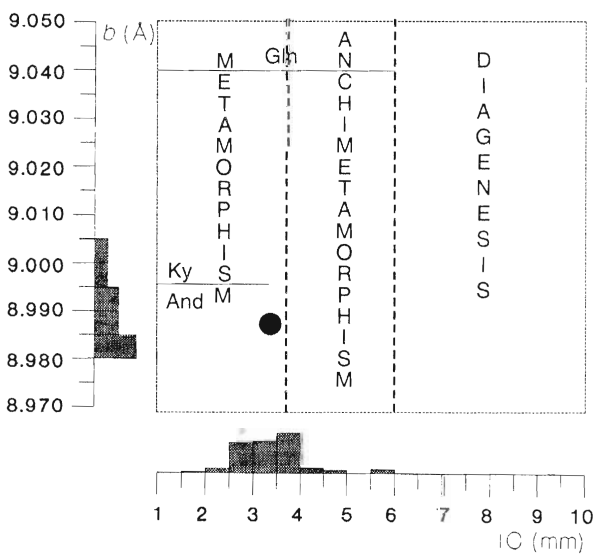

Fig. 1. P-T location of the Variscan metamorphism in the studied rocks, as inferred from data on the "illite crystallinity" (IC) sensu Kubler (1968) and *b* cell parameter (And=Andalusite; Ky=Kyanite; Gln=Glaucophane). In the ellipse: each semiaxis is 1.96s/ √n (where *s* is standard deviation and *n* the number of data).

The more abundant rock type makes up a monotonous sequence of fine-grained metapelites (slates) and metasemipelites, with subordinate intercalations of metasandstones and carbonate beds.

A metavolcano-sedimentary sequence also occurs, covering a large compositional range. The distinction of basic, intermediate and acidic suites was possible by means of petrographic and chemical analyses, but their stratigraphic position and interrelationships are unknown due to the poor exposure of the whole sequence.

Due to this problem, besides the poorly outcropping rocks, the cores of some boreholes were analyzed:

-borehole Szb-9: this crosscuts metapelites with carbonate and acidic metavolcanoclastic intercalations; the analysed rock samples lie above and below the rock level in which Arenigian acritarchs were found; (Albani et al., 1985);

- *borehole Szb-10*: it crosscuts similar lithologies;

- *borehole AÖT-1*: this encounters a volcano-sedimentary sequence with "porphyroids" and carbonate lenses rich in organic content;

- *borehole P-4:* this crosscuts acidic metavolcanoclastics;

- *borehole R-6:* this crosscuts a volcano-sedimentary sequence with intermediate metavolcanoclastics.

TAB . 1
Muscovite b cell data in different localities of the Eastern Alps, Carpatians and Hungary

Rock complex	Source	x	s	n
Eastern Alps (Austridi) Austria and Italy	Guidotti & Sassi, 1986	8.995 A	0.007	159
Eastern Alps (Southern Alps)	Italy Sassi et al., 1974	8.997 A	0.009	161
Fökajar area Hungary	Lelkes-Felvári et al., 1982	8.987 A	0.004	38
Tázlár area Hungary	Arkai & Lelkes-Felvári, 1987	8.998 A	0.007	11
Bakony Mts. Hungary	present paper	8.987 A	0.007	53
Gemericum (West Carpathians) Czechoslovakia	Sassi & Vozárová, 1987	8.997 A	0.003	59
Brusnik anticline (West Carpathians) Czechoslovakia	Mazzoli & Vozárová	8.994 A	0.005	50
Veporicum (West Carpathians) Czechoslovakia	Sassi & Vozárová, 1991	8.997 A	0.006	32
Gemericum (West Carpathians) Czechoslovakia	Sassi & Vozárová, 1991	8.992 A	0.006	55
East Carpathians Romania	Kräutner et al., 1975	8.997 A	0.008	84
Poiana Rusca Massif (South Carpathians) Romania	Kräutner et al., 1976	9.021 A	0.008	129

x - average value, s - standart deviation, n - number of samples

No biostratigraphic data were obtained from the last four boreholes. All rock types were studied from the petrographic point of view in order to ascertain the microtextural features, the mineral composition, the degree of metamorphic recrystallization versus the survived pre-metamorphic mineral and textural features of the metavolcanics.

As regards the P-T conditions of metamorphism, metapelites were preferred to other rocks for geothermobarometric purposes.

Mineral compatibilities were ascertained in about 100 rock samples. Considering that the mineral assemblages in these rocks do not give specific thermic and baric indications, the metamorphic conditions were established by means of the "crystallinity index" of illite (sensu Kubler, 1968) as regards temperature, and b cell dimension of muscovite as regards pressure (Guidotti & Sassi, 1976).

The "crystallinity index" of illite was measured in 110 rock samples on disoriented powder preparates of whole rock. The procedures proposed by Kubler (1968) were followed. 95 % of the values of the Kubler index range between 2.5 and 4.5 mm, clearly indicating conditions at the boundary between subgreenschist (i. e. very low-grade or anchizone) and greenschist facies.

The b cell dimension of K-white mica was measured in about 50 samples having suitable bulk and mineral composition, following the procedure and constraints proposed by Guidotti & Sassi (1976, and references quoted therein).

Analytical b results clearly indicate low-pressure conditions, related to quite high value of the metamorphic thermal gradient (close to 40 °C/km: Guidotti & Sassi, 1986).

The whole set of illite crystallinity and b data allows a definite location of this metamorphism in the P-T space to be established, as shown in Fig. 1.

As clearly detectable from Table 1, this geobarometric result is consistent with the numerous data existing in the literature, obtained by means of the same methodology in other Variscan areas of the Alpine-Mediterranean belt. Similarly low pressure conditions prevailed during the Variscan metamorphism in all these areas, with the only exception of the Poiana Rusca area, where the available b data indicate an intermediate pressure character.

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Alpine metamorphic recrystallization of the pre-Carboniferous metapelites of the Kohút crystalline complex (the Western Carpathians)

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Abstract

In the Kohút crystalline complex, which represents a geological unit situated just on the contact of two geological Western Carpathians units of lst order - the gemeric and the veporic units - pre-Upper Carboniferous metapelites (mica schists and gneisses) occur.

They have been metamorphosed in two metamorphic events. The older one (Variscan) underwent under medium pressure amphibolite facies conditions. Younger (Alpine) is characterized by medium - till high pressure conditions and temperature characteristic for the greenschist till low grade amphibolite facies. Occurrences of kyanite, chloritoide/staurolite and grossularite-rich garnet in metapelites correspond to maximal P-T conditions of the Alpine metamorphic recrystallization.

Introduction

The Kohút crystalline complex (= the Kohút zone) represents the rock filling of the innermost zone of the veporic unit. From the south gemeric unit is thrusted over it. According to some authors Margecany - Lubeník lineament (being the dividing element between veporic and gemeric units) represent the home area of the upper-most subtatric nappes.

Among the rock complexes forming the eastern part of the Kohút crystalline complex, metapelites belong to the most typical. They are represented mostly by mica schists and paragneisses (Bezák, 1988; Méres & Hovorka, 1991). Mica schist predominate in the Ostrá complex; the amphibolite being another member of this unit. Paragneisses form the Klenovec complex (Bezák, 1988). Metapelites under consideration have maximum distribution in the area between Muráň and Kokava (Fig. l).

The geological importance of the discussed geological unit metapelites were recognized by several authors in the past. The opinions on their origin could be summed up into following groups:

1) mica schist are considered to be the product of prograde (Assyntian or Variscan, resp.) metamorphic recrystallization of sediments;

2) mica schist are the product of retrograde recrystallization processes;

3) mica schists are the product of the Variscan metamorphic recrystallization and their mineral assemblage originated in its prograde as well as retrograde branches.

Stratigraphical ranking of the Kohút crystalline complex metapelites is problematic. Zoubek (1936) assumed mica schist to be the product of the Variscan metamorphic recrystallization of the Early Paleozoic sediments of geosynclinal provenience. Later on several authors accepted the Zoubek's view (Klinec & Planderová, 1979; Hovorka et al., 1987 and others). Consequently Máška & Zoubek (1961; in Buday et al., 1961) denoted metamorphic complexes occurring between the Muráň and the Margecany - Lubeník lineaments as Kohút serie and supposed to be the stratigraphical overlier of Early till Middle Proterozoic Tatra serie. Mentioned authors (l. c.) Kohút serie ranked to the Late Proterozoic. Kamenický (1967; in Maheľ et al., 1967) followed mentioned stratigraphy, but he introduced the new name for the Kohút serie - the Kokava serie. Last mentioned author supposed the metamorphic recrystallization of the Kokava serie to be of Assyntian age.

As it is seen from the above presented survey, the opinions on stratigraphy of the Kohút crystalline complex rock sequences are pronouncedly contraversial. It is a consequence of polycyclic metamorphic evolution (Méres & Hovorka, 1989, 1991) of the prevailing part of rock sequences of the geological unit under consideration.

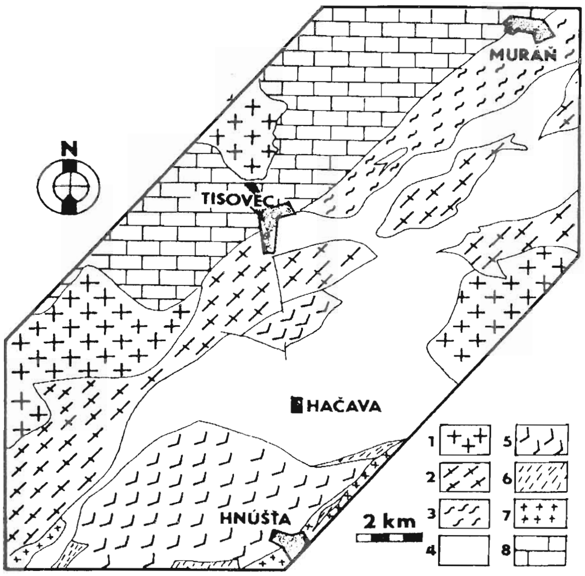


Fig. 1. Geological map of the central part of the Kohút crystalline complex (Compiled according to Méres & Hovorka, 1991). Explanations: Kráľova hoľa complex: 1 - hybrid granitoids, 2 - migmatites; Ostrá complex: 3 - the Muráň gneisses, 4 - mica schists; Klenovec complex: 5 - gneisses; Sinec complex: 6 - chlorite-muscovite phyllites; Rimavica complex: 7 - leucogranite; Mesozoic complexes: 8.

Metamorphic development of the metapelites

The very complicated problem of the Kohút crystalline complex metapelites metamorphic evolution is the distinction of individual metamorphic events. Our interpretation of the metamorphic evolution of discussed metapelites is based on the following aspects:

a) on the documented mineral assemblages in metapelites;

b) on the presence of two garnet generations in metapelites;

c) on already published geochronological data.

Mineral assemblages in metapelites

Our consideration dealing with the existence of the older metamorphic mineral assemblages is based on the chemical composition of metapelites as well as on the most probable PTX conditions of the older metamorphic recrystallization (Méres & Hovorka, 1990, 1991). From the chemical composition studies result that the protolith of mica schists had the character of cley schists (pelites) with variable admixture of finegrained quartz (gneisses). According our views PTX conditions of the older metamorphic recrystallization reached the amphibolite facies conditions. It results from the composition of Gar I, and from the observed pseudomorphs of Bt II after Bt I which are enclosed in Gar I only (Pl. I/c). Pseudomorph of Bt II after Bt I differ from the newly formed Bt II in the matrix by the presence of anomalous amount of small ilmenite crystals. They are concentrated on the biotite cleavage planes. Ilmenite is supposed to the product of Bt I (which was characterized by higher Ti content) breakdown. We did not find any systematic differences in the chemical composition of Bt II pseudomorphs in Gar I and newly formed Bt II in the matrix. From the Gar I and Gar II comparison and supposed higher Ti content in Bt I it follows that discussed minerals of lst generation originated under higher temperature as the minerals of the younger generation.

From the metapelites chemical composition and from indirect deduced PTX conditions of metamorphic recrystallization it follows that in the mineral assemblage of 1st metamorphic event plagioclase, quartz, muscovite, garnet I, staurolite and probably some of Al_2SiO_5 modifications could be present. From the older mineral assemblage mentioned above only Gar I have been so far identified.

In metapelites (mica schists, paragneisses) studied in prevailing amount are present rock forming minerals which are the result of younger regional metamorphic recrystallization. In mica schist this mineral assemblage is as follows: chlorite, muscovite, quartz and garnet II. According to the chemical composition of the rocks as well as depending on the intensity of the metamorphic recrystallization staurolite, chloritoid, kyanite, biotite, plagioclase and graphite are sporadically present. In paragneisses the assemblage is represented by: muscovite, quartz, plagioclase, with local presence of chlorite, biotite, garnet and graphite.

The above presented mineral assemblages prove that the younger metamorphic recrystallization reached low- till middle temperature (350 - 550 °C) and middle- till high pressure conditions (500 - 800 MPa). The results of geothermometry (Gar-Bt and Gar-St geothermometer; Méres & Hovorka, 1991) falls under presented temperature interval. Maximum pressure conditions are expressed by the kyanite presence in the stabile mineral association of metapelites under discussion.

Two garnet generations in metapelites

In mica schist as well as in paragneisses of the Kohút crystalline complex we have distinguished two genetic garnet types (Méres & Hovorka, 1991). The older garnet generation (Gar I) form in some samples cores of garnets porphyroblasts. It has the pyrope-almandine composition with low grossularite (till 5 %) and spessartite (till 10 %) molecules. In all samples studied, in which we have found this garnet type, also younger garnet generation (Gar II) is present. It forms rims around the garnet of 1st generation. Younger garnet has the pyrope-grossularite-almandine composition with variable spessartite molecule. It differs from Gar I namely by the higher grossularite molecule. The change of chemical composition between Gar I and Gar II is abrupt (P1 I/a). Gar II is present in some of rock samples studied in which no Gar I has been found. In this case Gar II is typicaly progressively zoned, in some samples it has snowball structure (P1 I/d). It is typical for the chemical composition changes studied in profiles that in rims of such garnets the highest pyrope molecule was detected - the garnet rims which are in the equilibrium with kyanite crystals are characterized by relativelly highest pyrope molecule (up till 15 % prp). For Gar II of this type the lowermost (till 2 percent) content of spessartite molecule is typical. The contents of CaO and MnO in sections is declining from the cores to the rims of individual garnet crystals (Méres & Hovorka, 1991). Changes in chemical composition of progresivelly zoned garnets from the centre to the rim is expressed by trend in the field of Gar II (Figs. 2 and 3).

Garnet analyses of mica schist and gneisses of the Kohút crystalline complex were presented recently (Korikovsky et al., 1990). It is evident (anal. D-17) that analyzed garnets correspond to two gar-

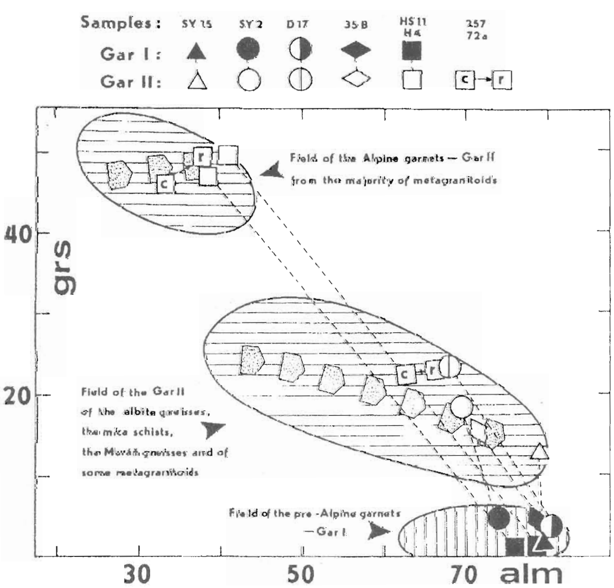


Fig. 2. Typical almandine and grossularite contents in relict pre-Alpine as well as newly-formed Alpine garnets. In fields in which composition of Alpine garnets is presented, trend (dotted arrows) of chemical composition changes in zonal garnets in metapelites as well as in metagranitoids is presented (analytical results are compiled from Vrána, 1980; Korikovskij et al., 1989 and Méres & Hovorka, 1991). Field of pre-Alpine garnets studied (Gar I) is identical with the field of garnets from the pre-Carboniferous metasediments of the amphibolite facies provenience of the central Western Carpathians (Hovorka & Méres, 1991). Field for Gar II of the Kohút crystalline complex metasediments is constructed on the base of individual analyses of progressivelly zoned garnet II (individual analyses are located on the cross section of garnet crystals; Méres & Hovorka, 1991). Mentioned is valid for Fig. 3, too. Garnet analyses are presented in Table 1.

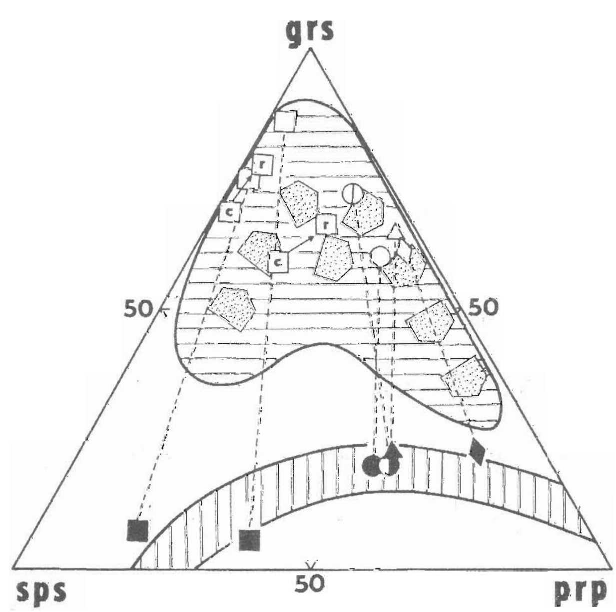
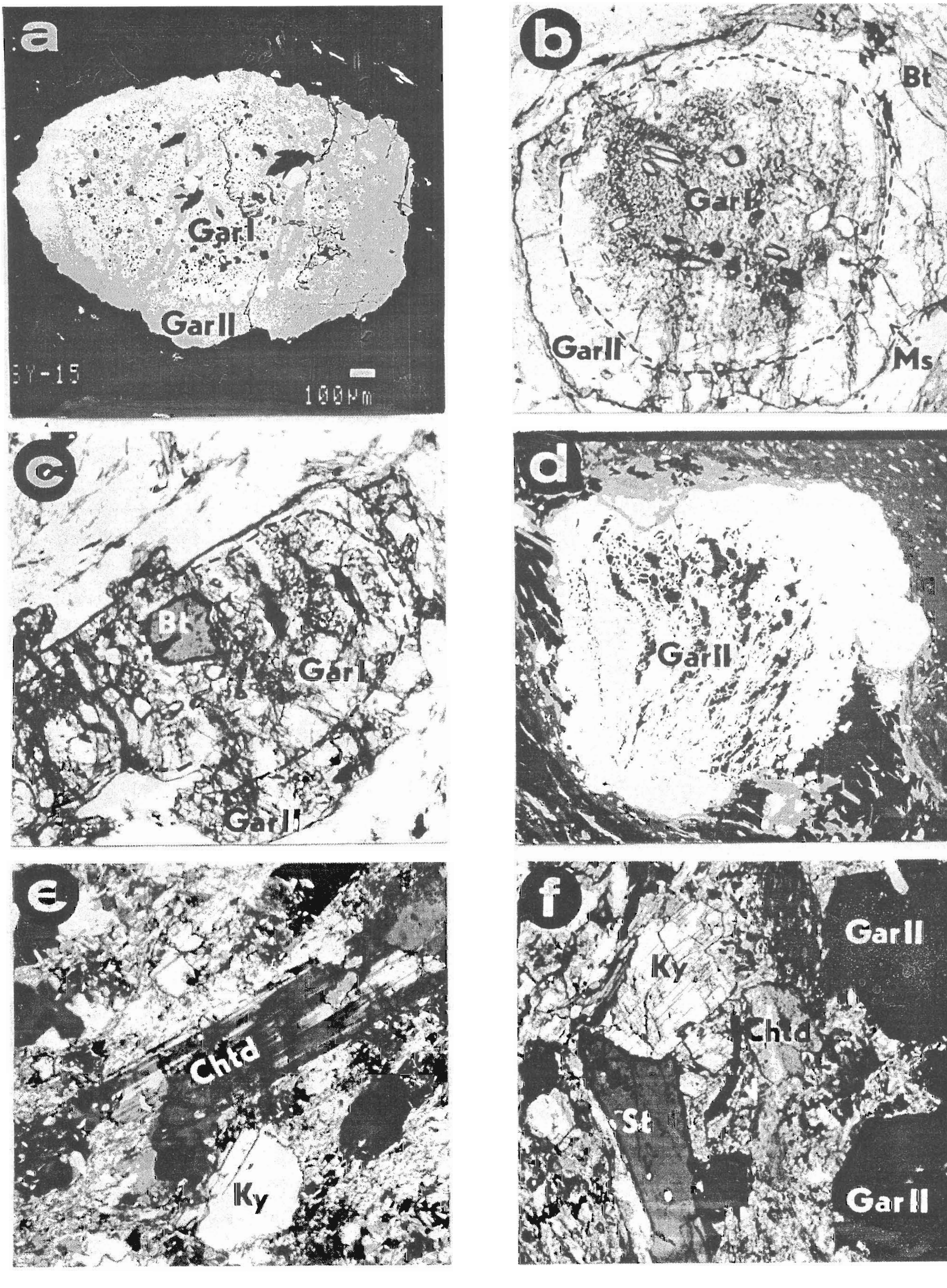


Fig. 3. Spessartite, grossularite and pyrope contents in pre-Alpine relicts of garnets and contents of mentioned garnet molecules of newly-formed Alpine garnets of metapelites, Muráň Gneisses and metagranitoids of the Kohút crystalline complex. Trend (dotted arrows) in the field of Alpine garnet in metapelites shows the change of the chemical composition of zonal garnets (core : rim) in rocks under consideration (Méres & Hovorka, 1991). The highest contents of pyrope and simultaneously the lowermost contents of spessartite are typical for the metamorphic assemblages with kyanite and staurolite (symbols as in Fig. 2).


a
GarI
GarII
SY-15
100μm
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toide rocks) and from the metasediments of the Veporic unit they construct the common biotite isochrone for the whole Veporic unit. This isochrone gives values 94 18 my. Mentioned authors (l. c.) concluded that among the results obtained broad span of results exists.

Discussion

The problem of the intensity and areal extent of the Alpine metamorphic recrystallization in the Western Carpathians crystalline complexes is along discussed problem. Cambel & Korikovsky (1986) concluded that Alpine diapthoresis in the Western Carpathians crystalline complexes is manifested in limited areas by mylonitization. This process is evident namely on the boundary between crystalline and mesozoic complexes; as a consequence blasthesis of sericite and chlorite is observable. Blasthesis of other minerals of metasediments discussed in this paper is considered to be the product of prograde as well as retrograde branches of the Variscan metamorphic recrystallization of regional extent. According to our opinion during the Alpine metamorphic recrystallization of metapelites total recrystallization of biotites, muscovites, plagioclasses and quartz of the older generation is characteristic. Simultaneously blasthesis of newly-formed granet II, chloritoid, staurolite and kyanite has been observed and documented. Recrystallization and blasthesis of mentioned minerals are known from the metagranitoides (Vrána, 1980) as well as from the Muráň gneisses (Hovorka et al., 1987). We consider the mentioned observation the main reason of identical K/Ar isotopic ratios from the granitoide rocks and metapelite micas, which in the most cases varies in the range 70 - 114 my. Mentioned values prove the assumption that in given time span an intensive thermal event took place in the Kohút crystalline complex.

The new blasthesis of other critical rock-forming minerals is proved by the following observation:

i) enclosed or intergrowth muscovites and biotites of the younger generation with the rims of garnets II (P1 I/b);

i) intergrowth and enclosures of staurolite in garnet II;

i) not found staurolite in Gar I;

i) syn-till posttectonic blasthesis of chloritoid and kyanite (P1 I/e, f);

i) chloritoide in thin sections seems to be relatively younger as staurolite;

i) progresivelly zoned Gar II (snowball garnets included).

No garnets of the composition of Gar I (in centre of Gar II) have been found. It suports present authors' opinion of two generations of garnets which originated during two events under different P-T conditons.

Korikovsky et al. (1989a) described the association of chloritoid, staurolite and garnet in the mica schist complex . They supposed the association to be the product of the Variscan metamorphic recrystallization of Al-rich sediments.

Mentioned authors (1989b) have studied garnets of metagranitoids of discussed geological unit. They reach conclusion of uneven composition of cores and rims of garnet crystals. They suppose cores of garnets of pyrope-almandine-spessartite type to be of magmatic provenance. Garnet rims which are characterized by high grossularite molecule content rank among the products of autometasomatic recrystallization of granite massifs. Intensive autometasomatosis according to the mentioned authors (l. c.) could yield in complete grossularization of primary-magmatic (pyralspite) garnets. Some discoveries are against the mentioned genetic interpretation. From the datas documented by the mentioned authors (l. c.) on Fig. 4 (p. 707) sharp boundaries of the mentioned 2 garnet types are evident. Grossularization of garnet I does not follow the cracks and the shape of newly-formed crystals

Pl. I. a - Two garnet generations in garnet mica schist. Sample SY-15; 5 km E of Tisovec; valley of the Losinec brook. b - Two garnet generations in chlorite-garnet-muscovite mica schist. Garnet II intergrowth with Alpine recrystallized muscovite. Sample SY-17; 0,8 km N of Hačava; outcrop in the railway. Magn. 30x, // polars. c - Pseudomorphs of Bt II after Bt I enclosed in Gar I. Garnet-chlorite-muscovite mica schist with chloritoid. Sample SY-9; cut in the brook valley near Čierťaž, 520 m above sea level. Magn. 48x, // polars. d - Snowball garnet (Gar II) with progresivelly developed chemical zonality. Garnet-albite-muscovite-biotite gneiss. Sample SY-11; Hnúšťa, abondoned quarry opposite Smrečina factory; SCAN. e - Alpine chloritoid and kyanite in mica schist. Sample SY-50; Lovinobaňa, 1,6 km SSW from the elevation point Hlbáky. Magn. 95x, X polars. f - Chloritoid-kyanite-staurolite-garnet-muscovite mica schist. Sample SY-58; 5 km SW of Muráň, cut in the brook near Pod Čierťažou. Magn. 30x, X polars.

TAB. 1
Composition of garnets (selected results)

	Metapelites						Muráň gneiss		Metagranitoides							
	2SY		15SY		D17		35B		257		72a		HS11		H4	
	1/2c	1/1r	2/3c	2/1r	5c	8r	c	r	c	r	c	r	26c	24r	6c	1r
	I	II	I	II	I	II	I	II	I	I	I	I	I	II	I	II
SiO_2	37.71	37.97	37.45	37.57	-	-	37.47	37.71	38.60	38.60	38.00	38.10	-	-	-	-
Al_2O_3	21.10	21.46	21.23	21.13	-	-	21.27	21.47	21.50	21.80	21.50	21.40	-	-	-	-
FeO	33.78	31.08	35.51	35.38	35.03	30.17	35.71	31.72	15.30	17.20	27.60	29.30	31.80	16.94	34.47	18.44
MnO	3.39	1.20	2.31	0.29	2.39	1.16	0.94	0.51	8.70	5.40	4.20	1.90	7.50	6.63	5.27	3.02
MgO	3.43	2.56	2.68	1.66	2.79	1.72	3.51	2.57	0.30	0.30	1.40	1.70	1.01	0.20	1.93	0.27
CaO	1.77	6.73	1.60	4.65	1.42	8.28	1.74	6.62	16.90	17.40	7.70	7.60	0.62	16.98	0.30	18.46
total	101.18	101.00	100.78	100.68	-	-	100.64	100.60	101.30	100.70	100.40	100.00	-	-	-	-
alm	74.09	68.34	79.48	79.35	79.20	67.00	79.10	70.00	33.35	37.70	62.34	66.62	75.90	36.90	78.90	40.50
grs	4.97	18.95	4.59	13.36	4.10	23.70	4.90	18.80	46.01	49.05	22.36	22.12	1.80	47.60	1.00	51.80
prp	13.41	10.03	10.69	6.64	11.20	6.70	13.90	10.10	1.20	1.18	5.65	6.89	4.20	0.70	7.80	1.00
sps	7.53	2.67	5.24	0.66	5.50	2.60	2.10	1.10	19.44	12.07	9.65	4.37	18.10	14.80	12.30	6.70

Sample symbols according to original papers: 2SY - albite gneiss (Méres - Hovorka, 1991); 15SY - garnet mica schist (Méres - Hovorka, 1991); D17 - garnet-mica gneiss (Korikovsky et al., 1990); 35B - Muráň gneiss (Hovorka et al., 1987); 257 and 72a - metagranitoides (Vrána, 1980); HS11 and H4 - "granitoides" (Korikovsky et al., 1989); c - core; r - rim; I - Gar I; II - Gar II.

is idioblastic. The thickness of the grossularite garnet rim is more-or-less equal in all present garnet crystals (approximately 100 μm). According to our opinion pyralspite rich garnet which form the core of crystal in accordance with the Korikovsky's et al. (l. c.) view is the primary-magmatic mineral. It is generally known as typical accessory mineral from the peraluminous granite type. Grossularite-type garnet forming the rims around the first type garnets, or forming the individual crystals is supposed to be the newly formed metamorphic mineral (Fig. 2, 3).

Our opinion is based on the identified Alpine recrystallization processes in metapelites as well as in metagranitoids. Simultaneously identical K/Ar ratios in micas and feldspars of both mentioned rock types are determined. For such interpretation also 2 garnet generations identified in metapelites (mica schists and gneisses) and in Muráň gneisses should be used (Tab. 1; Fig. 2, 3).

From the above mentioned results it could be stated that metapelites of the Kohút crystalline complex belong to the polymetamorphic rocks. The older metamorphic event took part under amphibolite facies conditions. In metapelites from that evolutionary stage only relicts of garnet I have been observed. Similar situation is valid for the Muráň gneisses complex.

We suppose mentioned numerical geochronological data (presented in the chapter "Geochronological data") to be the interval of Variscan granite intrusions of the Kohút crystalline complex. Simultaneously we consider the mentioned period to be the period of older metamorphic recrystallization of the metasediments (as well as of the other rock sequences) of the geological unit under consideration. The realization of Alpine recrystallization processes in middle- till high pressure conditions could be documented from the sourrounding geological units, too.

K/Ar ages of the metasediments of the studied geological unit have small dispersion around 94 my. Existing results from metagranite from this unit have greater dispersion. According to the authors' opinion it is the result of the presence of several Variscan granite bodies within discussed unit. Their geological position in the Alpine time-period was different. It is reflected in their uneven Alpine recrystallization. The most intensively recrystallized Variscan granite bodies (which are typical by intensive muscovitization and by the presence of garnet - Vrána, 1980; Korikovsky, 1989) have also small deviation around the value of 94 my. In the less intensively recrystallized bodies of Variscan granites the dispersion of K/Ar values is more pronounced. Higner (than 94 my) ages are, in accordance with the author's opinion, influenced by incomplete (partly) homogenization of older biotites and muscovites as the result of low temperatures of younger (Alpine) metamorphic recrystallization. Lower ages (less than 94 my) are characteristic for the Rochovce granite of the Alpine age and for the rocks of its exocontact aureole, too.

Identification of chloritoid and kyanite was the key argument for Vrána (1964a) to define Alpine chloritoid and kyanite isogrades in the Late Paleozoic envelope of Veporides. Higher metamorphic conditions are proved by the presence of Alpine garnets in metagranitoids (Vrána, 1964a, 1980), too.

In the innermost Western Carpathians zone, i. e. in complexes which in present day erosion level crop out in area eastward of the Kohút unit complexes which have been recrystallized in the medium/high pressure conditions are known. The most typical one is the complex of metasediments and metabasalts, which have been metamorphosed in the glaucophane isograd during the Alpine time period. They are a part of the Meliata group of Triassic till Jurassic age. Glaucophane schists (= glaucophanites) locally pass gradually to the rocks with still preserved original magmatic (ophitic) structure as well as magmatic clinopyroxenes.

Within the last years relicts of medium pressure mineral association of metabasalts with still well preserved pillow textures have been described from the Rakovec area (Hovorka et al., 1989). Since the metamorphosed volcanics in this case are of Early Paleozoic age, the age of their medium pressure recrystallization (the blue crossite being the most typical product of it) which have been followed by typical greenschist recrystallization (actinolite, albite, chlorite) is not proved definitelly. By analogy medium pressure recrystallization is the most probably of the Alpine age.

Concluding the problem we have to add that in the Kohút crystalline complex alpine contact-metamorphic recrystallization took part (Vrána, 1964b; Korikovsky et al., 1986; Fig. 4). This problematics is not dealt with in this presented paper.

Conclusion

Pre- Upper Carboniferous mica schists and gneisses of the Kohút crystalline complex are considered to be a polymetamorphic complex. It is represented by metapelites (mica schists) sometimes with elevated content of quartz together with the other rock types (gneisses). Their source area has been the upper continental crust (Méres & Hovorka, 1990, 1991). Variscan metamorphosis was reached most probably under amphibolite facies conditions. This statement could be deduced from the composition of Ist garnet generation, from still preserved pseudomorphs of Bt II after Bt I, which are enclosed in these garnets.

The second (Alpine) metamorphic recrystallization took part under greenschist facies conditions. It has the character of low temperature/medium pressure metamorphic recrystallization (T = 550 °C - max; P = 500 - 800 MPa). Such conditions are proved by mineral assemblages; together with Plg + Chl + Ms + Qtz gradually together with increasing PTX conditions also other critical mineral phases originated: biotite, garnet II, chloritoid/staurolite and kyanite. Alpine metamorphic recrystallization in mentioned PTX interval along to Margecany - Lubeník lineament affected not only metapelites (mica schists and gneisses) but other lithotypes, too (the Muráň gneisses, graphite metaquartzites, amphibolites, carbonates, granitoide rocks, migmatites). Mineral associations originated in individual lithotypes reflect the chemical composition of their protolith.

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Graptolite zones of Llandoverian in Albania (Korabi zone)

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Llandoverian beds create the lower part of black phyllitic schists with intercalations of silex called the Muhurr black schist formation in the tectonic zone of Korab (NE Albania) and are transsected by gabbrodiabase and minette dykes. The Llandoverian age is determined on the basisof discovered Graptolites (namely in the areas of Buzemadhe and Muhurr, Fig. 1). The Llandoverian schists of the Muhurr Formation contain, in comparison with the higher Wenlockian, Ludlowian and Pridolian levels, a large amount of Graptolites. In the sediments of Llandoverian age, three stages and eight graptolitic zones are (distinguisable) in the stratotype as in various European countries (Tab. 1 and 2).

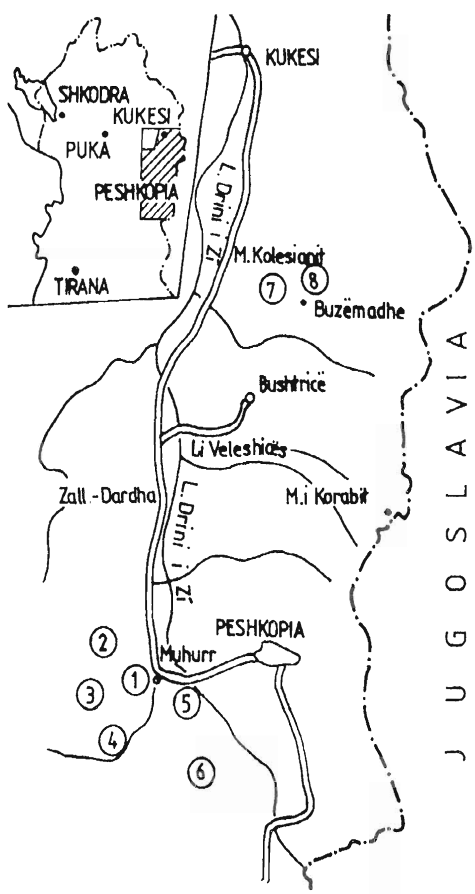


Fig. 1. Scheme of the area of stratigraphic profiles and sites of Graptolites occurrences. 1 - Muhurr, 2 - Bulac and Perroi i Bulacit, 3 - Kodra e Bardhe, 4 - Bufel, 5 - Kodra e Molles, 6 - Arab i Eperm, 7 - Buzemadhe, 8 - Laku i Tejes.

Zonal subdivision

Lower Llandoverian

At the present state of our knowledge the extent of sediments of this substage is limited and the age is determined by Graptolites found at Buzemadhe and Muhurr.

The *Monograptus cyphus* zone

This is the oldest biostratigraphic zone of the Llandoverian beds. The following Graptolites have been found at Buzemadhe: *Climacograptus citocrescens* and *Cl. gr. scalaris.* In Muhurr the Graptolite assemblage of this zone is richer, being represented by large number of individual of *Climacograptus gr. scalaris, Cl.* cf. *rectangularis. Dimorphograptus* sp. and *Monograptus (Pristiograptus) acinaces.*

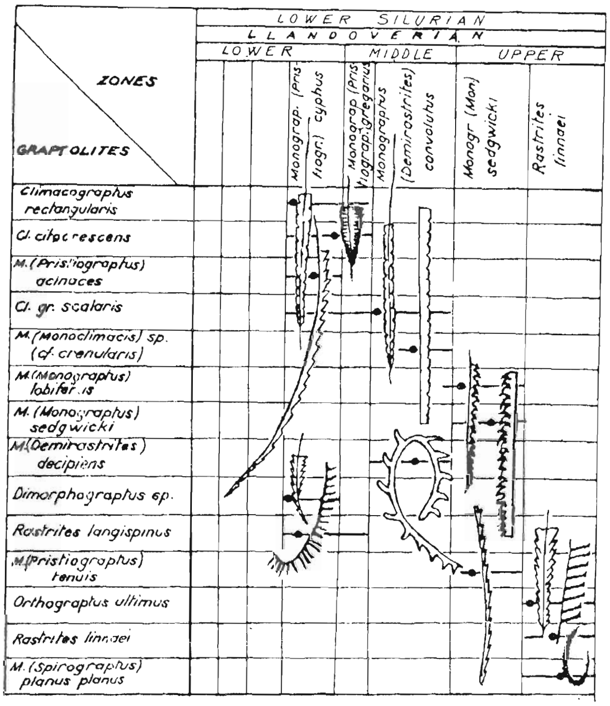
This level with Llandoverian Graptolites represents the zone of *M. cyphus*, but without the zonal index.

Middle Llandoverian

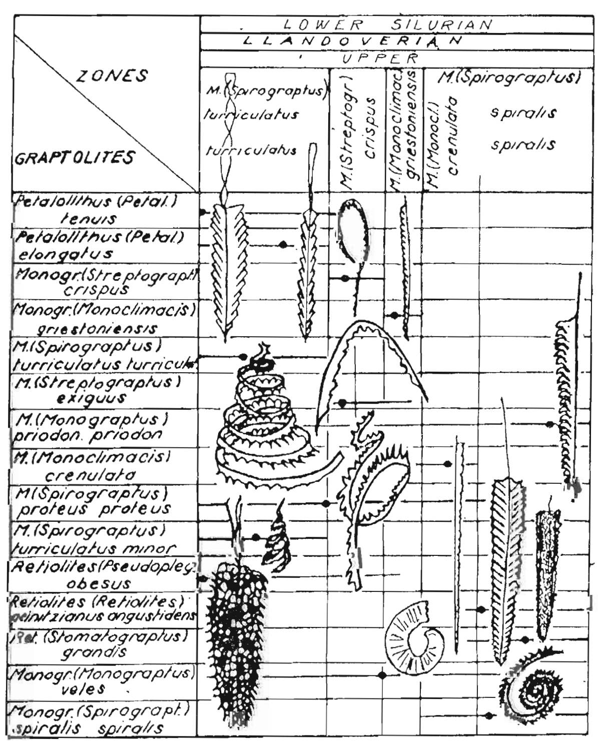
Sediments of this substage have also a limited extent and are found only in the Buzemadhe area.

The *Monograptus (Demirastrites) convolutus* zone

The zone is the upper one of the substage being represented by a thin level of black phyllitic schists in which several badly preserved pieces of (*Mono-*



Tab. 1. Zonal subdivision of the Lower Silurian (Llandoverian) on the base of Graptolites



Tab. 2. Zonal subdivision of the Upper Llandoverian on the base of Graptolites

climacis) p. cf. *crenularis*) occur of a type form of the *M. convolutus* zone (Tab. 1).

Upper Llandoverian

Sediments of the Upper Llandoverian create the thickest and best studied part of Llandoverian profile in the Korabi zone. The sediments occur at Buzemadhe and Muhurr.

The *Monograptus (Monograptus) sedgwicki* zone

It represents the lowermost zone of the substage. Based on the zonal indication and associated forms of *Climacograptus* sp. (cf. *scalaris*) and *M. (M.) lobiferus*, the zone was discovered in black silicic schists with black silex intercalations at Bufel and Koder e Molles as well as in the Bulac profile, relying on *M. (Pristiograptus) tenuis, M. (Demirastrites) decipiens* and *Rastrites longispinus*.

The *Rastrites linnaei* zone

The zone was determined at Buzemadhe on the basis of the associated form of *M. (Spirograptus) spiralis contortus*. At Muhurr (Kodra e Molles), it is determined by its zonal index (typical form in good conservation) associated with *Orthograptus* cf. *ultimus, M. (Sp.) planus planus, M. (Sp.) contortus* and *Diversograptus capillaris*. In the Bufel profile the schists of the zone are characterized by the presence of zonal index and other associated forms as *M. (Sp.)* sp. (cf. *planus*) and *M. (Sp.) spiralis contortus*.

The *Monograptus (Monograptus) turriculatus turriculatus* zone

This is the richest zone in Graptolites namely in Buzemadhe area. There, except for the zonal index which is found frequently and in good preservation, also the following Graptolites are found: *Petalolithus (Pet.) elongatus, P. (P.)* cf. *tenuis, M. (M.) priodon priodon, M. (M.)* aff. *priodon, M. (Prist.) nudus, M. (Prist.) nudus variabilis, M. (Spirograptus) turriculatus minor, M. (Sp.) proteus proteus, Rastrites* cf. *peregrinus, Diversograptus capillaris* etc.

The zonal index of this zone was not found at Muhurr.

The *Monograptus (Streptograptus) crispus* zone

This is a zone with great number of Graptolites as in most European countries and in its stratotype in England, it is accepted as an individual zone. At Buzemadhe the form of *M. (Sp.) with M. (M.) gr. priodon* and *Monograptus* sp. have been discovered. At Muhurr the zonal index has been found in profiles of Bufel and Koder e Bardhe (several individua) associated with *M. gr. priodon* and *M. (Sp.) tullbergi tullbergi.*

The *Monograptus (Monoclimacis) griestoniensis* zone

In Buzemadhe the zone is found in the suite of other Upper Llandoverian zones whereas in Muhurr it was determined in the Bufel profile.

The *Monograptus (Spirograptus) spiralis spiralis* zone

The zone represents the uppermost one of Llandoverian age and it was found only in Muhurr. The southernmost outcrop of the Muhur area, near Arab i Eperm contains the greyish schists of a relatively variable Graptolite assemblage which is guite rich in individua, represented mostly by the *Retiolites* genus. The assemblage contains *M. (M.) priodon priodon, M. (M.) gr. priodon, M. (Monoclimacis) vomerina hemipristis, M. gr. vomerina, M. (Spirogr.)* sp., *Retiolites geinitzianus angustidens, R. (Stomatograptus) grandis.* It is the proof of the summit of Llandoverian beds. Moreover, *R. grandis* in Central Bohemia creates a particular zone (Bouček, 1960), which is the last of the Llandoverian and this zone superpones over the *M. spiralis spiralis* zone. The more northerly zone was discovered in the Bufel and Koder e Bardhe profiles containing *M. (Monocl.) crenulata* and *R. geinitzianus angustidens.* The first species is proof of the base of the zone indicating the lower subzone. Schistoze sediments of the zone are largely developed in the Bulac profile and in the right side of the Perroi Bulac valley. The relatively rich assemblage of Graptolites over the zonal index conatins *Monograptus gr. priodon, M. (Pristiogr.) gr. nudus, M. (Streptograptus)* cf. *nodifer, M. (Strs.)* cf. *runcinatus, M. (Monocl.) vomerina vomerina, M. (M.) crenulata, M.* cf. *anguinus, M. (Spirogr.) proteus proteus, Retiolites geinitzianus angustidens* and *R. (Stromatogr.) grandis.*

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Mesozoic tectonic evolution of the epi-Variscan continental crust of theCentral Western Carpathians - a tentative model

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Abstract

Lateral compositional inhomogeneities of the epi-Variscan lower crust are regarded as primary causes of its heterogenous stretching during Mesozoic time. Variscan orogen in the Central Western Carpathians had generally the same E-W strike of principal structures as the Alpine one, but with an opposite - southern polarity. Variscan internides are framed by parallel belts of granitoid plutons and these belts represent also the most important elevated areas during the Mesozoic extension. A simple rheological model of the epi-Variscan crust is developed, in which zones with anatectic granitoids in the upper crust are underlain by mafic restites in the lower crust, while interstitial zones composed mainly of Variscan and/or pre-Variscan metamorphites are underlain by a quartz-dominated lower crust. Low yield strength and ductile delamination of zones with quartz-rich lower crust led to preferable distension in these zones, hence generating asymmetric rifting and normal faulting in the brittle upper crust. Granitoid-rich zones remained stronger without considerable stretching. Consequently, heterogenous extension was responsible for the formation of individual basement imbrications during the Jurassic - Cretaceous shortening and crustal stacking of the Central Western Carpathians.

Introduction

Pre-Alpine basement of the Central Western Carpathians (CWC) comprises complexes of magmatic and metamorphic rocks consolidated during the Variscan orogeny and low-grade to unmetamorphosed Upper Carboniferous - Permian strata. Little regard on basement composition and its influence on development of Alpine structures of the CWC was taken in earlier tectonic models (e. g. Andrusov, 1968). Plate tectonics brought strong emphasis on the quality of the crust and the whole lithosphere, but for early plate tectonic conceptions only rough division of the earth crust into oceanic, continental and transitional (sometimes called also subcontinental or paraoceanic) was sufficient. Just this rather vague term "transitional crust" was frequently used to account for considerable intracontinental mobility during the Alpine orogeny. However, the nature of transitional crust remained enigmatic.

Later Mahel' (1980, 1981), in order to explain differentiated Mesozoic subsidence in the CWC, introduced the conception of "heavy" or "unstabilized" crust with a great deal of mafic and ultramafic bodies in the pre-Alpine basement, tending to subside, and "light" or "stabilized" crust reinforced by granitic plutons, which tends to build up buoyant elevations. In its consequences this model presumes an isostatic rebound of the pre-Alpine crust triggered by Alpine mobility prior to shortening and fails to explain the facts that Variscan crust was isostatically balanced far before first truly Alpine crustal events, that volumetrically insignificant mafic and ultramafic bodies in the upper crust negligibly contribute to the total crustal density, and that material composition of the upper crust plays a secondary role in lithospheric deformations.

The aim of this contribution is, based on the reconstruction of Mesozoic tectonic evolution of the CWC, to characterize briefly the structural response of the epi-Variscan continental crust of the CWC area to Mesozoic plate and intraplate mobility. For this purpose, preliminary reconstruction of Variscan orogen in the CWC domain is outlined, Mesozoic sedimentary, magmatic, metamorphic and structural rock record is shortly characterized and an unsophisticated, strictly tentative two-dimensional rheological model is presented.

Variscan orogen and the CWC crust

In spite of numerous petrological, geochronological, biostratigraphical and structural observations, unfortunately no general tectonic model of the Variscan orogeny in the Western Carpathian area is

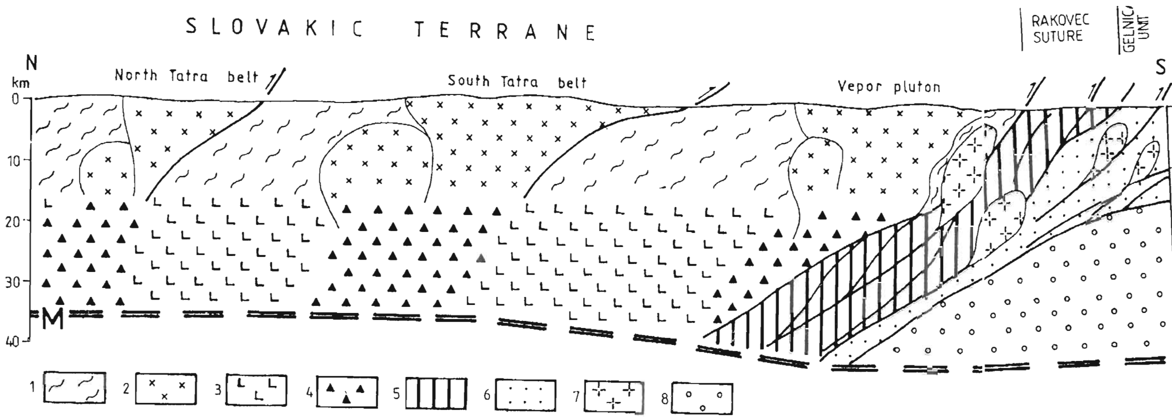

Fig. 1. Epi-Variscan (approximately Upper Permian) structure of the crust of the later CWC. Only principal units and structures are outlined, post-collisional phenomena are omitted. 1 - Variscan and pre-Variscan metamorphic complexes, 2 - early to late Variscan anatectic granitoid plutonites of the upper plate, 3 - quartz-dominated lower crust, 4 - mafic restites bellow granitoid plutons, 5 - Devonian to Lower Carboniferous low- to medium-grade volcanosedimentary rocks and ophiolites of the Rakovec suture, 6 - low-grade sediments and volcanics of the Gelnica accretionary complex, 7 - late Variscan S-type granitoids, 8 - hypothetic pre-Variscan basement of the lower plate of Variscan orogeny.

available. This deplorable shortcoming cannot be compensated by an oversimplified view presented in next paragraphs.

Generally, the Variscan belt of the CWC seems to be a prolongation of the southern branch of European Variscides (e. g. Matte, 1986) linking the Massif Central and Bohemian Massif. Most of the crystalline basement in the Alps and Carpathians belongs to this belt. Pre-Alpine complexes are composed of high- to low-grade metamorphic rocks and abundant granitoid plutons. The belt had a southern polarity (e. g. Frisch & Neubauer, 1989), which is now masked by superimposed north-verging Alpine thrusts.

Compared with the Eastern Alps, the CWC basement seems to be less complicated, the Alpine overprint is weaker and Variscan structures can be usually detected in micro-, as well as in meso and mega-scales. We distinguish three tectonostratigraphic superunits in the CWC crystalline basement: the Slovakic terrane, the Rakovec suture zone and the Gelnica accretionary complex (in N-S direction - Fig. 1).

Slovakic terrane (name derived from "Slovak block" after Máška and Zoubek in Buday et al., 1960 and "Slovakides" - the term for Central Western Carpathians of Mahel', 1983) involves basement of the Alpine Tatric and most of the Veporic unit. The main characteristics are: probable presence of high-grade pre-Variscan basement, Early Paleozoic volcano-sedimentary sequences metamorphosed predominantly under medium-grade conditions and abundant bodies of Variscan granitoids. Composite granitoid plutons were formed by manifold intrusions of anatectic melts derived mainly from lower-crustal sialic rocks in time of Variscan (Devonian - Lower Carboniferous) crustal stacking and thickening. Thermal equilibration after stacking and subcrustal, subduction-related processes brought about crustal heating and partial melting. Crustal parentage for the majority of granitoids has been assumed by several authors (e. g. Hovorka, 1980; Hovorka & Spišiak, 1983; Gubač, 1989), granitoids are partly peraluminous, hybrid and full of enclaves of metamorphic rocks. Two types can be distinguished: smaller, presumably older group of leucogranites with ambiguous I/S typology (Cambel & Petrík, 1982) and somewhat younger group of tonalitic plutons exhibiting mantle influence - mafic enclaves and dioritic pods (Petrík & Broska, 1989) and tending to the Caledonian I-type (Cambel & Vilinovič, 1987). Therefore the lower crust underneath may be dominated by intrusive mafic rocks of ultimately mantle origin (Petrík, pers. communication), besides of mafic restites after melting (Fig. 1). The formation of Variscan granitoids in the CWC covered a considerable time span (380 - 300 Ma).

The Slovakic superunit represents the internal metamorphic belt of the Variscan orogeny comparable with the Moldanubian and Ligerian zones in the extra-Carpathian Variscides including the Helvetic basement and small Pannonic and related terranes of the Central Eastern Alps (cf. Frisch et al., 1990; Frisch and Neubauer, 1989). At the same time it represents also the upper plate of Late Variscan (Carboniferous-Permian) subduction-collision orogeny.

The Rakovec suture zone covers present-day southernmost Veporic and northern Gemeric zones and comprises numerous slices of an active margin and ocean floor derived rock complexes: Devonian-Lower Carboniferous low-grade volcano-sedimentary sequences (Hladomorná dolina complex - Klinec, 1966; Predná hoľa complex - Bajaník et al., 1979; Lovinobaňa and Sinec complexes - Bezák, 1988; all in the Southern Veporicum),Gemeric Rakovec and Klátov groups with dismembered ophiolites (cf. Bajaník, 1975; Dianiška & Grecula, 1979; Grecula & Hovorka, 1987; Ivan, 1989 etc.), early Late Paleozoic Ochtiná and Črmeľ formations representing fillings of subduction-related remnant basins (Vozárová & Vozár, 1988) and amall, but abundant protrusions of ultramafics. Syn- to post-kinematic bodies of Late Variscan granitoids are also present. Rakovec suture forms a narrow belt with intricate structure accreted during the Lower Carboniferous (before Moscovian), which was divided into two principal basement units - the Veporic and Gemeric ones during the paleo-Alpine deformation (cf. Plašienka, 1984). One cannot exclude that Tatric basement of the Malé Karpaty Mts. in SW corner of the Western Carpathians also belongs to this belt.

Rock complexes of the Gelnica Unit, mostly flysch sediments, originated from southern supplies (Ivanička et al., 1989) formed originally an Early Paleozoic volcano-sedimentary cover of passive margin (Grecula, 1982) of the poorly known "southern continent" with a Pan-African basement - the lower plate during the Variscan orogeny. They were detached and accreted during the Late Paleozoic and intruded by late Variscan (300 - 260 Ma) S-type granitoids formed by melting of wet sediments and volcanoclastics during A-subduction and collision. Foreland of the Gelnica accretionary prism is represented by only slightly metamorphosed and deformed Paleozoic complexes mainly of carbonate platform origin of the Rudabánya, Uppony and Bükk Mts. in northern Hungary, creating a foreland fold-and-thrust belt. Rakovec and Gelnica accretionary complexes can be best correlated with the Koriden terrane and cover rocks of the Noric composite terrane of the Eastern Alps (cf. Frisch & Neubauer, 1989).

Short overview presented above enables us to define roughly probable compositional properties of the epi-Variscan continental crust in the CWC area (Fig. 1). Late Variscan transpressional structures and post-collisional pull-aparts and grabens are omitted in this scheme. Thus, the Slovakic terrane was composed of several parallel belts rich in anatectic granitoid bodies with mafic restites in the lower crust and belts with predominance of metamorphic rocks, which were probably underlain by a quartz-dominated lower crust - originally upper crustal rocks deeply burried during the Variscan crustal stacking. Three parallel granitoid belts in Fig. 1 may be an artificial phenomenon arisen from the Alpine strike-slip splitting.

Only few elements of the original epi-Variscan structure of the lower crust are recently preserved in the CWC crust. Not only Early Alpine crustal stacking superposed different upper and lower crustal domains, but also strong renewed Late Alpine transpression, transtension and splitting in time of the Tertiary tectonic escape of the Western Carpathians led to the formation of modern lower crust, in which except pre-Alpine elements large volumes of Meso-Cenozoic magmatic mobilizates participate as well (cf. Fusán et al., 1987.)

Mesozoic terranes and paleotectonic evolution of the CWC

Although the Western Carpathians cannot be analysed in terms of typical "suspect" terranes, in certain time intervals some portions of their recent collage created independent crustal elements often separated by oceanic domains from their later neighbours. Mesozoic tectonostratigraphic terranes of the Western Carpathians are summarized in Fig. 2. Only two types of terranes are distinguished - continental ribbons or "microcontinents" sometimes with strongly attenuated crust and oceanic domains preserved in slivers of dismembered ophiolite fragments and oceanic sediments. Terranes are near-travelled and usually show close evolu-

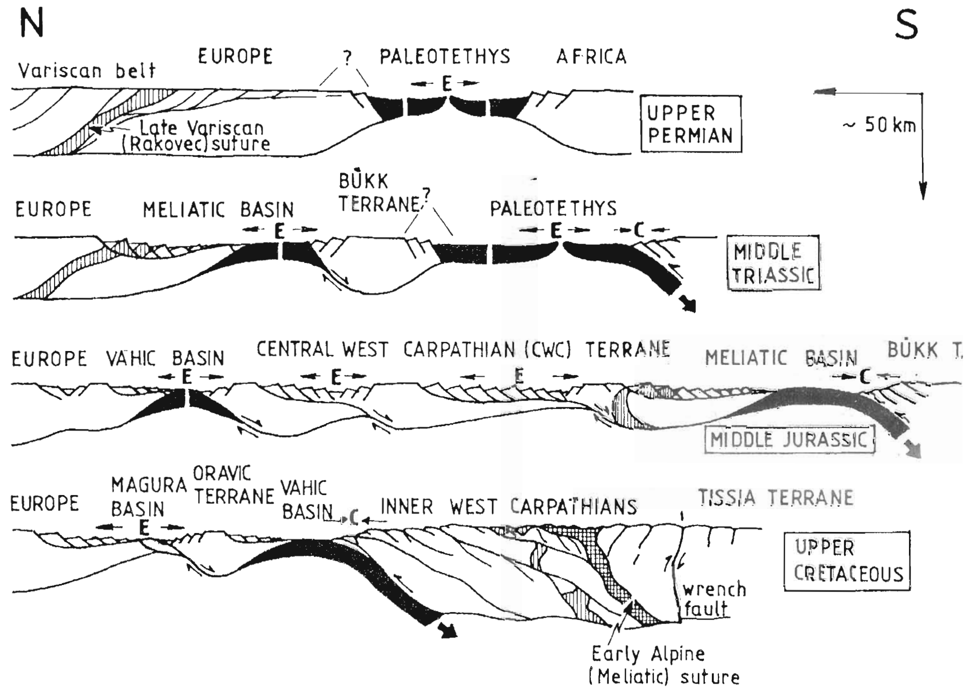

Fig. 2. Nomenclature of Mesozoic tectonostratigraphic terranes of the Western Carpathian area along a mid-Slovakian transect. The position of Paleotethys and Africa is not well constrained, Europe is inferred to be stable. Compared with the Alps, the Magura basin can be roughly analogized with the North Penninic (Valais) trough, the Oravic terrane (comprising most of the Klippen Belt units) with the Brianconnais high, the Vahic basin with the South Penninic (Piemonte-Ligurian) ocean, the CWC terrane with the Austroalpinicum, the Meliatic basin and Bükk terrane show affinities to the South Alpine - Dinaric belt. The Tissia terrane is an exotic microcontinent. Terranes south of the Penninic-Vahic basin formed together so-called "Kreios plate". E indicates extension, C compression, thick arrows driving slab-pull forces. Oceanic crust is shown black, sutures are hachured, continental crust is unornamented.

tionary interconnections and often also gradual transitions, thus they should be rather called "tectonotypes" according to Mahel´s terminology (Mahel', 1988).

Mesozoic processes of crustal deformations recorded by sedimentary, magmatic, metamorphic and structural rock phenomena show clear northward migration in the CWC terrane. In a broader sense this migration, i.e. "orogenic polarity", is also documented by asynchronous duration of two oceanic domains, which bounded the CWC terrane from both subequatorial sides - the southern branch of the Tethys (Triassic-Jurassic oceanic basin of the Meliaticum) and the northern one (Jurassic-Cretaceous oceanic basin of the Penninicum-Vahicum). The latter was in tension at least till the Albian, while compression in southernmost zones of the Western Carpathians began probably as early as in the Lower Jurassic.

Mesozoic paleotectonic development of the CWC can be divided into 4 main periods: 1) platformal epeirogenesis (Early-Late Triassic); 2) extensional tectonic regime, thinning of the crust (Middle Triassic-Middle Createaceous), 3) compressional tectonic regime, thickening of the crust (Middle Jurassic - Late Cretaceous, and 4) transpressional-transtensional tectonic regime, partial disintegration of the crust (Late Cretaceous). In the first Triassic period most of the CWC area obviously had a stabilized, continental crust of the epi-Variscan platform. Its gradual sinking to the sea level had an epeirogenic and/or eustatic character. Lower Triassic piedmont-beach clastics and Middle Triassic carbonate platform sediments covered almost the whole area of the CWC. Attenuation of the epi-Variscan crust and collapse of carbonate platforms to bathyal depths started at the southern edge of the CWC already in the Middle Triassic (Upper Anisian in the Meliatic and adjacent units) and exhibits northern migration - at the Ladinian-

Carnian boundary it reached the Southern Veporicum (black shales and cherty limestones of the Foederata group) and during the Carnian the Northern Veporic and Krížna zones (Lunz or Raibl event). In the Norian times, stable shelf conditions (Hauptdolomite) returned in the Southern Veporicum and continental-lagoonal environment dominated in zones north of the Veporicum (Carpathian Keuper group), while the Meliatic basin remained in deep-marine sedimentary conditions.

At the beginning of Jurassic the CWC suffered a new, much stronger tensional impulse. Fundamental rebuilding of the paleogeographical pattern took place at that time and longitudinal deep-water basins and shallow-water highs were established for the next 100 Ma. Deep-water Zliechov trough was created between the Veporic and Tatric zones, which became the sedimentary area of the later Krížna cover nappe. A similar, but probably narrower basin (Šiprúň or Fatra trough) appeared inside the Tatricum, which may have been in some places joined with the Penninic (Vahic) oceanic furrow rimming the Tatricum from the north. However, at least in the south-western part of the CWC a narrow, fault-bounded subaeric ridge persisted (an equivalent of the Lungau swell of the Eastern Alps - Plašienka, 1987). Extensional activity in the Krížna and Tatric zones culminated by small portions of basanitic hyaloclastic lavas of upper-mantle origin (Hovorka & Spišiak, 1988) piercing strongly thinned crust in the Barremian-Lower Albian.

Onset of compressional movements and basement shortening had a marked northern progradation as well. Its beginning is dated as Early-Middle Jurassic in Meliatic and Bükk terranes (flysch, olistostromes), thus reflecting southward subduction of Meliatic basin substratum (blueschist metamorphism) ending before the Oxfordian (cf. Kozur, 1990). Basement shortening of the Gemeric-Southern Veporic zones occured probably during the Late Jurassic-Early Cretaceous (no sedimentary record). Further to the north, the onset of flysch sedimentation in the Lower Albian of the Krížna (Zliechov) trough reflected a nascent underthrusting zone at the northern edge of the Veporicum, where the thin attenuated crust of the Zliechov basin substratum was gradually consumed. During the Middle Albian, the flysch sedimentation covered most of the CWC area, including previously uplifted zones. This may be attributed to the flexural downbuckling of lithosphere loaded by compressional orogenic wedge advancing from the hinterland, as well as to thermal subsidence caused by shrinking of cooling top parts of astenosphere, upwelled during the preceding period of maximal extension.

From the structural point of view, the underthrusting of zones with thinned crust (A-subduction) is manifested by the paragenesis of the first Alpine deformational stage AD_1, the main structures being subhorizontal ductile shear zones, large recumbent folds and basement nappes. In the Krížna zone, the cover elements were detached from their reduced and underthrust basement substratum along horizons of Werfenian and Keuper shales and evaporites, and accumulated in a pile of duplexes and recumbent folds of the collisional accretionary prism (Fig. 5b). This pile was later (during the Turonian) gravitionally spread and overthrust more to the north lying Tatric zones. In deeper structural levels, more rigid basement elements of the Veporic rear buttress and buoyant South Tatric ribbon continent came into collision (Plašienka, 1983), which is structurally recorded by the deformational stage AD_2 having compressional and partly transpressional character. Linear mocrofolds with subvertical axial plane cleavages are the dominant structures. Uplift and cooling were the accompanying features. Structural recording of stages AD_1 and AD_2 displayed foreland migration; the front of AD_1 is expressed by consuption of zones with attenuated crust and by subhorizontal thrusting, the front of AD_2 by collision of thicker rigid crustal segments and their accretion to the hanging wall of the compressional orogenic wedge.

Advancement of the compressional front through northern zones of the CWC may be estimated on the basis of the termination of flysch sedimentation - it was Cenomanian in the Krížna basin, Cenomanian-Turonian in the Southern Tatricum and Turonian-Coniacian in the Northern Tatricum. During and after Coniacian, shortening fully attacked areas located north of the Tatricum, i.e. the oceanic crust of the Penninicum-Vahicum. Rootless cover

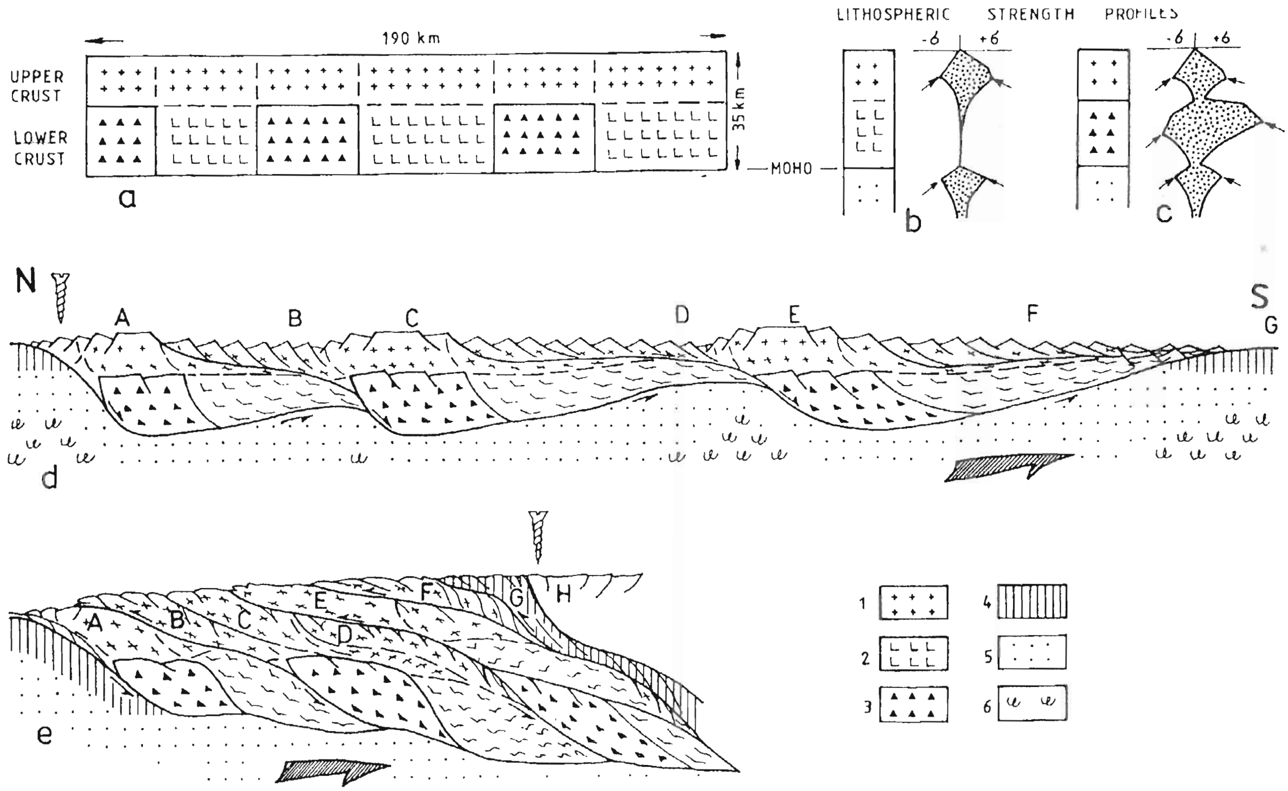

Fig. 3. Paleotectonic evolution of Mesozoic crustal deformations of the CWC. a) simplified (cf. Fig. 1) rheological model of the epi-Variscan crust. Upper crust is mostly rigid-brittle, rheology of the lower crust depends on its composition; b) yield strength envelope for a continental lithosphere with quartz-dominated crust and olivine-dominated mantle and c) strength envelope for a lithosphere with quartz-dominated upper crust, feldspar and/or pyroxene-dominated lower crust and olivine-dominated mantle. Distensive stresses are given by negative sign, compressive positive. Dotted areas represent rigid behaviour, arrows indicate brittle-ductile transitions. Taken and adapted from various authors; d) cross-sectional plane strain model of Mesozoic extension of the CWC. Only basement complexes are shown. Note that demonstrated is the maximal extension reached, which was partly asynchronous; e) model of Mesozoic shortening and crustal stacking of the CWC. Figures a, d and e are areally balanced. A - North Tatric swell, B - intra-Tatric Šiprúň basin, C - South Tatric swell, D - Krížna basin, E - North Veporic swell, F - Gemeric-Meliatic basin, G - Meliatic oceanic domain, H - Bükk terrane. 1 - quartz-dominated upper crust, 2 - quartz-dominated lower crust, 3 - mafic lower crust, 4 - oceanic crust, 5 - mantle lithosphere, 6 - astenosphere. Large arrows indicate the sense of bulk simple shearing of the lithosphere.

nappes (Choč and higher ones) were obviously during the Upper Turonian gravitionally overthrust onto partly peneplainized CWC area. They are structurally alien to the CWC and do not respect their orogenic polarity. Their homeland is supposed to have been located further to the south-west of the CWC.

Tectonic model of crustal deformations

For better understanding the crust-scale deformations in the CWC, a simplified, two-dimensional rheological model (viewed in cross-sections) was developed. Epi-Variscan crust prior to Alpine movements is taken as a consolidated, isostatically balanced double-layer, 35 km thick. Quartz-dominated segments with similar composition as the upper crust are usually wider than quartz-depleted, mafic segments (Fig. 3a). Other material as well as all structural inhomogeneities are not taken into consideration. This pattern was obtained by schematization of Fig. 1.

Rheological behaviour and yield strength of polymineralic rocks building up the Earth crust fully depend on rheology of their weakest mineral constituent, on condition that it creates at least one third of the total rock volume (e.g. Suppe, 1985). Quartz is the weakest wide-spread mineral of the sialic continental crust (at least of its upper part), therefore the upper crust and some segments in the lower crust can be modelled as wet or dry quartz, respectively Rheological profile of a quartz-dominated continental crust is shown in Fig. 3b. Such crust is then two-layered, with rigid-brittle upper third and largely ductile lower two thirds. A considerable weak detachment zone exists in the lower crust and especially at the Moho-level, dividing quartz crust and olivine-dominated mantle lithosphere.

Quartz depletion of the lower crust due to anatexis and upward intrusion of acidic melts led to the different rheological profile (Fig. 3c). Lower crust is composed of mafic, feldspar and/or pyroxene dominated rocks, i.e. restites and cumulates after melting. In this case, two weak zones are developed in the crust - one at the base of quartz-rich upper crust, the other at the Moho again. Nevertheless, the crust as a whole is now considerably stronger compared with the previous model and decoupling on weak zones is less pronounced.

Mode of extension

As we can read from the sedimentary record of the Mesozoic paleotectonic history, the CWC crust sustained an extensional tectonic regime for approximately 100 Ma. Four longitudinal deep-water basins were formed - bounding Vahic and Meliatic of oceanic nature and within-plate Šiprúň and Krížna troughs (Fig. 3d). The overall stretching factor is supposed to range between 1,5 and 2 in crustal segments with strongly thinned, but still continuous, continental crust. Subsiding basinal domains were separated by relatively narrow continental ribbons with prevailing shallow-water sedimentation - the Northern and Southern Tatric and Northern Veporic swells. However, this is a partly artificial pattern - in reality, the foundation, maximal extension and subsequent shortening of attenuated basinal crustal domains were asychronous and show clear northward progradation. Orogenic polarity and foreland migration of tectonic events is in this case regarded to be the manifestation of stress gradients as a result of distensive and later compressive forces exerted upon the southern margin of the CWC plate. The source of the stress field is slab-pull of oceanic lithosphere attached to the CWC edge subducted towards the south (Fig. 2) and / or the drag force acting at the base of the lithosphere generated by uni-lateral convective mantle flow. In both possibilities bulk deformation of the continental lithosphere should have been an uniform-sense simple shear according to the Wernicke´s model (Wernicke, 1985).

Asymmetric model of rifting can be inferred from paleogeographic interpretation of the sedimentary record as well as from later shortening history. If slab-pull is accepted as the main driving force, stress gradients and unequal strain rates in weak detachment zones of individual crustal segments may have been the principal dynamic reasons for asymmetric extension. Stronger coupling of the lower mafic crust and mantle lithosphere (Fig. 3c) is of special importance, because stress concentration behind the mafic segments brought about lithospheric necking due to widespread lower quartz-rich crust delamination and resulting tectonic, isostatic as well as thermal subsidence in axial zones of rifted asymetric basins (Fig. 3d). In the final stage of assymmetric extension these zones may have eventually served as break-away zones and places, where newly formed oceanic crust appeared, as it probably happened in certain time intervals in southern parts of the Meliatic (Middle Triassic) and Vahic (Middle Jurassic ?) basins. In such a case the oceanic crust production is the result of tectonic denudation of mantle by crust unroofing rather than of typical ocean-floor spreading. Consequently, oceanic crust should have been composed mainly of ultramafics, i. e. serpentinites and gabbros; basaltic pillow-lavas being less important. It is highly probable that extension progressed through oblique-slip faulting and that basins were partly of pull-apart type - a feature, which is not included in our two-dimensional model.

Extension in the brittle upper crust was maintained by block tilting over a low-angle crustal detachment fault jointing the trailing edge of the upper plate with the lower shear boundary at Moho level of each mafic crustal segment (Fig. 3d). Two types of high-angle normal faulting occurred in the upper crust: listric faults at the trailing edges of upper plates above low-angle detachment faults and "domino" or "book-shelf" type faults on opposite sides of basins. Domino faulting and block tilting was more conspicuous in zones with complete break-away (Northern Tatricum - Fig. 3d).

Mode of shortening

The Gemericum, Veporicum and Tatricum as the principal basement units of the CWC upper crust were formed during the paleo-Alpine (Cretaceous) orogeny. Upper crustal crystalline thrust sheets rooted deeply in the lower crust are nicely visible on deep reflection seizmic line 2T (Tomek et al., 1989), so demonstrating crustal duplexes. Northward foreland migration of basement shortening is well documented through sedimentary as well as structural rock record.

Compressive stresses and shortening of the CWC basement resulted from closing of the Meliatic basin by southward subduction of its oceanic crust and collision of the Bükk terrane with the southern edge of the CWC terrane (Fig. 2). However, the bulk deformational regime of the CWC lithosphere remained the same as during the extensional stage - uniform sense simple shear, but now producing crustal imbrications as a consequence of closing of an open oceanic space and appearing of a rigid buttress in the rear part of the CWC crust. The driving forces for crustal deformations remained also the same - in the case of slab-pull it should have been the negative buoyancy of the mantle lithosphere still decoupled from bellow the CWC crust and subducted towards the south.

Processes of crustal shortening and stacking widely utilized preexisting geometry and rheology of the CWC basement. South-dipping low-angle detachment normal faults changed their shear sense for thrust faults, whereas north-dipping high-angle domino type faults were locked. Crustal delamination and thinning in basinal domains accelerated by diastathermal heating due to asthenosphere upwelling in break-away zones led to preferable deformation in these zones - the maximal previous extension changed for maximal shortening. Former trailing edges of upper plates were inverted into leading edges of crustal duplexes (e. g. basement of the northern part of the Meliatic basin into the Gemeric thrust sheet and basement of the northern part of the Šipruň basin into frontal Tatric basement nappes - Fig. 3d, e).

Discussion

Several implications of the presented model are able to account for certain controversial features of the CWC Alpine tectonics. Emphasis on the precompressional extension strongly influencing also basement structures is of special importance.

Metamorphic gaps and pinched synclines

In some parts of the CWC crystalline basement areas (southern slopes of the Nízke Tatry Mts. are typical example) there are rather small bodies of low-to very low-grade metasediments directly juxtaposing medium- to high-grade metamorphites and plutonites. Metamorphism of both rock groups is usually considered to have been synchronous - Variscan (for the summary see Spišiak & Pitoňák, 1990). Common opinion presumes Variscan cover nappes overlying eroded or tectonically denuded basement. However, the process of denudation is poorly understood and nappe emplacement in uplifting areas is highly improbable. We attempt to propose now other mechanism - extensional listric faulting.

The down-throw on listric surfaces at the trailing edge of the upper plate during asymmetric crustal extension may reach several kilometers and extensional allochthons can be produced (Wernicke, 1985; Lister et al., 1986). Through this process, near-surface cover and low-grade basement rocks may have juxtaposed or superposed deep crustal high-grade rocks (Fig. 4a). Recent position of low-grade metasediments in tight pinched synclines is a result of compression of two rock groups with contrasting competency - cuspate-lobate folding (Ramsay, 1982) and reverse or oblique-slip faulting of tilted bottom tips of upper-crustal rocks squeezed into crystalline basement (Fig. 4b). These bottom tips may have been eventually complicated by second-order antithetic normal faults or roll-overs (Fig. 4a), which caused dismembering of the cover complexes into numerous tiny slices usually bounded by mylonitic zones or fault cataclasites. In this aspect, extensional unroofing of deep crustal rock complexes is essentially an Alpine feature, confirmed by the common participation of Late Paleozoic and Mesozoic cover rocks in structures

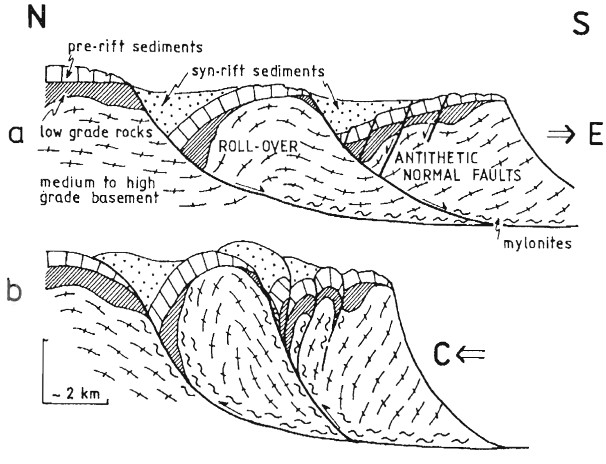

Fig. 4. Origin of pinched synclines filled by low-grade rocks amidst crystalline basement. a) formation of tilting extensional blocks bringing together Variscan low-grade metasediments and unmetamorphosed cover rocks with metamorphites and plutonites of the crystalline basement (Variscan and/or pre-Variscan). Space problem due to listric faulting is resolved by compensating roll-overs or antithetic normal faulting; b) compression (or transpression) and 30 % shortening generates tight synforms and slices of low-grade and cover rocks.

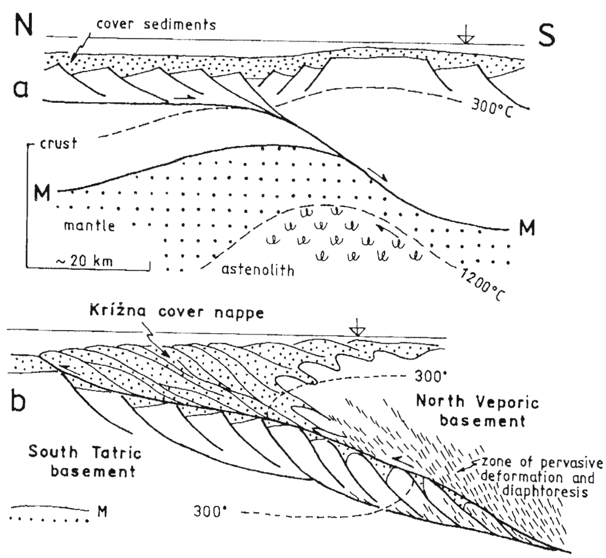

Fig. 5. Development of Alpine diaphtoresis of the Northern Veporic crystalline basement. a) period of maximal crustal thinning and astenosphere upwelling (Lower Cretaceous). Geothermal gradient rose to about 50 oC per km; b) underthrusting period (Cenomanian). Fluid saturation of overheated upperplate crust brought about strain softening, pervasive deformation and diaphtoresis of the crystalline basement.

with Early Paleozoic low-grade metasediments (e. g. the Trangoška syncline).

Sources of Alpine diaphtoresis

Extensive diaphtoresis is known from medium- to high-grade pre-Alpine crystalline basement of the Northern Veporic zone since pioneer works of Zoubek in the thirties, summarized by Máška & Zoubek (in Buday et al., 1960), Kamenický (1977) and Krist (1980). Later findings of low-grade Early Paleozoic rocks cast doubt upon the occurrence of widespread diaphtoresis (Klinec, 1966), but recent papers of Putiš (1987, 1989) confirmed the presence of both - originally medium to high-grade, widely diaphtorized crystalline basement , as well as low to medium-grade volcanosedimentary rocks (the Janov grúň complex - Miko, 1981). Diaphtoresis acted mainly under low-grade conditions similarly as progressive metamorphism of the Permo-Mesozoic cover and is clearly linked to Alpine structures. The problem is the source of heat and metamorphic fluids necessary for diaphtoresis, this may be solved by superposed effects of extension and overthrusting again.

In times of Mesozoic extension, the basement of the Northern Veporic belt was lying close above the zone of maximal crustal delamination and thinning, over a zone of incipient mantle denudation and astenosphere upwelling (Fig. 3d, 5a), which led to diastathermal heating and isoterms rising in the crust. Subsequent period of shortening established an A-subduction zone, in which the Krížna basin basement was underthrust below the Northern Veporic edge (Fig. 5b). Thus the hot, dry upper plate overrode the cold, wet basement and tegument sediments of the lower plate. Thermal balancing heated the lower plate, whose pore fluids were expulsed upward into the footwall of the upper plate, where syntectonic retrometamorphism and volatile saturation in the crystalline basement gave rise to diaphtoritic schists, phyllonitic and mylonitic rocks (Fig. 5b). Since the Moho temperature may have reached as much as 700 - 800 °C, generation of early Alpine crustal magmatic mobilizates cannot be excluded in this zone.

Alpine basement nappe types

Two different basement nappe types are distinguished in the CWC - the Northern Tatric and Northern Veporic types. The former is represented e. g. by the Bratislava nappe of the Malé Karpaty Mts., characteristic features of which are: relatively thin-skinned nature (not more than 5 km), without diaphtoresis in the basement, internal Variscan structure is widely preserved, overthrust plane partly frictional (cataclasites), recumbent folds and imbrications are not typical. The latter features: thick-skinned wedge-shaped principal thrust sheet, extensive diaphtoresis, pre-Alpine structures mostly destroyed, highly strained ductile overthrust plane, recumbent folds and basement imbrications are decisive phenomena.

All these features can be explained by means of structural position before shortening (Fig. 3d). Northern Tatric basement nappes originated from the thin-skinned trailing tip of the extensional upper plate above a thicker, relatively cold crustal segment. Pre-existent master low-angle detachment fault as well as secondary listric normal faults inverted into overthrust planes, hence limiting the geometry of the basement nappes. On the contrary, the Northern Veporic thrust sheet was detached at the brittle-ductile transition in mid-crustal levels, the basement was generally ductile due to diastathermal heating and later fluid saturation (Fig. 5) and was not able to serve as a rigid buttress to support shear stresses necessary for the detachment of the Krížna basin basement, which was almost completely underthrust beneath the Northern Veporic basement wedge. Overthrust plane was a deep-seated, ductile shear zone with slow strain rates causing strain dispersion and penetrative deformation throughout the whole overriding thrust body (Fig. 5b).

Conclusions

Tentative rheological modelling of Mesozoic crustal deformations taking into account paleotectonic evolution of the CWC leads to several important implications:

- heterogenous stretching and thinning of the CWC crust can be better explained by lateral material inhomogeneities in the lower crust, rather than in the uniformly brittle upper crust;

- extensional basement deformations, namely crust-scale low-angle detachment normal faults are widely utilized during subsequent period of shortening and crustal stacking on condition that stretching and shortening had identical, geometrically constant dynamical sources;

- foreland migration of extension and subsequent shortening indicates uniform-sense simple shear of the lithosphere and stress gradients generated by slab-pull driving forces;

- generally speaking, the composition and rheological behaviour of the pre-Alpine basement and the whole epi-Variscan continental crust crucially determines the mode of Alpine crustal deformations.

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Geology and petrotectonics of some shear zones in the West Carpathian crystalline complexes

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Abstract

Ductile - brittle and ductile shear zones have been distinguished, in the Central West Carpathian crystalline complexes. They were formed under anchimetamorphic and/or low-metamorphic conditions.

The age of the shear zones decreases towards the north. The following polarities and ages of the Central West Carpathian shear zones have been recognized: 1. Late Hercynian, 2. Late Kimmerian, 3. Middle Cretaceous and 4. post-Upper Cretaceous shear zones as well as their relationships to subduction or transpressive tectonics.

Models of preferentially oriented quartz c-axes resulted from slip and rotation along quartz basal planes. They define more precisely the sense of the shear as well as relationship of S-C structures.

Introduction

This article summarizes structural-geological and petrotectonic data on several shear zones in the Central West Carpathian crystalline complexes which provide the reader with the overall view of the character and conditions of the shear zone formation, from the Tatric outer margin (northern Subtatricum) of the Central West Carpathians (Malé Karpaty, Považský Inovec Mts.) through the southern Tatricum and/or southern Subtatricum (Tríbeč, central Nízke Tatry Mts.), northern Veporicum (Pohorelá line in the Slovenské rudohorie Mts.) as far as the Veporic/Gemeric contact (Lubeník - Margecany line in the Slovenské rudohorie Mts.).

Discussed topics include the age of the shear zones as well as their relationship to deformation stages in the crystalline complexes from the Late Hercynian period to Paleogene.

The author took advantage largely of his own experience gained during geological mapping of the above-mentioned regions as well as results of structural and petrostructural analyses. The author´s views are confronted with the present-day conceptions of the West Carpathian structure (Mahel', 1986).

The terms "northern" and "southern Subtatricum" are only preliminary names to designate Tatric tectonic subunits which cannot be defined more accurately for the time being. These subunits situated at the Tatric margin, however, undoubtedly display a more complex history than the central Tatric. The "near-klippen zone" (Mahel', 1980b) in broader terms may also comprise the northern Subtatricum.

The continuation of the crystalline tectonic units (Fig. 1) beneath the overlying ones can be compared with the map of the West Carpathian pre-Tertiary basement (Fusán et al., 1987).

Malé Karpaty Mts. - Borinka and Modra shear zones

In the Paleoalpine structure of the Malé Karpaty, the crystalline complexes along with their Permian - Mesozoic envelope are incorporated into three partial nappes (Plašienka & Putiš, 1987):

1. The northernmost paleotectonic unit and at the same time also the lowermost tectonic unit is the subautochthonous Mesozoic of the Borinka unit with the Tatric subautochthonous crystalline complex. The crystalline complexes are largely hidden at depth but their character is evident from clasts of olistoliths in the Mesozoic Borinka sequence. The crystalline clastogene material includes: metapelites (mostly phyllites), metamorphosed mafic rocks (largely green schists), and lesser granitoid rocks.

2. The southernmost paleotectonic unit and at the

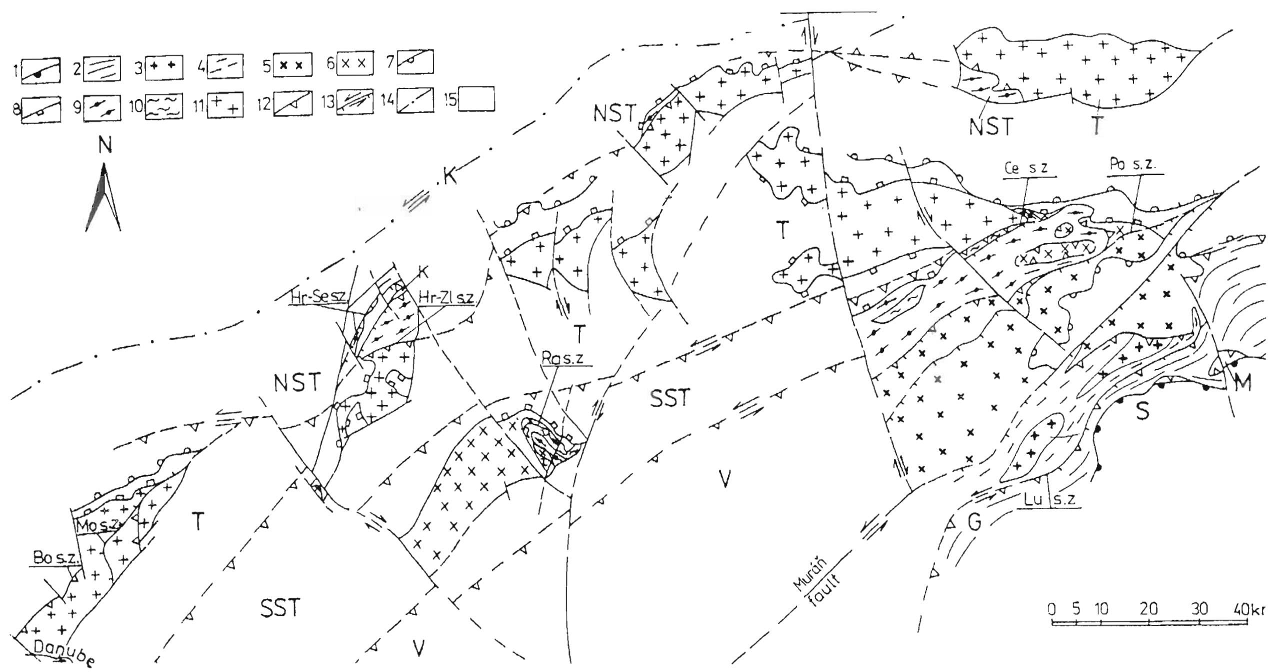


Fig. 1. Tectonic sketch of major shear zones in Tatric and Veporic complexes of the Central West Carpathians. Abbreviations: K - Klippen Belt, NST - northern Subtatricum (Infra-Tatricum), T - Tatricum, SST - southern Subtatricum (Supra-Tatricum), V - Veporicum, G - Gemericum, M - Meliaticum, S - Silicicum, Bo s. z. - Borinka shear zone, Mo s. z. - Modra shear zone, Hr-Zl s. z. Hrádok-Zlatníky shear zone. Hr-Se s. z. - Hrádok-Selec shear zone, Ra s. z. - Razdiel shear zone, Ce s. z. - Čertovica shear zone, Po s. z. - Pohorelá shear zone, Lu s. z. - Lubeník-Margecany shear zone; 1 - Silica nappe thrust plane, 2 - Gemericum, 3 - Permian (?) granitoids, 4 - southern Veporicum crystalline, 5 - granitoids and crystalline schists of central and northern Veporicum (Kráľova hoľa nappe system), 6 - nappe Veporicum thrust onto southern Subtatricum, 7 - thrust plane of Choč and higher nappes, 8 - Krížna nappe thrust plane, 9, 10 - crystalline complexes of southern and northern Subtatricum: 9 - mostly mica-schist-amphibolite complexes, 10 - gneiss and migmatite complexes, 11 - Tatric crystalline (granitoids, gneisses, amphibolites, migmatites), 12 - thrust planes of crystalline nappes, 13 - tectonic boundaries, faults, often with lateral displacements, 14 - Klippen Belt axis at the Outer/Central West Carpathian contact, 15 - undivided West Carpathian geological units.

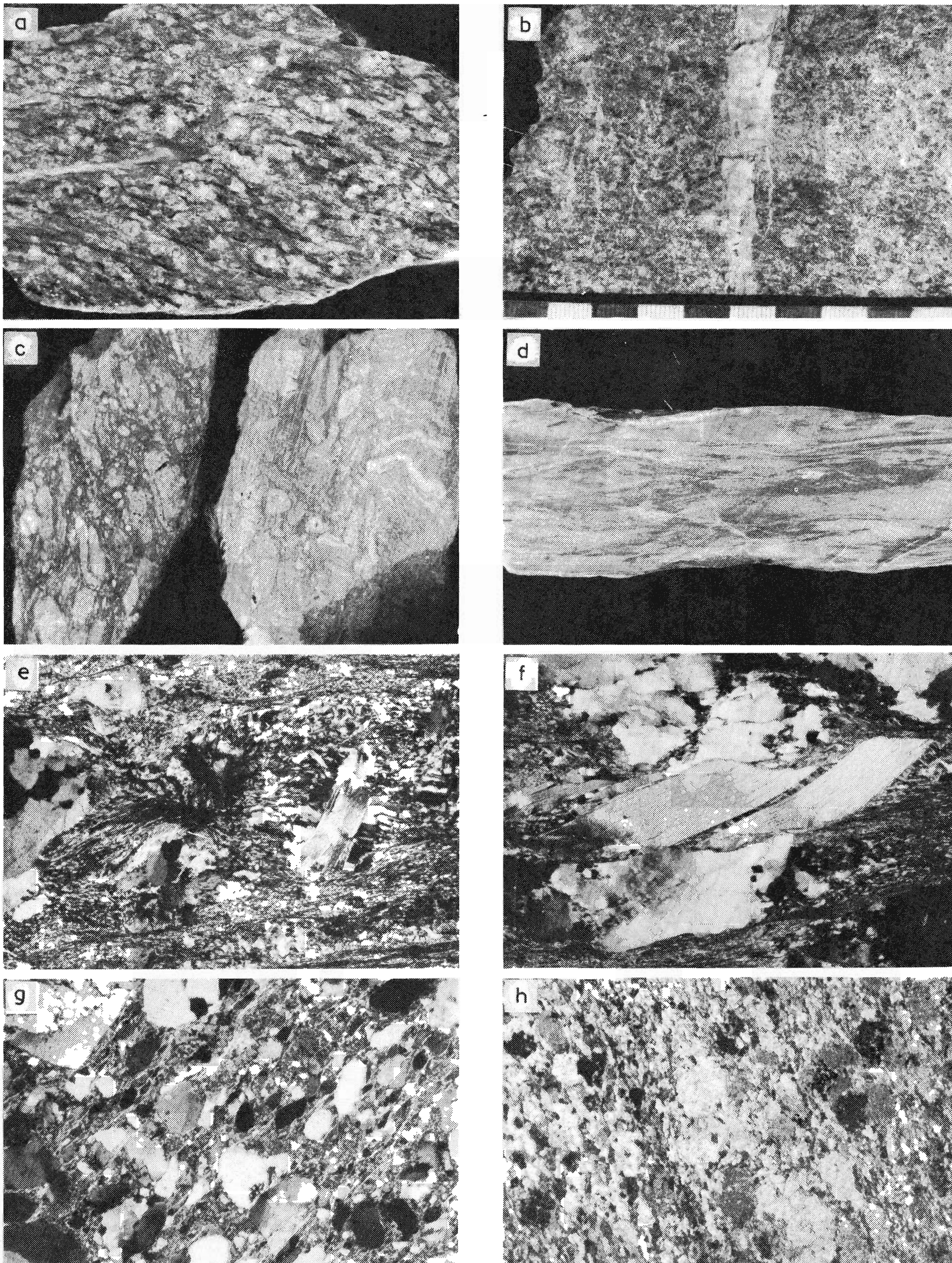
same time the uppermost tectonic unit is the allochthonous Bratislava unit (the Bratislava nappe, sensu Maheľ, 1980a) thrust in the form of a nappe over the Borinka unit towards the NW. The Bratislava nappe consists of the Tatric crystalline complexes (granitoids and metamorphosed rocks) with the Tatric Mesozoic envelope sequences: the Devín, Kuchyňa, Kadlubka, Orešany ones (Maheľ in Maheľ et al., 1967). The Borinka unit (southern Penninicum, sensu Plašienka, 1987) positioned near the southwestern margin of the Malé Karpaty thus represents a semiinlier in the Tatric Bratislava nappe.

3. In another semiinlier of the Bratislava nappe there is the Orešany unit situated near the northeastern margin of the Malé Karpaty. The unit is composed of crystalline complexes (Doľany sequence metamorphic rocks) and Mesozoic rocks, some facies of which resemble calcareous shales of the Borinka unit. In the paleotectonic scheme (prior to the mid-Cretaceous folding) as well as in tectonic section, the Orešany unit lies between the Borinka and Bratislava units.

The thrust plane of the Bratislava nappe is marked by two conspicuous shear zones: Borinka and Modra which are of the same age (77 and 74 m. y., respectively, K-Ar dating, Kantor, 1987.

In the lower part of the overthrust crystalline nappe in the Borinka area (Fig. 1) there is an assemblage of rocks and structural elements characteristic of ductile-brittle shear zones. The zone is mostly tens of metres thick and is made up of my-

Fig. 2. Borinka shear zone in the Malé Karpaty. a - granite S-C mylonite (roughly real size of the sample; the same applies to the other figures without a scale), b - quartz-filled tension fissure perpendicular at stretching lineation in granite mylonite, c - tectonic breccias of metapelites, d - isoclinal horizontal microfolds in Borinka unit Lias marbleized limestones, e - rotated muscovite clast (between C planes) with pressure-shadow quartz in granite mylonite, X nicols, magn. 20 x, f - antithetic slides in muscovite clast in granite mylonite by right-lateral shear, X nicols, magn. 20 x, g - slightly recrystallized Lower Triassic quartzite of Borinka (?) unit, X nicols, magn. 20 x, h - distinct recrystallization of Lias Borinka Limestone, X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lonitized granites and/or also metapelites, indistinct S-C-mylonites (Fig. 2a) as well as ultramylonites and tectonic breccias (Fig. 2c) directly at the shear zone (cf. Putiš, 1986a).

In accordance with quartz microstructure (Figs. 6a-e), asymmetric structures of rare S-C mylonites (Lister & Snoke, 1984) indicate that the nappe was thrust towards the northwest. SE - NW stretching lineation is distinct only in carbonates of the subjacent Borinka unit and partly in granite mylonites where it is accompanied by a transversal (NE - SW) system of en-echelon fissures filled with quartz (Fig. 2b) or calcite. Mylonites commonly contains α-type quartz (Lister & Price, 1978) which is flattened and recrystallized mostly only along the margin and/or along deformation lamellae and bands as well as microshears into an aggregate of small γ-type grains. Feldspars are slightly deformed as well as highly sericitized and albitized. Rare biotite underwent chloritization. Muscovite is frequently rotated between C-planes, antithetically dissected and extended, with fissures being filled with quartz (Fig. 2f) or pressure-shadow quartz (Fig. 2e). Kink-type deformations are common.

In the southwestern Bratislava nappe, Hercynian structures dipping to the NW at 50 - 70° are cut by an Alpine thrust plane dipping SE at 30°. Further north, in the centre of the Borinka shear zone, Hercynian planar structures rotated to the NW into an Alpine position subparallel to the shear zone, and mylonitic foliation was superimposed onto them.

The Modra shear zone (Putiš, 1987a, Plašienka ,1990) in the central and northern Malé Karpaty crystalline complexes is regarded a higher structural level lying hundreds of metres away from the main thrust plane of the Bratislava nappe which was thrust over the Orešany and Borinka units (Fig. 1). The Modra zone has a fold-shear Paleoalpine tectonic style.

Later deformation stages (Laramidian and Savian) are characterized by steep (as much as 60°) fold-shear tectonic style bearing signs of subsequent dextral lateral displacement along the reactivated Modra shear zone (confluence of thrust planes of the Krížna and higher nappes). The system is antithetic relative to the main system - sinistral, in association with the formation of the Carpathian arc.

Považský Inovec Mts. - Hrádok-Selec and Hrádok-Zlatníky shear zones

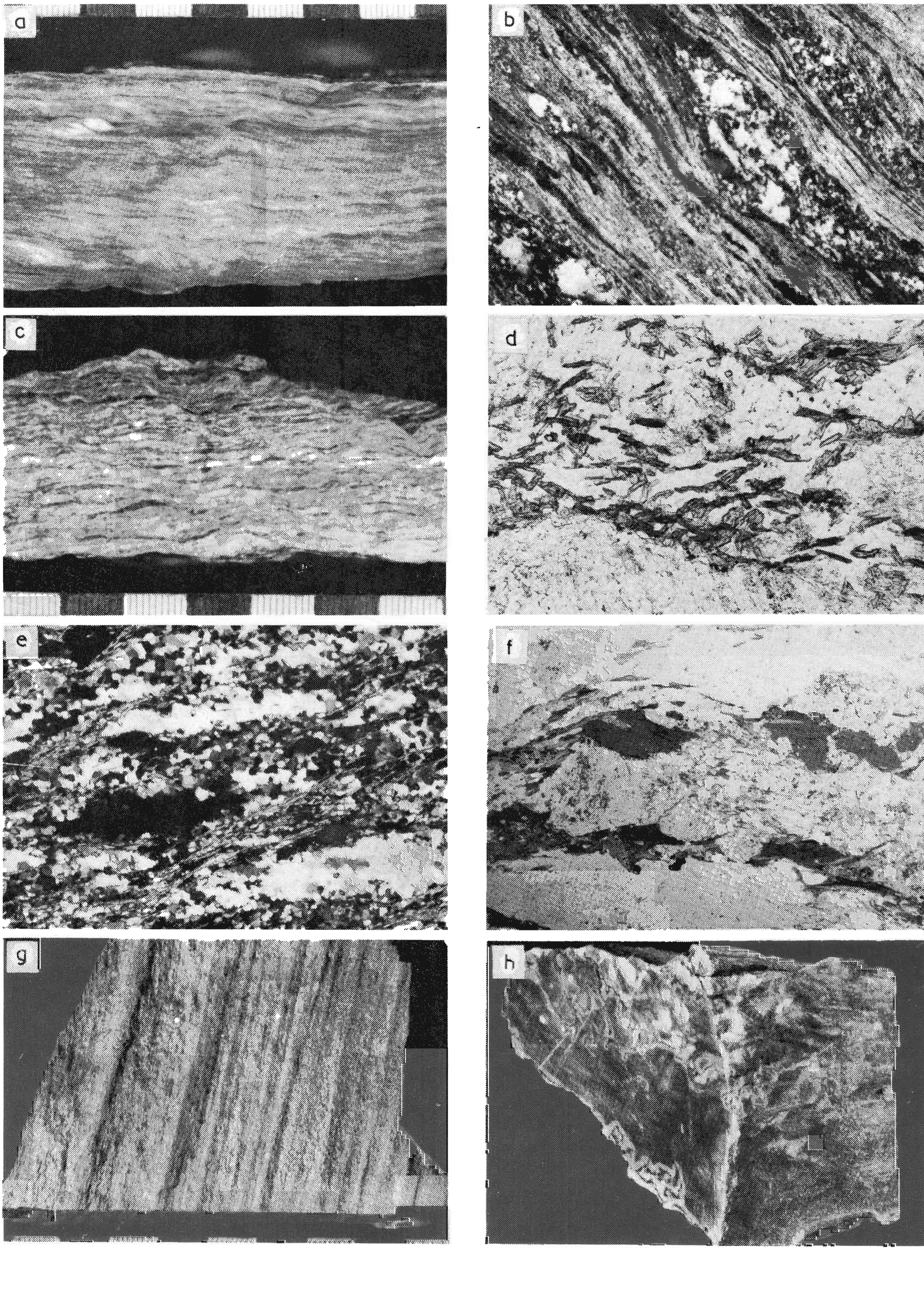
The Považský Inovec crystalline complex is also part of the outer Tatric of the Central West Carpathians and like the Malé Karpaty and western Vysoké Tatry crystalline complexes, it is composed of lower tectonic slices of the Tatricum.

The typical Tatric crystalline complex comprises Late Paleozoic complexes made up of paragneisses, amphibolites, migmatites and granitoids of S-I, S affinity (Cambel et al., 1985) Upper Devonian - Lower Carboniferous in age. In the Považský Inovec, only the southern tectonic block conforms to the above-described character. It was overthrust onto the different, northern block along a phyllonite shear zone tens of metres thick (Figs. 1, 3a, b).

The northern block crystalline is of a fairly unique character, Within the Tatricum such crystalline complexes are known to exist only in the nearby Tríbeč (Krist, 1971) and more distant western Vysoké Tatry Mts. (Gorek, 1956). The Považský Inovec northern crystalline block is composed mainly of mica schists and mica-schist gneisses with staurolite, cyanite, andalusite and sillimanite (Putiš, 1981, 1983; Korikovsky & Putiš, 1986). The Alpine dynamometamorphism in the shear zones took place under anchimetamorphic conditions.

The Považský Inovec crystalline blocks differ from one another in their tectonic styles, particularly as regards the Alpine tectonics. The Early Paleo-

Fig. 3. Hrádok-Zlatníky shear zone in the Považský Inovec (a, b). a - granite mylonite with relics of feldspar porphyroclasts (left shear on C planes), b - highly recrystallized quartz porphyroclasts, feldspars are replaced by sericite-muscovite on C planes in mica-schist mylonite. Razdiel shear zone in the Tríbeč (c, d). c - crenulation cleavage in granodiorite mylonite, d - mylonitization planes in granodiorite with authigenic fine-grained biotite$_2$ and chlorite, X nicols, magn. 35 x. Pohorelá shear zone in the Slovenské rudohorie and eastern Nízke Tatry (e, f). e - highly recrystallized Lower Triassic quartzite of Struženík unit (horizontal - S-planes, oblique - C-planes, left shear),X nicols, magn. 35 x, f - metatonalite with magmatic biotite$_1$ (coarse-grained) and mylonitic biotite$_2$ (fine-grained), // nicols, magn. 20 x. Lubeník-Margecany shear zone in the Slovenské rudohorie (g, h). g - Lower Triassic columnar metaquartzite of Struženík unit with distinct stretching lineation, h - metamorphosed Middle Triassic limestones of Struženík unit with stretching lineation marked by constriction sheath folds.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zoic crystalline complexes (Čorná & Kamenický, 1976) of the southern block were only subjected to slight Alpine tectonic reworking and are therefore dominated by the pre-Alpine tectonic style of open macrofolds with metamorphic foliation and B-axis running WNW - ESE. The most important manifestations of the Alpine tectonics include phyllonitization of the basal part of this overthrust block as well as mylonitization along N-S to NNW - SSE faults. In contrast, the northern, tectonically lower crystalline block is dominated by the Alpine tectonics resulting in reverse-fault to slice-klippen tectonic styles trending generally NNE - SSW with incorporated paraautochthonous Late Paleozoic - Mesozoic envelope including presumably Klippen-Belt Klape-type allochthonous Middle and Upper Cretaceous sediments (Kullmanová & Gašparíková, 1982; Leško et al., 1988).

Dynamometamorphic anchimetamorphic recrystallization is related mainly to two deformational stages (Putiš, 1986b): 1. Paleoalpine, with tectonic style of west-vergent overthrusts and horizontal folds, 2. Mesoalpine, post-Upper Cretaceous, with fold-thrust, reverse-fault and slice-klippen tectonic styles. In both cases, brittle deformation and recrystallization prevail over neoblastosis (quartz, sericite, chlorite).

Asymmetric mesostructures (Fig. 3a) as well as quartz microstructure models (Fig. 6h, i) indicate sinistral shear at the contact between the Považský Inovec northern and southern blocks with E-W to NNE - SSW stretching lineation. Other shear planes in the northern block also converge into this direction (Fig. 1).

Tectonic convergence of the northern and southern block, accompanied by partial overthrusting towards the north-northwest lasted from the Paleoalpine (93 - 53 m. y., K-Ar dating of sericite-muscovite and K-feldspars, Arakeliants, written information) to the post-Upper Cretaceous period (E-W stretching lineation). It came to an end in association with collision-transpressive tectonics in the Klippen Belt area, as suggested by the structural style as well as by the SW stretching lineation and rotation.

The Tríbeč Mts. - the Razdiel shear zone

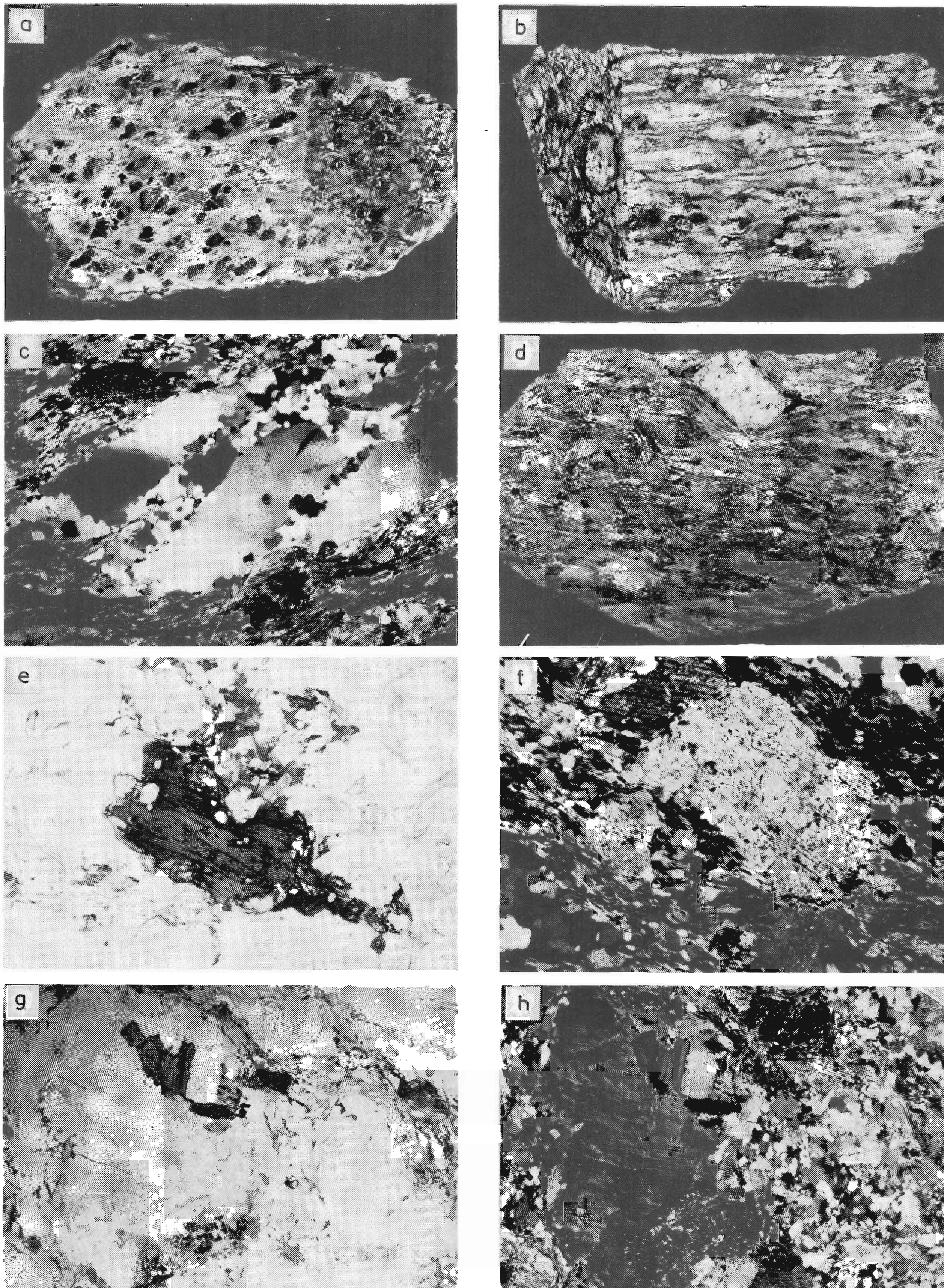
The northern, Razdiel tract of the Tríbeč (Fig. 1) is a NW-vergent fold-shear megastructure whose B-axis strikes NE-SW and plunges to the NE. NW-SE to WNW - ESE stretching lineation as well as quartz microstructure (Fig. 6j, k) suggest that the upper limb was transported to the NW - WNW.

The Alpine megastructural core is made up of mica schists and amphibolites of the Razdiel complex, tectonically overlain by gneisses, migmatites and leucocratic granites of the Kamenné vráta complex which, in turn, are overlain by the uppermost slice (up to 200 m thick) of highly mylonitized granodiorite in the chlorite-biotite zone (Putiš, Krist et al.,1992 in press). The succession in the lower limb exposed in an inlier is reverse. The envelope Permian underwent only anchimetamorphism (Korikovsky, Ďurovič, Boronikhin, in press).

Although the envelope Permian was incorporated into external tracts of this megastructure, the internal nappe structure of the complexes may be original, pre-Permian. This is also suggested by higher-temperature mylonitization of granitoids (Fig. 3c) with authigenic chlorite and biotite (Fig. 3d) as well as phyllonitization of mica schists with authigenic albite in contrast to anchimetamorphosed envelope Permian sediments.

The whole geological structure of this region as well as metamorphic conditions of mylonitization are very similar to those in the northern Veporicum and/or southern Tatricum (southern Subtatricum), i. e. in the root zone of the Krížna nappe.

Fig. 4. Pohorelá shear zone in the Slovenské rudohorie - Fabova hoľa massif (a - d, f) and eastern Nízke Tatry - Kráľova hoľa massif (e, g, h). a - flattened quartz grains at S-planes normal to horizontal C-shear planes in augen mylonite of porphyrite granite, b - elongated feldspar and quartz grains at plane XZ in porphyric granite mylonite, c - microshears in quartz with associated recrystallization in granite eyed mylonite, d - rotated K-feldspar porphyroblast in porphyric metagranite (left shear), e - coarse-grained biotite (magmatic) with sagenite enclosed between mylonitization planes with authigenic biotite$_2$ in metagranite, // nicols, magn. 35 x, f - rotated plagioclase porphyroclast in metatonalite, X nicols, magn. 20 x, g - chloritized biotite$_1$ enclosed in microcline of eyed metagranite with biotite$_2$ and sericite-muscovite$_2$ at mylonitization S-planes, // nicols, magn. 20 x, h - same but X nicols, magn. 20 x.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The Nízke Tatry Mts., centre - the Čertovica shear zone

The Čertovica shear zone has been defined as a plane along which the Verporicum was overthrust towards the northwest onto the Tatricum and at the same time as a scar left after the expulsion of the Mesozoic Krížna nappe to the NW (Andrusov, 1936; Zoubek, 1930, 1931, 1935; Biely, 1961; Biely & Fusán, 1967; Siegl, 1978).

Tectonic history of this shear zone includes at least three principal deformation stages. The oldest of them seems to be at least Late Hercynian in age (LHD, pre-Permian) when three different crystalline complexes - the Ľubietová, Hron and Jánov grúň ones - tectonically approached one another as partial nappes of the Hron nappe system (Putiš, 1989). All these three partial nappes along with the metamorphosed (Plašienka et al., 1989) Veľký bok-type Mesozoic, i. e. the Krížna nappe roots, were incorporated to horizontal folds of the north-vergent first Alpine deformation stage (AD1). The folds are several km large and are of regional importance (Zoubek, 1953).

In the horizontal-fold structural style north of Bacúch, the Jánov grúň partial nappe along with part of the overlying Permian - Mesozoic paraautochthonous envelope seems to be a deeper tectonic unit relative to the Hron partial nappe. The Jánov grúň partial nappe represents the core of a horizontal megafold (AD1) which, in the Bacúch Valley, is deeply eroded and overlain by the upper limb of the Hron partial nappe.

This north-vergent structural horizon of the deformation stage AD_{1-1} is unconformably overlain by an upper horizon of the deformation stage AD_{1-2} composed of (the Sihla-type) tonalites of the Vápenica nappe within the Kráľova hoľa nappe system (Klinec, 1966, 1976; Putiš, 1981, 1987b, 1989) which originated south of the Pohorelá line (Fig. 1).

Compressive Alpine AD_2 deformation stage (pre-Upper Cretaceous) gave rise to the superposed NE - SW fold-shear structure bearing slight medium-sized as well as large signs of the "flower structure" - style related also to lateral movements along the Čertovica line.

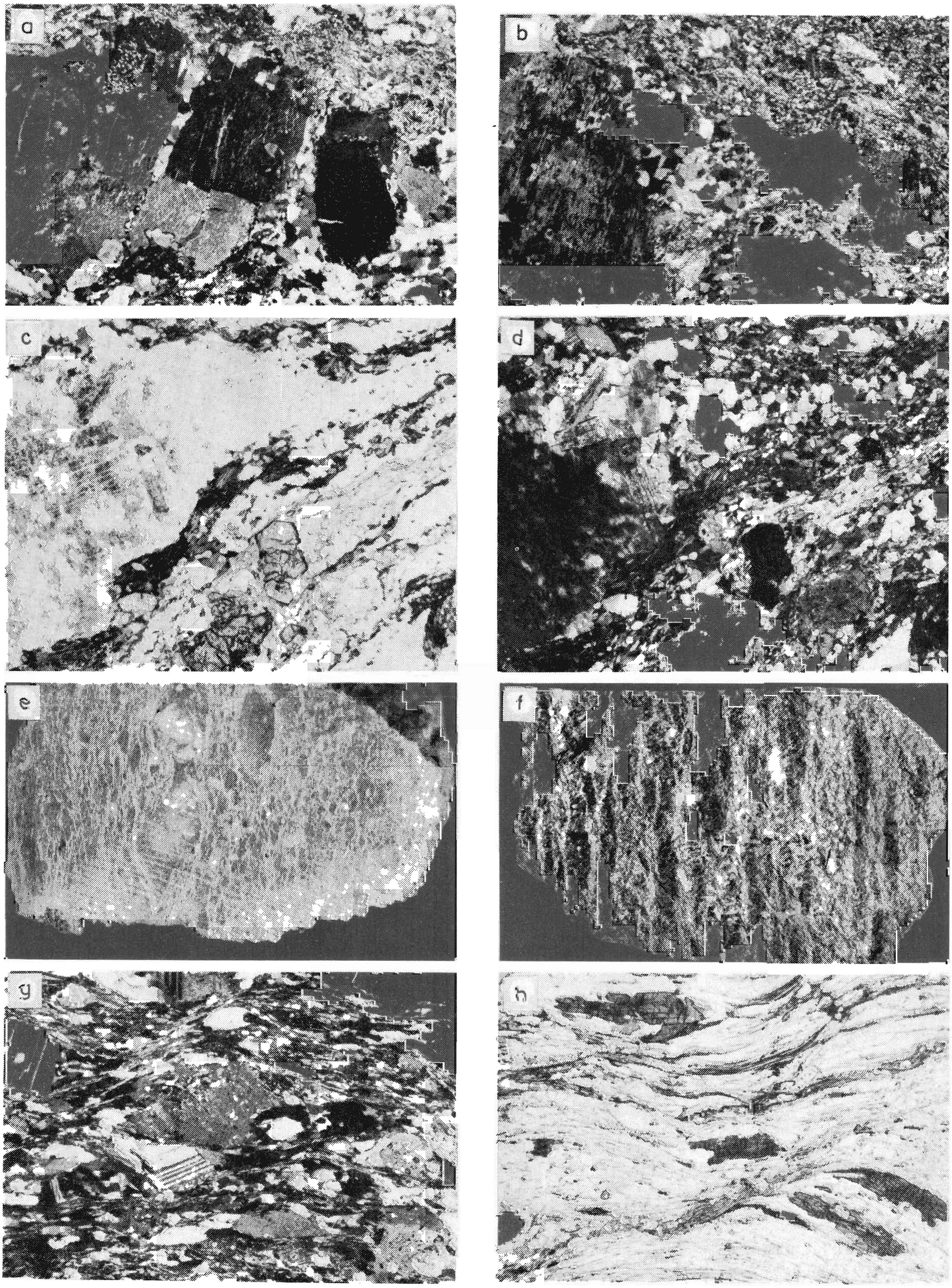
Conceptions accepted so far (Maheľ et al., 1967; Maheľ, 1986) assign the area between the Certovica and Pohorelá lines into the Veporicum. From the parautochthonous Mesozoic view-point, however, the Veľký bok sequence and the Krížna nappe Mesozoic (with thick Jurassic-Cretaceous sediments, shallow-water as well as abyssal facies) converge more to the Tatricum, i. e. the zone of the general Jurassic-Cretaceous extension. It would perhaps be more convenient to regard this area as the southern Tatricum, Subtatricum (sensu Putiš, 1991) rather than the Veporicum whose central part (roughly south of the Pohorelá line) had a clearly different history as early as since the Late Hercynian period (see following text).

Nízke Tatry Mts. - east and the Slovenské rudohorie Mts. - northwest - the Pohorelá shear zone

This region is intersected by the so called Pohorelá line (Fig. 1) interpreted by Zoubek (1930, 1931, 1935) as well as by other authors (Klinec, 1966; Maheľ, 1986) as an intraveporic line, which may be either a reverse-fault plane between the Krakľová (northern) and the Kráľova hoľa (southern) subzones (Zoubek, l. c.) or northward thrust plane of the Kráľova hoľa complex above the Hron complex (Klinec, l. c.).

The study of the inner structure of the originally determined complexes (Klinec, 1966, 1976) has revealed that not only the Hron complex (cf. Miko, 1981) but also the Kráľova hoľa complex (cf. Putiš, 1981, 1982, 1987b, 1989) has a complicated structure and so does the Pohorelá line (Putiš, 1989, 1991; Hraško & Hók, 1990).

Fig. 5. Pohorelá shear zone in eastern Nízke Tatry - Kráľová hoľa massif (a - d) and Lubeník-Margecany shear zone in Slovenské rudohorie (e - h). a - elongated K-feldspar porphyroclast with tension fissures filled with quartz and $biotite_2$ in eyed metagranite, X nicols, magn. 20 x, b - highly albitized K-feldspar porphyroclast, mylonitic quartz, calcite, epidote-zoisite minerals in granite mylonite, X nicols, magn. 20 x, c - on the left is K-feldspar porphyroclast at mylonitization planes with authigenic garnet, $biotite_2$, $sericite-muscovite_2$, epidote-zoisite minerals, ± chlorite in eyed metagranite, // nicols, magn. 35 x, d - same but X nicols, magn. 35 x, e - sinistrally rotated porphyroclasts of quartz and leucocratic granite between C-planes in Permian metaconglomerate, f - kink folds whose B-axis is perpendicular at stretching lineation in Permian metasandstone, g - S-C planes (S - subhorizontal, C - dipping to the left) in Permian metasandstone, X nicols, magn. 35, h - Permian metamorphosed shale with chloritoid, biotite and mus-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The Kráľova hoľa nappe (Fig. 1) composed largely of mylonitized granitoids of the Late Hercynian Vepor pluton was tectonically differentiated into two partial nappes north of the Pohorelá line. The lower Kráľova hoľa partial nappe (more than 1 km thick) is made up mostly of mylonites to augen gneisses (metagranites) that originated from porphyric granites and granodiorites with abundant xenoliths of the metamorphosed gneiss mantle. The upper, Vápenica partial nappe (up to 500 m thick) consists of the Sihla-type metatonalites and exceptional metagranodiorites (Putiš, 1981, 1987b, 1989).

The parautochthonous Permian-Triassic envelope (Struženík or Foederata sequence with Middle Triassic carbonates and Upper Triassic dark calcareous shales) clearly differs from Permian-Triassic envelope sequences in the Tatricum and Subtatricum (with Carpathian continental-lagoonal Keuper in the Upper Triassic). This southern tract of the Central West Carpathians along with the Gemericum and Meliaticum appears convergence to the southern zones that underwent a general Triassic and/or Triassic-Jurassic extension. It is therefore more convenient to assign only the Kráľova hoľa nappe system into the Veporicum (without the Hron nappe system in the north) with the envelope Struženík (Foederata) sequence (cf. Putiš, 1991) and/or also some other units situated to the south of the above-mentioned ones and north of the Lubeník-Margecany line (cf. Bezák, 1988).

The intrusion and crystallization of the northeastern part of the Vepor pluton took place 300 - 270 m. y. ago (300 m. y. - the Sihla type, 270 m. y. - the Vepor and Ipeľ-type) as was determined by U-Pb method on zircons (Bibikova et al., 1990). The solidification stage (270 - 240 m. y.) was characterized by tonalite and granodiorite recrystallization accompanied by the formation of Ti-low biotite, muscovite, epidote, calcite, chlorite, albite, ± garnet as well as associated mylonitization under the conditions of the biotite, locally garnet zone. Mylonitized granitoids are also present in Permian conglomerate intercalations in the Struženík sequence. Low-temperature (up to 300 °C) hydrothermal processes of the Alpine tectonic cycle (93 ± 10 m. y.) resulted in rejuvenization (younging) of Rb-Sr ages of biotites, whereas their K-Ar ages of the post-magmatic Late Hercynian stage remained unaffected (Bibikova et al., l. c.). The latest history of the pluton is indicated by ages of its uplift to the thermal level of some 100 °C more than 80 m. y. ago, as is indicated by fission track measurements on apatites (Kráľ, 1977, 1982).

Structural analysis of the Vepor pluton body distinguished older S-structure (flow structure with xenoliths, general areal anisotropy) striking WNW - ESE and dipping SSW at 10 - 30° that is marked by superimposed ductile-mylonitization S1 planes in the chlorite-biotite, locally even biotite-garnet zone (Fig. 5c, d) with very ductile quartz and ductile-brittle feldspars (Fig. 5a). WNW - ESE L_1 lineation plunging to the ESE at 10 - 30° on S planes can hypothetically be interpreted as a result of the Late Hercynian extension and crustal thinning of the zone where the Veporicum started to exist.

Younger C plane shear system striking ENE - WSW and dipping SSE at 40 - 60° has distinct subhorizontal ENE - WSW stretching lineation L_2. Alpine C planes (Fig. 3e, f) of the Pohorelá shear system are devoid of biotite, and contain only chlorite, sericite, calcite, pumpelleite, quartz and albite.

The Pohorelá line roughly separates the Veporicum from the southern Subtatricum, but a more important line is represented by the thrust plane of the Kráľova hoľa nappe system (Veporicum) over the Hron one (southern Subtatricum).

In the northern and southern Veporicum, stretching lineation is subparallel to basement-nappe boundaries as well as to two regional shear zones - Pohorelá, which roughly separates the Tatricum from the Veporicum, and Lubeník - Margecany, roughly separating the Veporicum from the Gemericum. Both these shear zones clearly deform preexisting structures (Alpine as well as pre-Alpine) and, along with the Čertovica line, seem to have resulted mainly from the Middle Cretaceous transpressive tectonics which put an end to fairly free nappe movements from the south to the north in accordance with the general paleotectonic polarity.

North to northwestern vergency of the nappes is also indicated by regional horizontal folds of the deformation stage AD_1 in the southern Subtatricum as well as by thrusting of the Veporic tonalite nappe over the southern Subtatricum to the dis-

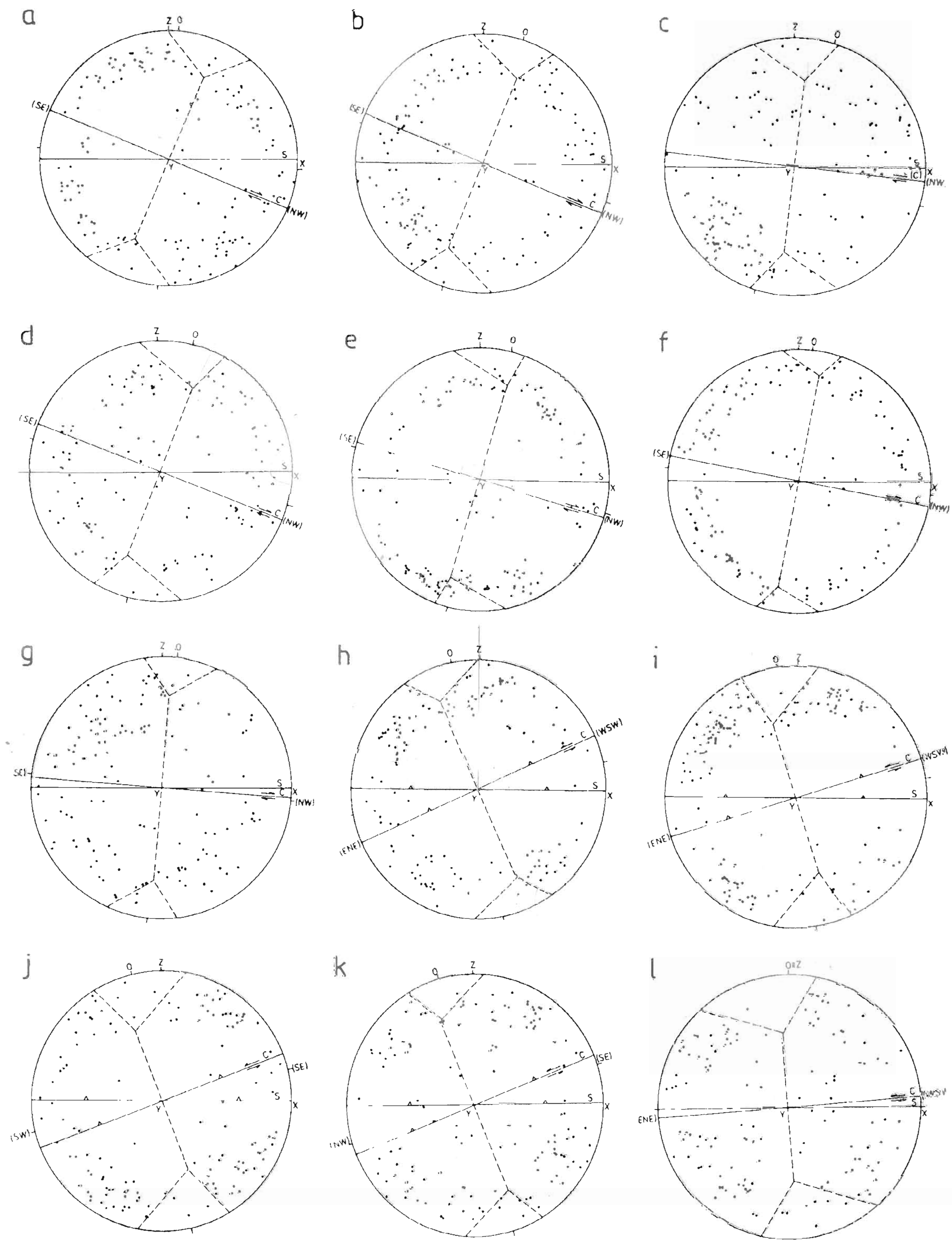
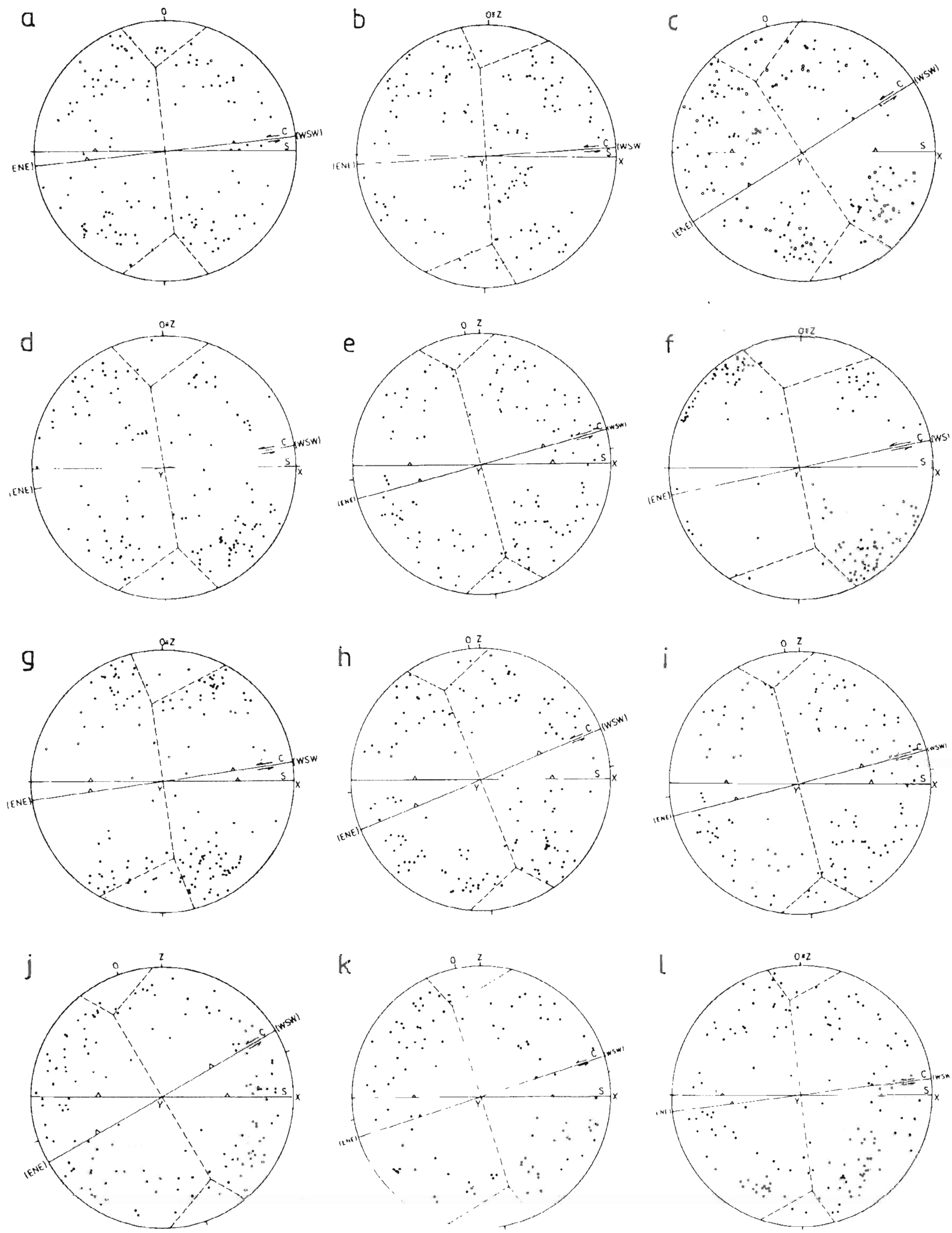


Fig. 6 and 7. Point diagrams of some 150 quartz C-axes measured by U-stage method. O - scale of U-stage N-axis, XYZ - finite strain ellipsoid axes, S - S planes, C - shear C planes. Triangles at S and C planes indicate their dip. Diagrams give γ– quartz type, except for 6 c, g and circles in diagram 7 c; α – quartz type is after Lister and Hobbs (1980). Bo s. z.: 6 a - e - granite mylonite (cf. Fig. 2a, e, f), 6f, g - Lower Triassic quartzite (cf. Fig. 2g); Hr-Zl s. z.: 6h, i - granite mylonite (cf. Fig. 3a); Ra s. z.: 6j, k - granodiorite mylonite (cf. Fig. 3c); Po s. z.: 6 1 - paragneiss of southern Subtatricum,



Po s. z. northern margin, 7a - c - Vepor-type eyed granite mylonite (cf. Fig. 4a - e, g, h, 5a - d); 7d, e - Sihla-type tonalite mylonite (cf. Fig. 3f, 4f), 7f, g - Lower Triassic quartzite (cf. 3e, g); Lu s. z.: 7h - Permian quartzitic metamorphism (cf. Fig. 5g); 7i - Permian arkosic metaquartzite (cf. 5f), 7j - Permian quartzitic metaconglomerate (cf. Fig. 5e), 7k - Permian arkosic metaquartzite, 7 l - quartzitic metaconglomerate.

tance of at least 15 - 20 km (Fig. 1). Stretching lineation of this trend is known in the Malé Karpaty, Považský Inovec and Tríbeč.

Asymmetric structures of S-C-type mylonites (Fig. 4a - d, f) as well as quartz-microstructure models (Fig. 61, 7a - g) determined by the U-stage method prove left shear in the Pohorelá shear zone, i. e. the Veporic block was displaced from the WSW to ENE and the Tatric (southern Subtatric) one from the ENE to WSW (Fig. 1) with assumed lateral displacement of at least several kilometers (cf. 3g, h).

The Slovenské rudohorie Mts., centre - the Lubeník-Margecany shear zone

The Veporic/Gemeric boundary runs along the Lubeník-Margecany shear zone which is of the WSW - ENE direction in the western part (Fig. 1). The Veporicum in the territory concerned is composed largely of envelope Permian metasandstones with exceptional Triassic rocks of the Struženík (Foederata) envelope sequence. S-C-type mylonites of metasandstones (Fig. 5g), quartzitic metasandstones and shales commonly occur. S planes (bedding and metamorphic foliation) strike WNW - ESE, whereas shear C planes and subhorizontal stretching lineation strike ENE - WSW.

This territory has for long been assumed to have undergone the highest-grade Alpine regional metamorphism in the West Carpathians in the chloritoid-cyanite and garnet-biotite zone (Vrána, 1962, 1964, 1966, 1980). Rock studies on profiles across the so called southern-Veporic Permian, however, have revealed only chloritoid and biotite (in addition to other metamorphic minerals: muscovite, chlorite, albite, quartz) bound exclusively to S planes (Fig. 5g, h). Like in the Pohorelá shear zone, C planes here have been found to contain only quartz, sericite-muscovite and chlorite (Fig. 5g, h).

The Alpine regional metamorphism therefore culminated prior to the transpressive formation of the Lubeník-Margecany shear zone. This zone also contains equally deformed leucocratic granites, probably Permian in age. Their contact-metamorphic effects have been studied by Vozárová & Krištín (1985). In our opinion, the Alpine metamorphism was caused by increased geothermic

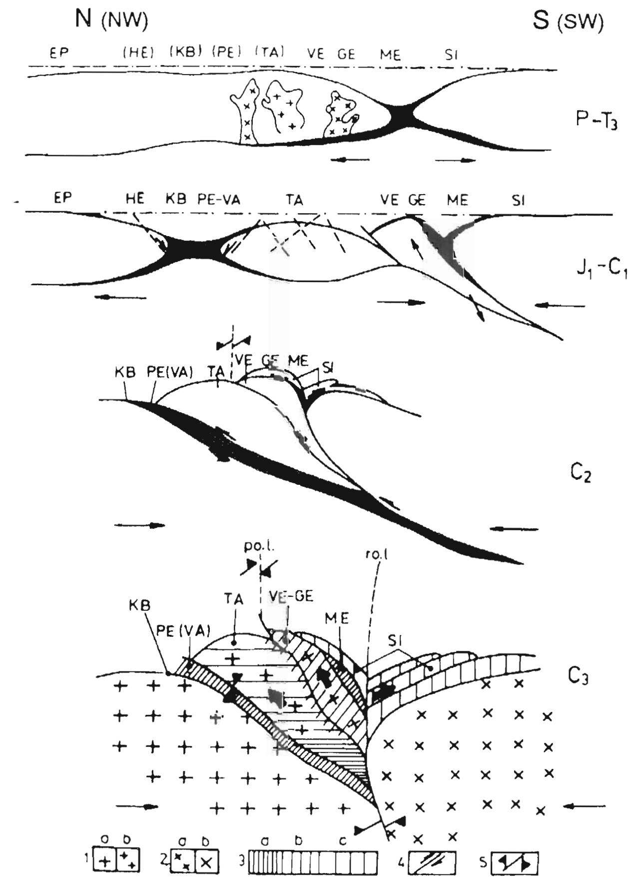


Fig. 8. Paleotectonic-kinematic model of the Western Carpathians during Permian to Late Cretaceous period (Putiš, 1991). Abbreviations: EP - European platform, HE - Helveticum, KB - Pieniny Klippen belt, PE - Penninicum s. l. (incl. Outer Flych zone), VA - Vahicum (Mahel', 1981, 1986), TA - Tatricum, VE - Veporicum, GE - Gemericum, ME - Meliaticum, SI - Silicicum, po.l. - Pohorelá line, ro.l. - Rožňava line. la - crystalline basement folded mainly by Eoalpine folding, 1b - Variscan S-type granitoids,2a - Variscan I-I, S-types of granitoids, 2b - crystalline basement folded by Kimmerian and Eoalpine folding, 3 - grade of the Kimmerian-Eoalpine metamorphism: a - low-, b - low - very, low-, c - very low - diagenesis, 4 - vertical Eoalpine movement direction, 5 - strike-slip.

gradient and tectonic overburden represented by the Gemeric nappe thrusted northward onto the Veporicum and later, probably in part, slipped back to the south.

The asymmetric relationship between S and C planes (Fig. 5e, g, h) as well as quartz-microstructure models in XZ sections (along stretching lineation and perpendicular to C-foliation) suggest left shear in the Lubeník shear zone (Fig. 7h-1).

Considerable differences in the Alpine metamorphism of the Veporicum (as well as the Gemericum and Meliaticum) relative to the Tatricum may also be caused by (partly) different age of their metamorphism which, in zones south of the Tatricum, may be Kimmerian in age in association with closing of Triassic and/or Triassic-Jurassic basins (Fig. 8) such as the Meliata Ocean trough.

Conclusion

The analysis of the investigated shear zones indicates that those in the Malé Karpaty, Považský Inovec, Tríbeč and the central Nízke Tatry, i. e. in the Tatricum and "Subtatricum" are ductile-brittle shear zones formed under metamorphic conditions (at temperature below 300 °C).

The Veporic-Gemeric shear zones (Pohorelá and Lubeník) are ductile shear zones formed under conditions of the chlorite-zone low-grade metamorphism (C-planes). Vepor pluton metagranitoids also contain preserved (Late Hercynian) mylonites of the biotite-garnet zone (S-planes).

Quartz-microstructural models show that C-axes of this mineral are much more preferentially oriented in the Veporicum than in the Tatricum. The principal mechanism to cause this quartz C-axis preferential orientation were slip and rotation on quartz basal planes in single-shear regime (Lister & Hobbs, 1980). The models correspond mostly to planar deformation.

The age of the shear zones decreases towards the north and therefore e. g. the Hrádok-Selec shear zone (Putiš, 1981, 1986b) also includes Middle and Upper Cretaceous sediments, their facies being similar to the Klippen-Belt Klape unit (Kullmanová and Gašparíková, 1982; Leško et al., 1988).

In contrast, transpression in southern zones of the Central West Carpathians could have taken place much sooner, presumably as early as before the Middle Cretaceous folding (e. g. Pohorelá and Lubeník-Margecany shear zone) in association with the Kimmerian tectonics during the closure of the Triassic and/or Triassic-Jurassic basins and ocean troughs (e. g. Meliaticum).

The above data indicate the following polarity and age of the Central West Carpathian shear zones: 1. the Late Hercynian shear zone associated with the Permian-Triassic extension of the Veporicum (as well as the Gemericum and Meliaticum); also related to the Late Hercynian formation of basement nappes in the southern Subtatricum; 2. the Late Kimmerian one (the Pohorelá and Lubeník - Margecany shear zone) - associated with collision-transpressive uplift of the Kimmerian (?) - metamorphosed Veporicum (but also the Gemericum and Meliaticum) to the level of the Tatricum; 3. the Middle Cretaceous one - the Čertovica shear zone (Nízke Tatry), Razdiel shear zone (Tríbeč), Hrádok-Zlatníky shear zone (Považský Inovec), Borinka - Modra shear zone (Malé Karpaty) related to the subduction of the Penninic - Vahic Ocean, transpressive Paleogene tectonics after the closure of the southern and central tracts of the Penninic Ocean. This type probably also contains a zone of highly deformed Paleogene rocks of the Buková depression in the Malé Karpaty (cf. Kováč et al., 1988).

Discussed topics also comprise definition of the Tatric southern margin as the southern Subtatricum (root zone or root area of the Krížna nappe). Similarly, the northern Subtatricum should also be considered. It includes the Považský Inovec-northern-block-type mica-schist crystalline complexes. They bear signs of indistinct Hercynian stabilization, virtually devoid of granitoids, with Permian bimodal-type volcanism known e. g. in the Považský Inovec (Putiš, 1986b) but also in the Tríbeč and the Malá Fatra. Similarly, the Alpine tectonic history and tectonic style of these zones are much more complicated than those observed elsewhere in the central Tatricum.

We cannot rule out the possibility that the so called southern and northern Subtatricum represent the same . Variscan tectonic unit concerning their crystalline basement composed of two different nappes one (migmatites, gneisses, amphibolites, granites) over anotrher (mica-schists, amphibolites, loc. serpentinites) e.g. in the West High Tatra Mts., Považský Inovec, Tríbeč and Low Tatra Mts.

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Petrographic evidence of fluid-controlled strain softening in an orthogneiss along a mylonitic zone at St. Martin im Passeiertal-San Martino in Passiria (Eastern Alps)

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Abstract

Sericitic schists outcropping at the west end of the Penserjoch - Passo Pennes syncline were considered by previous authors as Permoskythian sediments. In the ambit of new field work and related petrographic analyses it has been recognized, that the west end of this syncline is affected by intense deformation and that the sericitic schists outcropping there really are mylonites. Protoliths of these mylonites were granitoid orthogneisses and paragneisses. Microscopical analyses revealed that mylonitization was facilitated by strain softening due to enhanced fluid flow within the basement. The source of fluids is individuated in the Permo-Mesozoic sediments overlying the basement. During mylonite formation temperature should not have exceeded 300 °C approximately.

Fine fractions of sericite from the mylonitic rocks were analysed with the K-Ar method. Although the ages of the < μ2 m fraction largely vary, they suggest that the mylonite zone formed during Lower Oligocene to Upper Miocene.

Relics of a weakly metamorphosed sedimentary cover are preserved in the Austroalpine Meran-Mauls/Merano-Mules basement (Eastern Alps), within a narrow syncline extending from Mauls-Mules in the Eisack-Isarco valley (in the east) to the Fartleis-Valtellese valley (in the west). The eastern and the central parts of this sedimentary cover consist of Permoskythian metaconglomerates, sericitic schists and prevailing Triassic limestones. The sericitic schists, which in the literature are considered to be of Permoskythian age, also outcrop in the western part of this syncline.

Fine fractions from the sericitic schists of the eastern and central part were analysed by Frank et al. (1977) with the K-Ar method. Modal ages in the range 15-20 Ma were obtained by these authors, and were interpreted to support the existence of a Tertiary metamorphic overprint in this area.

New detailed field analyses of the western part of the Penserjoch-Passo Pennes syncline were carried out by the author within the framework of a geochronological, structural and petrographic investigation. The main aim was to ascertain the Alpine metamorphic and deformational history of this area, taking into special consideration the sericitic schists. In the ambit of this field work and related petrographic analyses, an unexpected result has been obtained: the sericitic schists are not Permoskythian sediments as assumed in the literature, but mylonites. The description of their main features, protholits, genetic mechanism and chronology is the aim of the present report.

These mylonites make up a NE-SW striking mylonitic horizon. This is more than 400m thick and crosscuts the Fartleis-Valtellese valley immediately to SE of the Fartleisalm.

Petrographic analyses show that protoliths of these mylonites were granitoid orthogneisses and paragneisses. However, it cannot be excluded that Permoskythian phyllites also occur, at least in those parts where intense deformation do not further permit with certainty the protolithic nature to be recognized.

Because of strain partitioning in a meter scale, especially at the marginal part of the mylonitic horizon, it can be observed that schists successively richer in sericite anastomose lenses of relictic orthogneisses: therefore a complete transition can be observed from domains with dominantly progressive shortening strain to domains with dominantly progressive shearing-only strain (Bell et al., 1986). Detailed microscopical analyses revealed that mylonitization was facilitated by strain softening due to enhanced fluid flow within the basement. In

TAB. 1

K-Ar data

Sample	Lithology/sample locality	Grain Size	$\%K_2O$	$^{40}Ar_{rad}ccm10^{-6}$	%rad	Model age Ma	Analysed sample
1/937	Mylonite/Fartleisalm	< 2 µm	4.80	5.66	55	30±2	S
2/939	"	"	4.90	5.39	51	28±2	S
3/945	"	"	4.20	5.88	56	35±2	S
4/946	"	"	5.00	6.12	64	31±2	S
5/947	"	"	6.40	4.84	13	19±6	S
6/944	"	"	4.70	4.21	14	22±6	S
7/943	"	"	3.80	3.76	33	25±3	S
8/945	"	2-6 µm	4.20	6.43	64	39±2	S
9/946	"	"	4.40	7.99	67	46±3	S
10/944	"	"	3.90	5.57	54	36±3	S
11/943	"	"	3.90	5.94	73	38±3	S
12/947	"	"	3.90	6.47	69	42±2	S
13/946	"	6-11 µm	3.70	7.93	71	53±3	S
14/945	"	"	3.60	6.84	75	48±3	S
15/934	Augen-gn./St.Martin	150-250 µm	8.60	92.58	97	274±11	W
16/934	"	"	6.80	33.44	85	125±6	B
17/935	Granite-gn./St.Martin	"	7.90	76.22	97	246±16	W
18/933	Mica sch./St.Leonhard	"	6.60	21.50	83	84±4	W

S - sericite, W - white mica, B - biotite

fact, within the shortening domains in which protoliths were preserved from advanced mylonitization, it is possible to ascertain that the development of the mylonitic foliation is spatially related to sericitized feldspar domains and took place after the fluid-controlled sericitization of feldspars. In particular, it can be observed that sericitization of feldspars proceeded along the main pre-mylonitic schistosity (which is mostly delineated by quartz-feldspar bands). The mylonitic foliation developed at a low angle to the main schistosity, preferentially crosscutting these intensively sericitized feldspar domains within the quartz-feldspar bands. As consequences:

i) sigmoidal, highly strained quartz bands formed, with mostly sutured grain boundaries and already beginning nucleation of new grains;

ii) sericitized feldspars were completely deformed, leading to sericite-rich domains;

iii) old white mica flakes were affected by grain size reduction, contributing to sericite formation. Microprobe analyses revealed that both generations of white micas (i.e. the original large flakes and the new fine-grained sericite) are phengitic in composition. Whereas this composition corresponds to a typical feature of the Alpine white micas, it is not surprising for old, probably Hercynian white micas from granitoid rocks: the occurrence of K-feldspar implies high phengite contents in the coexisting white mica, as pointed out by Guidotti & Sassi (1976, p. 125).

With respect to the source of the fluids responsible for the sericitization of the feldspars, the Permo-Mesozoic sediments overlying the basement can reasonably be admitted to have released a significant amount of fluids (during their diagenesis, and then also during their metamorphism): therefore, we can speculate either that before the Tertiary mylonitization these sedimentary sequences also underwent Cretaceous metamorphism, or that release of fluids occurred during a Tertiary tectonometamorphic reworking of this area.

Temperature during mylonite formation should not have exceeded 300 °C approximately. This is clearly indicated by petrographical data including: intensively sericitized feldspars, generally sutured grain boundaries of strained quartz crystals, formation of very small subgrains within highly strained quartz bands. A consistent indication is supplied by the isotopic data, both those available in the literature and the new K-Ar data (Tab. 1). Hercynian cooling ages were obtained from biotites of micaschists outcropping near the Penserjoch-Passo Pennes syncline (Thöni, 1980), whereas a partial Ar and Sr loss (Del Moro et al., 1982) was recognized only within a narrow area along the boundary between the basement not affected by the Alpine overprint (to the south) and the basement in which an amphibolite facies Alpine overprint (to the north) occurs.

In order to determine the age of the mylonitic event, fine fractions of sericite from the mylonitic rocks were separated and analysed with the K-Ar method. As can be seen from Table 1 (samples 1 - 14), there is an evident correlation between grain size of the sericite and its age: the model ages the <2μm fraction fall in the range 20-35 Ma, those of the 2 - 6μm fraction fall in the range 35-45 Ma, whereas the two samples of the coarse fraction (6-11μm) gave ages around 50 Ma. Probably, the temperature conditions prevailing during the formation of the mylonite zone were too low to allow the radiogenic Ar to be completely released from the >2μm sericite fractions. Consequently, an unambiguous indication of the time of deformation cannot be simply derived from the age data of the different fractions. However, although the ages of the <2μm fraction largely vary, they suggest that the mylonite zone formed during Lower Oligocene to Upper Miocene, probably related to the development of the Periadriatic Lineament.

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Is the present plate tectonic setting of the Arabian Plate (Greenland of Africa) comparable with the Early Tertiary evolutionary state of the Greenland Plate (Arabia of South America) ?

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Abstract

Due to the eccentric rotation of the spinning Earth around the gravitational centre of the Earth-Moon (-Sun)-system, the Earth presumably behaves like a gigantic hypocycloid gearing. As a result, the lower mantle, the Pacific basin, and the oceanic states of the Wilson or Oceanic Cycle (Rift/Red Sea state, Atlantic state, Pacific state, Collision/Himalayas state), being arranged from East to West through 360° around the globe, are constantly displaced eastwards relative to the Earth´s crust or the convecting upper mantle-crust system, once around the globe in 200 to 250 m. y. The continents migrate westwards around the Pacific basin both in the North and South. They collide at the equator to the West of the Pacific, continuously adding segments to an eastwards growing collisional mountain belt (sequential collisions; zip fastener principle), that becomes older towards the West. Since the Permian, this belt has lapped some 1 1/4 to 1 1/3 times around the globe in the northern hemisphere, in the form of a spiral.

During the course of its eastward displacement the Pacific retains approximately the same size; it does not shrink. What is lost at the stern from the Pacific basin in the West through sequential collisions between the island arcs, microcontinents and continents from the northern and southern hemispheres in the Asia, Australia and equatorial West Pacific backarc basin settings (N-S compression), is added at the bow in the East by sequential openings of a series of Caribbeans and Mediterraneans (N-S extension; rift propagation towards the East; zip fastener principle), withdrawal of the continents in the North and South America, Europe and Africa settings towards the Poles and by incorporation of a succession of Atlantics. Thus, a Mid-Atlantic Ridge will slowly be replaced by an East Pacific Rise, an Azores-Gibraltar Fracture Zone by a Galapagos Rift Zone; a Mediterranean will develop into a Caribbean and a Black Sea into a Gulf of Mexico. Arabia (the Greenland of Africa ?) should continue to force its way North through the Alpine Belt and the Russian Platform of Eurasia. It should in future arrive at a plate tectonic setting comparable to that presently occupied by Greenland (the Arabia of South America ?). After having disengaged from Asia in the near geological future, Europe will rotate clockwise as North America has done during post-Laramide times since the Eocene/Oligocene. A Western Indian Ocean will become a South Atlantic, a Red Sea-Gulf of Aqaba-Levant Fault a Central/North Atlantic-Baffin Bay and a Persian Gulf-Mesopotamian Plain a North Atlantic. Like the Ocean states of the Wilson Cycle, plate tectonic characteristics move eastwards with the angular velocity of the Pacific basin and of the lower mantle. Relative to a distinct plate tectonic setting of an ocean, a continent or continental margin, a future or later evolutionary state at the Earth´s surface is always depicted in a setting simultaneously developed further to the West and a past or earlier state in a setting simultaneously occurring further to the East.

In the foreland, to the South of the North Greenland Fold Belt (Ellesmerian Orogeny, Innuition Orogen: Late Carboniferous and earlier; Eureka Orogeny/Orogen: Late Cretaceous/Early Tertiary) hydrocarbon accumulations could be expected in quantities comparable with the southern foreland deposits of the Himalayan-Alpine-Laramide-Kimmerian (-Atlas-Caribbean Andes-West Antarctica-SW-Pacific microcontinents) Belt of New Guinea, Indonesia, Burma, Bangla Desh, India, Pakistan, Arabian Plate, North Africa, northern South America, (East Antarctica ?; SW-Pacific backarc basins - Lord Howe Rise/Challenger Plateau) and California.

Introduction

Two major oceans in succession (Paleo-Tethys farther North; Neo-Tethys to the South) separated the southern continent Africa/Arabia (part of West Gondwana) from the northern megacontinent Eurasia, during the Mesozoic. The western part of the southern margin of Eurasia had previously been deformed by the Variscan diastrophism (e. g. Kopeth Dagh belt and the southern Caspian area in the eastern part of this segment; Flügel, 1964; Brinckmann, 1976; Lensch et al., 1984). During the Late Triassic, the Iranian Plate, formerly like the Turkish Plate part of the northern margin of Gondwana (identical with Sengör´s "Kimmerian continent" or Stöcklin´s "Central Domain"), collided with Laurasia (Indosinian orogeny) and formed the Elburz Fold and Thrust Belt (North Iran suture; closing of the Paleo-Tethys). Following this collisional event the Iranian Plate formed part of Laurasia. Shortly after the Indosinian closure of the Paleo-Tethys, the Iranian Plate separated from Gondwana along a Mediterranean/Caribbean type rift ocean. Through continuous withdrawal of Gondwana to the South the Pacific-wide Neo-Tethys Ocean opened to the South of the Iranian Plate during the Jurassic. Simultaneously, Kimmerian movements deformed Northern and Central Iran (Pacific type, subduction-related deformations). During Kimmerian to Late Kimmerian times (Late Jurassic to Early Cretaceous) part of the Iranian Plate (Central East Iran Microplate) broke away again from Laurasia in a NW-Pacific type backarc basin, microcontinent and island arc setting. After the closing and obliteration of the Neo-Tethys Ocean during the Late Cretaceous/Early Tertiary and the renewed collision between Gondwana and Laurasia/Iranian Plate (closing of West Pacific type backarc basins; collisions between continents, island arcs, microcontinents and oceanic crust from NW- and SW-Pacific type backarc basins; ophiolite/oceanic crust obduction, oceanic crust subduction, thrust faulting; continent, microcontinent and island arc suturing) during the Tertiary (Alpine-Himalayan orogeny), all that currently remains from the former Pacific-wide Paleo- and Neo-Tethys Oceans and NW- and SW-Pacific type backarc basins are remnants from their oceanic and backarc basin crust in the form of ophiolite sequences, scattered and locally aligned along the suture zones between the collision fragments of continental crust. The two-sided Tertiary Alpine-Himalayan collisional mountain belt, which surrounds and occupies the Iranian and Turkish Plates, can be subdivided into a northern branch with the Indosinian Paleo-Tethys suture (North Iran suture) along the two-sided Late Triassic Elburz Fold and Thrust Belt (Kopet Dagh, Elburz Mountains of Iran; Lesser and Greater Caucasus Mountains of Armenia and Aserbaidshan, Pontides Thrust Belt of Turkey; intermediate complex folded mountain system on the Iranian Plate to the South, that resulted from Indosinian and Alpine microcontinent and island arc collisions) and a southern branch between the neotethyan margins of Laurasia/Iranian-Turkish Plates in the North and Arabia/Africa or West Gondwana in the South (Semail Ophiolite Nappe of the Oman Mountains of Arabia; the two-phase Late Cretaceous/Early Tertiary and Late Miocene/Pliocene Zagros Fold and Thrust Belt of Iran with the Mio-/Pliocene Zagros Thrust Zone in the South; Bitlis Fold and Thrust Belt of Iraq and Turkey; Taurides Fold and Thrust Belt of southern Turkey with the South Anatolian Thrust Zone; Troodos Ophiolite Complex, Cyprus; etc.) Wellman, 1965; Stöcklin, 1977; Sengör, 1979, 1985, 1986, 1987; Sengör et al., 1985; Sengör & Yilmaz, 1981; Scott, 1981; Khain, 1984; Lensch & Schmidt, 1984; Lensch et al., 1984; Rotstein, 1984; Schmidt & Soffel, 1984; Cherven, 1986; Dercourt et al., 1986; Dewey et al., 1986; Snyder & Barazangi, 1986; Ben-Avraham, 1987; Jancovic & Petracheck, 1987; Dewey, 1988; Domas, 1988; Searle, 1988; Bushara, 1989; Cisternas et al., 1989; Papanikolaou, 1989; Philip et al., 1989).

Major hydrocarbon occurrences accumulated in the northern and southern forelands of the Alpine-

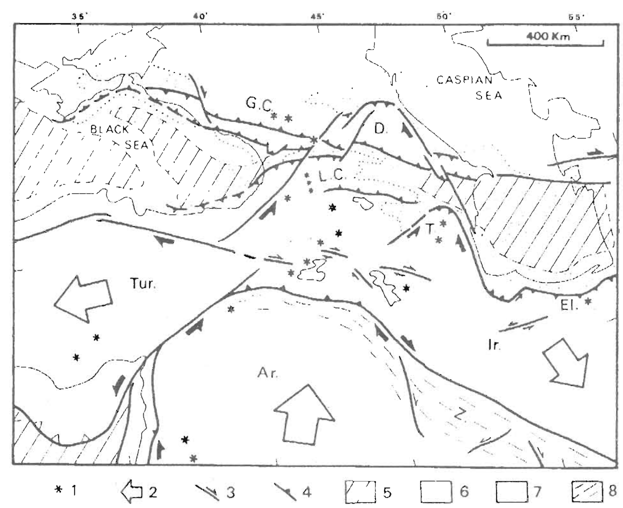


Fig. 1. The Arabian Plate (according to Philip et al., 1989) forcing its way North with its wedge-shaped northern front along the N-S-trending sinistral Aqaba-Levant fault and the dextral Zagros fault through the Alpine-Himalayan collisional mountain belt into the Russian Platform.
Present-day tectonic features characterizing the areas surrounding the Caucasus. 1 = recent volcanoes; 2 = relative motion with respect to Eurasia; 3 = major strike-slip falts; 4 = major thrust faults; 5 = oceanic or intermediate crust; 6 = continental crust; 7 = main sedimentary basins; 8 = recent folding at the border of the Arabian plate. G.C.-Great caucasus; D-Dagestan; LC-Lesser Caucasus; T-Talesh; El.-Elbros; Ir.-Iranian block; Tur.-Turkish block; Ar-Arabian plate; Z-Zagros; K-Kasbeg volcano; E-Elbrus volcano

Himalayan orogenic belt during syn- and post-collisional times. During the Late Tertiary (Late Oligocene, Mio-/Pliocene; from 32 to 24 m. y. onwards) the opening of the Red Sea (oceanization during the Pliocene) separated the Arabian Plate from Africa (Bonatti, 1985; Ben-Avraham, 1987; Bonatti & Seyler, 1987; Bayer et al., 1989; Voggenreiter & Hötzl, 1989; May, 1991). The Arabian Plate has continued up to the present to force its way North along the N-S trending sinistral Gulf of Aqaba-Levant transcurrent fault and to penetrate with its wedge-shaped northern front into the Alpine-Himalayan collisional mountain belt and into the Russian Platform of Laurasia (Fig. 1). Arabia´s still ongoing northward drive is responsible for the young Mio-/Pliocene movements along the active Zagros Thrust Zone, along which the Arabian Plate is subducted or underthrusted beneath the Iranian and Turkish Plates of Laurasia. In particular the tectonically complex Lake Van Region of Eastern Anatolia in the southern branch of the Alpine-Himalayan belt is a conjunction of strike-slip faults parallel to the trend of the Alpine belt, with a NNE-trending zone that extends northward from the Gulf of Aqaba-Levant fault (Fig. 1). As a consequence of the continuous northward push of the impinging Arabian Plate, the Turkish block withdraws laterally westwards along the dextral North Anatolian strike-slip fault and the Iranian block is extruded laterally along the Zagros fault relative to Arabia towards the East (Fig. 1).

The stress accumulating through the northward movement of the Arabian Plate in the East and the different plate tectonic behaviour of the Levant/Mediterranean Region to the West is compensated along the N-S trending sinistral Aqaba-Levant strike-slip fault. Whereas Arabia has continued its northward drive up to the present, Africa on the other hand has reversed its former northward thrust since Mio-/Pliocene times (some 22 to 8 m.y. ago) and is rifting South, away from Eurasia and Arabia. The former compressive regime in the Alpine-Mediterranean Region has given way to a tensional regime accompanied by thinning of the continental crust, by tectonic denudation/erosion or extensional unroofing/exhumation and isostatic rise of metamorphic core complexes along low-angle detachment faults and finally by the exhumation of oceanic crust (upper mantle; oceanization; e.g. Tyrrhenian Sea) (Selli, 1981; Wernicke, 1981; Cello et al., 1985; Reuther & Eisbacher, 1985; Weijermars, 1985, 1987, 1988; Channell, 1986; Cello, 1987; Lister et al., 1987; Pawlides & Mountrakis, 1987; Davis, 1988; Dewey, 1988; Hashida et al., 1988; Laubscher, 1988; Doblas & Oyarzun, 1989; Trurnit, 1991d).

The principal aim of this paper is to establish whether or not all of these seemingly chaotic, uncoordinated movements of plates along strike-slip faults, towards suture/obduction/subduction zones or away from rifts/zones of seafloor spreading in this key area in the southern part and to the South of the Alpine-Himalayan collisional mountain belt can nevertheless be fitted into a greater plan and be comprehended as part of a long-term physical process or a reconstructable sequence of events, thus assisting in the unravelling of the convective pattern of the Earth. If it is possible to prove that all plates on the globe move according to a long-term, stable, predictable convective pattern, the secondary aim of this paper is to predict the future plate tectonic behaviour of the Arabian Plate and, through the method of comparative plate tectonics, possibly to find one or several plates that have already passed through a similar sequence of plate tectonic settings, through which Arabia has passed and is preparing to pass in the near geological future (50 to 70 m. y. to reach the present setting of the nearest forerunner-plate).

The concept of the eastward migrating global tectonic megacycles presented in this paper integrates most of the current version of the plate tectonic model (Trurnit, 1983a-c, 1984a-d, 1985a-d, 1986a-d, 1987a-c, 1988a-d, 1989a-e, 1990a-c, 1991a-d). However, in order to eliminate major inconsistencies in the presently accepted model, a new dynamic aspect is introduced: the eastward displacement of the Pacific basin and of the additional ocean states of the Oceanic or Wilson Cycle (Rift/Red Sea state, Atlantic state, Pacific state, Collision/Himalayas state; Fig. 2) relative to the Earth´s crust. It has also been found convenient to give new connotations to some terms already introduced: *Pangaea* is used for all continental crust, past and present. *Tethys* is taken as synonymous with the Pacific. The circum-Pacific ring of sub-

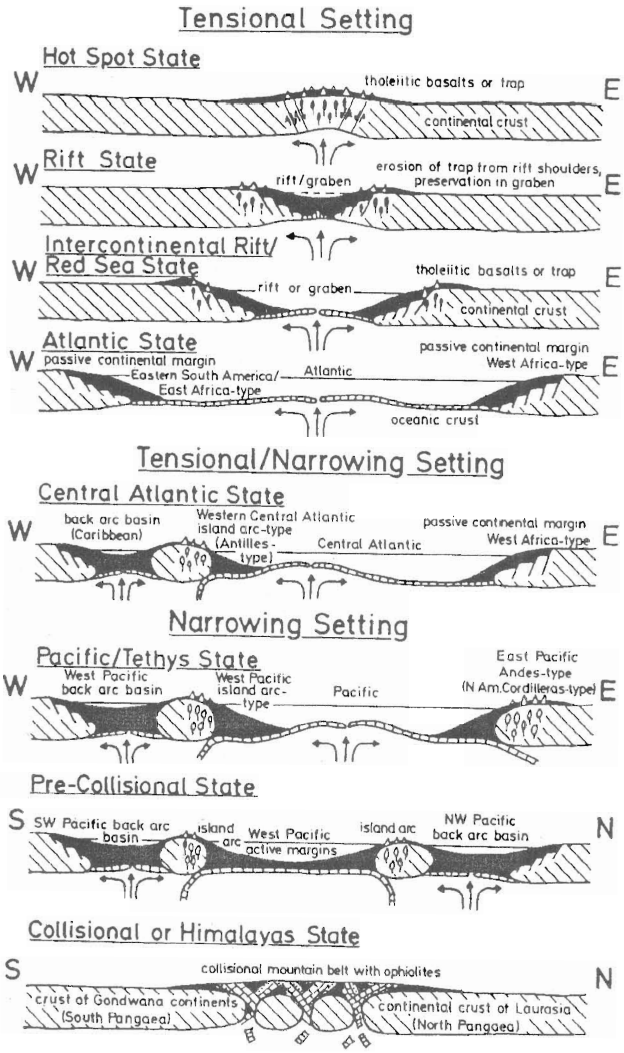


Fig. 2. The Oceanic, Eugeosynclinal or Wilson Cycle (Rift/Red Sea state, Atlantic state, Pacific state, Collisional/Himalayas state) describing the opening and closing of an ocean (for Mediterranean/Caribbean type rift states see Fig. 3c, 13, 15, 16).

duction zones separates a *Pacific or Tethys* area with mainly active continental margins from a *Continental or Pangaea area with Intra-Pangaea Oceans* (Atlantic, Red Sea/Indian Ocean, Arctic Ocean, Circum-Antarctic Ocean, etc.) and mainly passive continental margins. The Pangaea area in turn is subdivided into a *North Pangaea* area and a *South Pangaea* area with the North Pangaea and the South Pangaea continents broadly distributed over the northern and southern hemispheres.

The North Pangaea and the South Pangaea area are separated by the West-East-trending, two-sided, bilaterally verging Himalayan-Alpine collisional mountain belt of Laurasia, that connects the West and East Pacific through the Pangaea area. This belt splits West and East of the Pacific into one-sided, unilaterally verging mountain belts, that circle the Pacific in the North and South. Relative to this globe-circling mountain system, the Pacific is a hinterland and the North and South Pangaea areas are forelands of enormous extent. Vergence is towards the foreland and away from the hinterland.

Orogenic belts form between the margins of two colliding continents at the end of the Oceanic or Wilson Cycle of ocean opening and closing (Fig. 2). During the opening phase of the cycle an ocean passes with the aid of seafloor spreading through the Rift/Red Sea and Atlantic states with mainly passive continental margins and during the closing phase of the cycle with the aid of subduction/obduction through the Pacific and Collisional States with mainly active continental margins.

The Wilson Cycle shows the narrow Red Sea type oceans as being the forerunners of wide Atlantic type oceans and the Rift or Atlantic type oceans with passive margins pre-dating Pacific type oceans surrounded by subduction zones or active margins. However, the Wilson Cycle does not explain the sequence and relative age of the margins East and West of the Pacific (Trurnit, 1987b). Do one-sided Cordilleran and Andean type margins to the East of the Pacific pre- or post-date one-sided West Pacific island-arc type margins ? Or can both one-sided types independently of each other be forerunners of two-sided collisional mountain building ?

The Extended Plate Tectonic Model presented here demonstrates that throughout Earth´s history, continental margins and oceans have always passed through similar sequences of plate tectonic settings between their birth in a rift and their death in the collision zone West of the Pacific. The arrangement of the present-day ocean types, passive and active continental margins and of the Alpine collisional mountain belt between South Asia and Europe, that exhibits a westward directed sequence of gradual evolution and ageing, represents an actualistic scenario for the pre-, syn-, and post-collisional evolution of older collisonal mountain belts, vanished oceans and former continental margins.

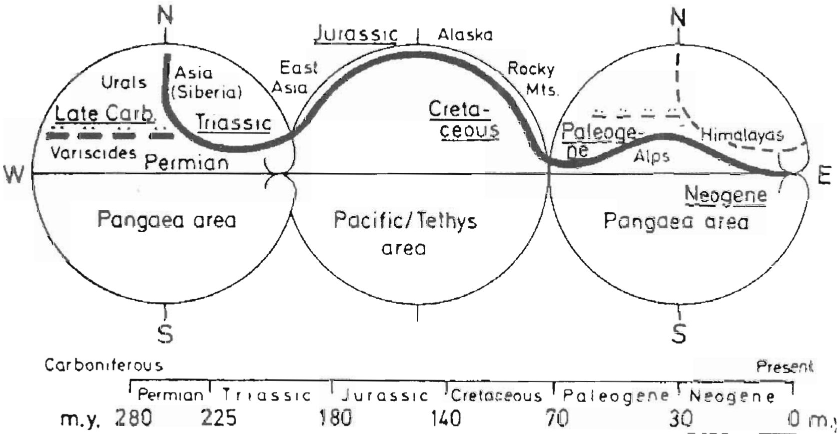
Eastward displacement of the Pacific basin relative to the Earth´s crust and the collisional mountain belt spiral of the Late Paleozoic-Mesozoic-Cenozoic North Pangaea

Six global tectonic phenomena indicate an eastward displacement of the Pacific basin (not the Pacific crust) relative to the Earth´s crust (a westward displacement of the Earth´s crust relative to the Pacific basin):

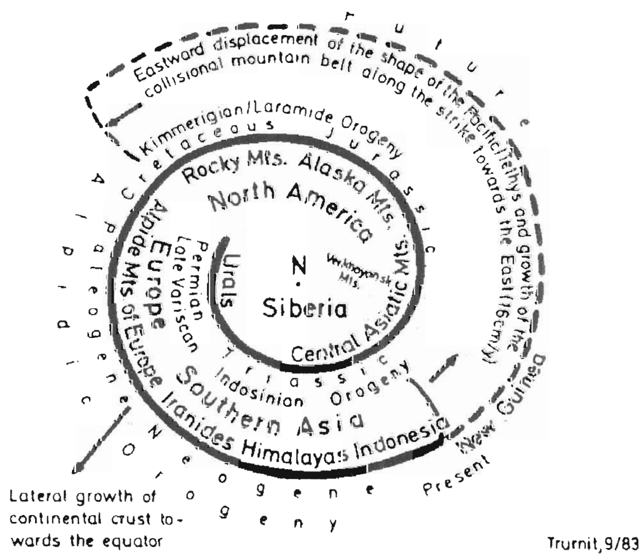
1. *The orogenic or collisional mountain belt that starts at the equator to the West of the Pacific Ocean and extends West* from New Guinea/Indonesia (collision still taking place) via the mountain ranges of Vietnam, Thailand and Burma, the Himalayas, the Karakorum, the Transhimalayas of Tibet and Sinkiang, the Central and South Pamirs, the East Hindukush (Neogene collision), the mountain ranges of South Afghanistan and of Pakistan, the Oman mountains of Arabia, the Iranides (Elburz mountains; complexly folded mountain systems of the Iranian Plate; Zagros Fold and Thrust Belt), the mountain ranges of Turkey and Cyprus (Pontides; Central mountain ranges of the Turkish Plate; Taurides; Troodos Ophiolite Nappe) (Tertiary collisions), the Alpine mountain chains of Europe (Late Cretaceous/Paleogene Alpine collisions) (Himalayan-Alpine collisions: closed Neo-Tethys Ocean), the North American Cordilleras/Rocky Mountains (Late Cretaceous Laramide collision), the mountain ranges of Alaska and Northeast Asia (South Anyuy; Olazeia Oloi; Chersky-Verkhoyansk) (Jurassic/Early Cretaceous Kimmerian/Late Kimmerian-Early Laramide collisions) and via some East Asiatic mountain ranges (East Sachalin; Japan: Hokkaido Central Belt; Hida Belt of Honshu - Triassic Akiyoshi collision; Sikhote-Alin, Mongolo-Okhotsk-Chukotsk Belt) (Late Triassic/Early Jurassic Late Indosinian to Early Kimmerian collisions), parts of SE-Asia (Loei Fold Belt of Thailand), parts of the Central Asiatic Mountains (Sungpan Kantze and Sinkiang Fold Belt Systems, Nan Shan, Qin-Ling Shan, Bayan Har Shan, Anyemaqen Shan, Kun Lun, Karakorum, Altyn Tagh, North Pamirs, Hindukush, Paropamisus, Elburz Fold and Thrust Belt of Iran, etc.) (Triassic Indosinian collisions) (closed Paleo-Tethys Ocean) to the Urals (Permian collision) (Stille, 1924, 1945; Gansser, 1973; Burchfield & Davis, 1975; Dickinson, 1977, 1979; LeFort, 1975; Molnar & Tapponier, 1977; Stöcklin, 1977; Sengör, 1979, 1985, 1986, 1987; Churkin & Trexler, 1981; Kosygin & Parfenov, 1981; Mitchell, 1981; Sonnenfeld, 1981; Li et al., 1982; Fujita & Newberry, 1983; Natalin & Parfenov, 1983; Nur & Ben-Avraham, 1983; Saleeby, 1983; Stauffer, 1983; Khain, 1984; Kovalenko et al., 1984; Zonenshain et al., 1984; Howell et al., 1985; Mattauer et al., 1985; Hahn et al., 1986; Kimura & Tamaki, 1986; Tapponier et al., 1986; Tobisch et al., 1986; Crowell, 1987; Hamilton, 1987; Jayko et al., 1987; Searle et al., 1987; Weijermars, 1987, 1988; Boulin, 1988; Eisbacher, 1988; Jolivet et al., 1988; Sano & Kanmera, 1988; Xingyuan, 1988; Coleman, 1989; Drewes, 1991). This belt is divided into segments by the ocean states of the Wilson Cycle and in terms of the collision ages it becomes progressively older towards the West (Fig. 3a). Since the Permian it has wound approximately 1 1/4 to 1 1/3 times around the Earth, i. e. the cratonic nucleus of Laurasia or North Pangaea, in the form of a spiral (Fig. 3b).

In order to explain the existence of this belt, the continents must migrate westwards around the Pacific basin both in the North and South (the Pacific basin must be displaced eastwards between the continents of the northern and southern hemispheres). The continental or Pangaea area splits permanently to the East of the Pacific in the Mediterranean and Caribbean settings (rift propagation towards the East) and closes permanently at the equator to the West of the Pacific by means of sequential collisions of the continents from the northern and southern hemispheres in the collisional setting (progressively later closing towards the East), continuously adding segments to the collisional mountain belt, which grows towards the East according to the zip fastener principle (Fig. 3c). During their formation the individual segments of the collisional mountain belt were aligned approximately parallel to the equator. The ophiolites in the suture zones of the Late Paleozoic-Mesozoic-Cenozoic North Pangaea collisional mountain belt and of all older orogenic mountain belts appear to mark the track of the eastward migrating Pacific basin (modern Tethys Ocean, Iapetus Ocean, etc.) (Trurnit, 1983c). Contrary to a postulation of Stille

a) Eastward Migrating Orogenic Paroxysm in the Collisional Mountain Belt Forming since the Permian



b) Collisional Mountain Belt Spiral of the North Pangaea forming at Present (Orogens, Orogenies)



c) The "Tethyan Zipper" : Origination of the Parallel to Equator Collisional Mountain Belt Spiral of the Late Paleozoic - Mesosoic - Cenozoic North Pangaea According to the Zip Fastener Principle

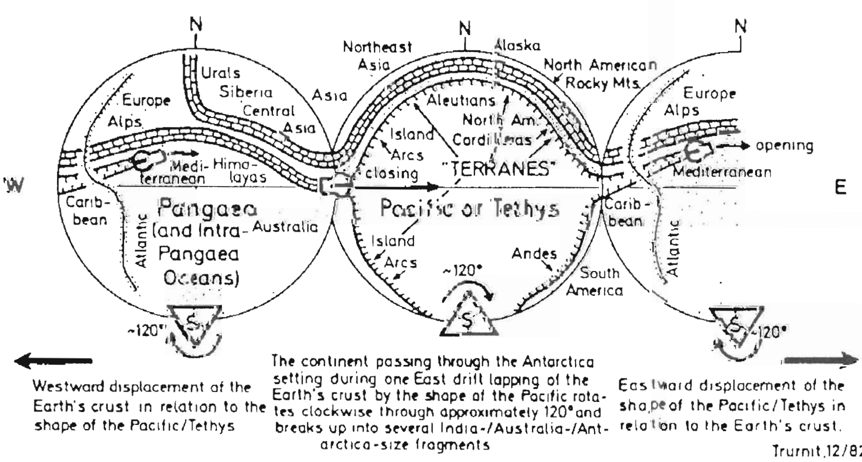


Fig. 3. Late Paleozoic-Mesozoic-Cenozoic North Pangaea collisional mountain belt spiral. a - Eastward advancing orogenic activity in the collisional mountain belt forming since the Permian at the equator to the West of the eastward migrating Pacific/Tethys; equatorial view; b - North polar view; c - Eastward displacement of the Pacific basin relative to the Earth's crust; opening of Mediterraneans/Caribbeans to the East of the Pacific (rift propagation towards the East; zip fastener principle) and continuous formation of the collisional mountain belt to the West of the Pacific through sequential collisions of the continents from the northern and southern hemispheres (zip fastener principle). Some 120° clockwise rotation of the continent from the South Pangaea/Gondwana breaking up in the Antarctica setting during one East drift lapping of the Earth's crust by the Pacific basin. Island arcs are pre-collisional, tectonostratigraphic terranes post-collisional phenomena.

(1924, 1945) which is still defended by Schwan (1984), two or more orogenic phases or continent/continent collisions can neither be worldwide occurring, simultaneous and coeval events nor can they be confined to definite and short-term intervals between longer eras of tectonic quiescence (Gilluly, 1949; Rutten, 1949; Oxburgh, 1974; Dennis, 1982; Sengör, 1985, 1986, 1987). There is rather continuous collision between an endless sequence of continents from the northern and southern hemispheres in only one setting and that is at the equator to the West of the eastward migrating Pacific basin (Fig. 3).

2. *Lateral continental growth by cyclically repeated orogenic events taking place at different times for the individual longitudes and continental margins* (Gastil, 1960). Examples are the Late Permian/Early Triassic collision of the Southern Urals succeeded by the Tertiary Ira-

nides collision, the Triassic Indosinian collision of Central and South Asia (closed Paleo-Tethys) succeeded by the Neogene Himalayas collision (closed Neo-Tethys) (Fig. 3), the Late Cambrian/Early Ordovician Ross collision followed by the Late Devonian/Early Carboniferous Borchgrevink collision in Antarctica, the Cambrian Delamerian-Tyennan collision succeeded by the Devonian Tabberabberan collision in East Australia, the Late Ordovician/Early Silurian Taconian-Caledonian collision (closed Paleo-Iapetus Ocean) followed by the Early Carboniferous Late Acadian-Caledonian collision (closed Neo-Iapetus Ocean) in Eastern North America and Western Europe and others.

While the sum of the orogenic states of an orogeny (pre-collisional, one-sided Cordilleran/Andean type to the East of the Pacific; pre-collisional, one-sided island arc type of the West Pacific; syn-collisional, two-sided collision/Himalayas type or paroxysm of orogeny between Laurasia and Gondwana at the equator to the West of the Pacific in the Pangaea area; post-collisional, extensional Mediterranean/Caribbean type further to the West, and to the East of the Pacific) will last one East drift lapping of the Earth´s crust by the Pacific basin or one Orogenic Cycle (one West drift lapping of the Pacific basin by the Earth´s crust), a certain orogenic state (e. g. Cordilleras/Andean type to the East of the Pacific) will be repeated after the collisional mountain belt or the Pacific basin have lapped the Earth once.

Lateral growth of continental crust is effected by winding the collisional mountain belt in form of a two-lap spiral around a cratonic nucleus (e. g. Laurasia). It is only in Asia, between the West Pacific and the Iranides/Southern Urals, that segments from the completed older lap and the still incomplete younger lap of the Late Paleozoic-Mesozoic-Cenozoic North Pangaea collisional mountain belt spiral are in contact with one another up to the present (Fig. 3). The age difference between the Indosinian collision and the Himalayas collision in Central and South Asia (closed Paleo-Tethys and closed Neo-Tethys) is 200 to 250 m. y.

3. *The phenomenon of the "exotic, suspect, displaced, accreted, allochthonous or tectonostratigraphic terranes"* (Coney et al., 1980; Churkin & Trexler, 1981; Fujita & Newberry, 1983; Jones et al., 1983; Natalin & Parfenov, 1983; Nur & Ben-Avraham, 1983; Saleeby, 1983; Bruns et al., 1984; Howell et al., 1985; Keppie, 1985; Parfenov & Natalin, 1986; Hallam, 1986a; Trurnit, 1986c, d; 1988a, d; 1989a, 1991d; Hutton, 1987) that were first recognized in the Cordilleras of Western North America and later on, also in the remainder of the Circum-Pacific Region, with the exception of the western margin of South America (Figs. 3, 13, 15, 16). The terranes along the North Pacific rim to a large extent arrived at their present-day position from more southerly latitudes. With the extended plate tectonic model presented here, namely the eastward displacement of the Pacific basin in relation to the Earth´s crust, the existence of tectonostratigraphic terranes is easily explained: continents of the southern hemisphere which had previously collided sequentially at the equator to the West of the eastward migrating Pacific basin against the continents of the northern hemisphere, should after half an East drift lapping of the Earth´s crust by the Pacific basin have disengaged again from the northern continents in the tensional Mediterranean and Caribbean settings to the East of the Pacific (tongue of the Pacific), leaving their former northern margins in the form of tectonostratigraphic terranes attached to the northern continents (lateral continental growth). These later migrate westwards around the Pacific basin in the North (Figs. 3, 6, 7, 13, 16), accompanied by post-collisional basin-and-range type extensional provinces (tectonic denudation/erosion or extensional unroofing/exhumation and isostatic rise of metamorphic core complexes; Wernicke, 1981; Haller, 1982; Froidevaux, 1986; Trurnit, 1986c, 1987a-c, 1988a-d, 1989a-d; Wust, 1986; Heller et al., 1987; Jayko et al., 1987; Lister et al., 1987; Davis, 1988; Laubscher, 1988).

4. *The contrast between the gentle inclinations of the subduction zones at the Andean type continental margins to the East of the Pacific and the steep inclinations of the subduction zones at the island arc type continental margins of the West Pacific* (Isacks & Molnar, 1971; Nelson & Temple, 1972; Mitchell, 1973; Moore, 1973; Uyeda & Kanamori, 1979; Mitchell & Garson, 1981; Uyeda, 1981, 1983, 1984, 1987; Trurnit, 1984c; 1988a, d; 1989a, 1991d). As it descends into the interior of the Earth along the circum-Pacific ring of subduc-

tion zones the oceanic crust of the Pacific, assuming that it belongs to the convecting upper mantle, should be attached to the lower mantle running ahead of the upper mantle-crust system with East drift. Assuming that the crust of the Pacific belongs to the convecting lower mantle cooling directly and independently from the upper mantle at the Earth´s surface, there should be an eastward or southeastward directed, jetstream-like counterflow above the mantle-core boundary (Pacific hot spot tracks; e. g. Hawaiian islands-Emperor seamounts chain) so that the lower mantle as a whole (and with it the Pacific basin) drifts East relative to the Earth´s upper mantle (in that case existing in the Pangaea area only) and crust. In both cases the downward dipping slabs of the oceanic crust in the subduction zones should be pulled towards to East. During backarc basin opening, an island arc migrates oceanwards (Pacific-, Atlantic-wards) towards the East (trench suction; Oxburgh, 1974) (Fig. 4).

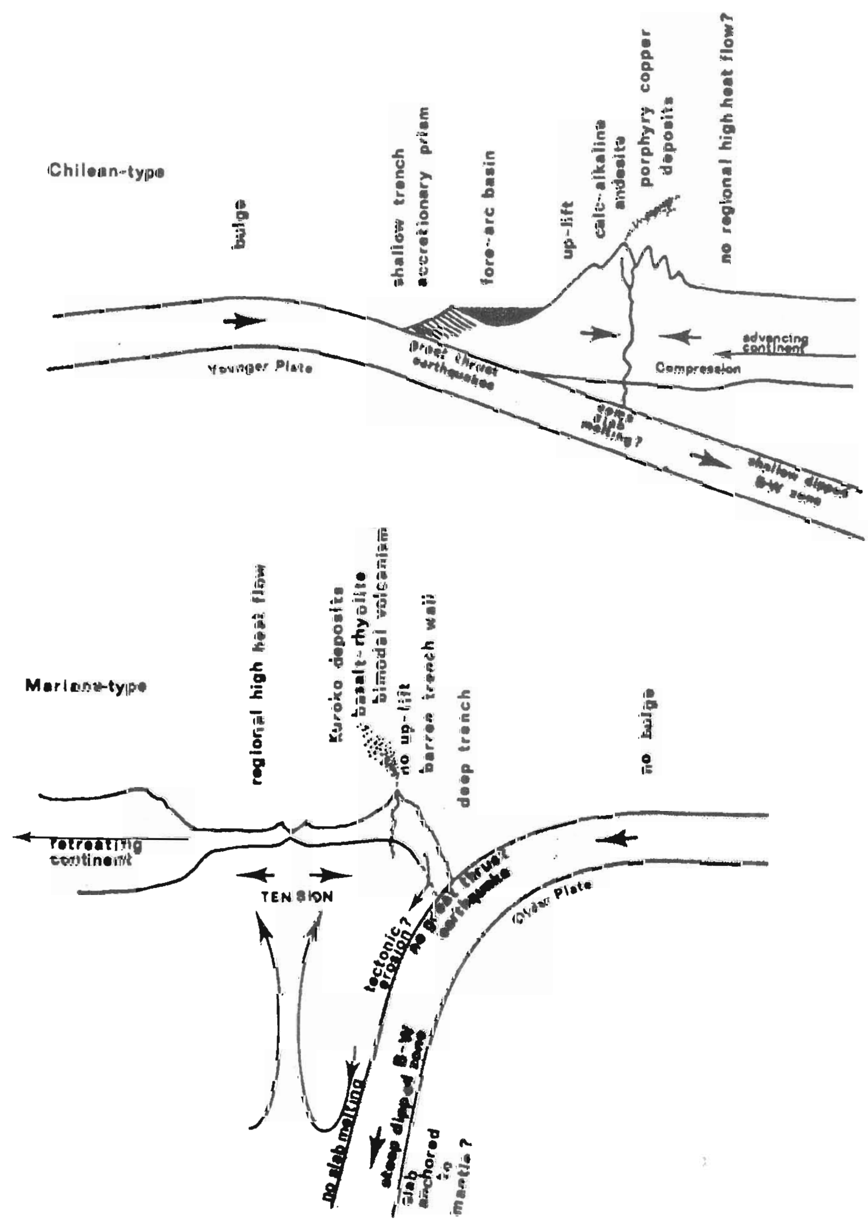


Fig. 4. Inclination of subduction zones (Uyeda, 1981, Fig. 8). a - East of the Pacific: Chilean type or Andean type margins; b - In the West Pacific: Mariana type or Island arc type margins.

5. *The arrangement of ocean types from East to West (younger to older) through 360° around the globe in the evolutionary sequence of the Oceanic, Eugeosynclinal or Wilson Cycle* (Wilson, 1966, 1968; Dewey & Burke, 1974) (Figs. 2, 3c, 9, 13, 15, 16) describing the opening and closing of an ocean with subsequent collision of the continental margins and collisional mountain building (some 180° Rift/Red Sea state and Atlantic state with mainly passive continental margins; some 180° of Pacific state with mainly active continental margins; terminated with the Collision/Himalayas state). Analogous to the ages of collisions at the end of the Wilson Cycle in the Himalayan-Alpine-Laramide-Kimmerian-Indosinian-Uralian collisional mountain belt increasing towards the West, the individual ocean states of the Wilson Cycle should also be arranged from the younger to the older evolutionary states from East to West through 360° around the globe. Provided that the Pacific basin (Pacific state, closing phase or second half of the Wilson Cycle) migrates eastwards ahead of the collision zone at the equator to the West of the Pacific (eastwards growth of the collisional mountain belt), the remaining Atlantic and Rift/Red Sea states representing the first half or opening phase of the Wilson Cycle in the Pangaea area further towards the East, should also migrate towards the East ahead of the Pacific basin and the collisional zone.

The present understanding of the Wilson Cycle is one of ocean opening with the aid of seafloor spreading (e. g. the Atlantic; Wilson, 1966), and of ocean closing with the aid of obduction/subduction (e. g. the Pacific) on approximately the same longitudes and/or latitudes with always the same continental margins bordering the same expanding and later contracting ocean (e. g. the Proto-Atlantic or Iapetus Ocean by the closure of which the Appalachians/Caledonides were formed). The extended plate tectonic model presented here is unorthodox in the sense that the ocean basins, the continental crust and the continental margins each pass through different sequences of global tectonic settings during the course of their evolution (during ocean opening and closing), i. e. they are each subjected to different global tectonic megacycles. As a result, oceanic crust (ophiolites) and miogeosynclinal continental margin sequences from collided continents, West Pacific type microcontinents and island arcs found in collisional mountain belts will in most cases not have a common history. At the end of the Oceanic, Eugeosynclinal or Wilson Cycle continental margins separated in a rift state in the Pangaea area to the East of the Pacific will in the majority of cases not meet again during the collision state in the collisional mountain belt at the equator to the West of the Pacific. The Wilson Cycle describes only the history of an ocean basin but not the evolution of the oceanic crust from that ocean nor the evolution of the continental margins bordering that ocean.

During the course of an Oceanic, Eugeosynclinal or Wilson Cycle, rift oceans of the SW-Pacific backarc basin type and of the Red Sea/Indian Ocean type will be displaced West relative to the Pacific or Tethys basin to become a North and a South Atlantic (i. e. because of the eastward displacement of the Pacific basin in relation to the Earth´s crust they will occupy a progressively more westerly setting in the Pangaea area during the course of their opening and evolution) and will finally be absorbed by the Pacific basin advancing from the West. The continental or Pangaea area continuously opens at the Pacific bow in the East in the extensional Mediterranean and Caribbean settings (rift propagation towards the East) (Channell, 1986; Cello, 1987; Lister et al., 1987; Laubscher, 1988) and continuously closes again at the Pacific stern in the West (sequential collision) between the continents in the Asia and Australia settings (zip fastener principle) (Figs. 3, 13, 15, 16). For an individual continent or continental margin the Pacific appears or "rises" in the West and disappears or "sets" in the East. During the course of its eastward displacement the Pacific retains approximately the same size; it does not shrink. What is lost at the stern from the Pacific/Tethys basin in the West through sequential collisions between the island arcs, microcontinents and continents from the northern and southern hemispheres in the Asia, Australia and equatorial West Pacific backarc basin settings (N-S compression; growth of the collisional mountain belt towards the East), is added at the bow in the East by the sequential openings of a series of Caribbeans and Mediterraneans (N-S extension; rift propagation towards the East), withdrawal of the continents in the North and South America, Europe and Africa settings towards the Poles and by the incorporation of a succession of Atlantics/rift oceans formed beforehand in the Pangaea area in the East into the Pacific (disappearance of Pacific/Tethys in the West and creation of new Pacific in the East) (Figs. 3, 13, 15, 16)(Passeríní et al., 1991). Thus, a Mid-Atlantic Ridge will slowly be replaced by an East Pacific Rise, an Azores-Gibraltar Fracture Zone by a Galapagos Rift Zone; a Mediterranean will slowly develop into a Caribbean and a Black Sea into a Gulf of Mexico. *Arabia (the Greenland of Africa?)* should continue to force its way North through the Alpine belt and the Russian Platform of Eurasia. It should in future arrive at a plate tectonic setting comparable to that presently occupied by *Greenland (the Arabia of South America?).* After having disengaged from Asia in the near geological future, Europe will rotate clockwise as North America has done during post-Laramide times since the Eocene/Oligocene (Figs. 13, 14, 16). A West Indian Ocean will become a South Atlantic, a Red Sea-Gulf of Aqaba-Levant fault a Central North Atlantic-Baffin Bay and a Persian Gulf-Mesopotamian Plain a North Atlantic. Plate tectonic characteristics like the ocean states of the Wilson Cycle move eastwards across the Earth´s crust with the angular velocity of the Pacific basin and of the lower mantle. The identity of an ocean will slowly be transferred eastwards to the neighbouring ocean: West Pacific to the East Pacific, East Pacific to the Atlantic, Atlantic to the Red Sea/Western Indian Ocean, Western to the Eastern Indian Ocean and the Eastern Indian Ocean to the SW- Pacific backarc basins. Relative to a distinct plate tectonic setting of an ocean, a continent or continental margin, a future or later evolutionary state at the Earth´s surface is always depicted in a setting simultaneously developed further to the West and a past or earlier state in a setting simultaneously occurring further to the East. Based on these considerations the sequence and relative age of the margins East and West of the Pacific can be established and the information contained in the Wilson Cycle be completed: Cordilleran and Andean type margins bordering the East Pacific should represent the forerunners of West Pacific island arc, microcontinent and back-arc basin type margins and both types in turn should be the embryonic pre-collisional states of the collision type margins at the equator to the West of the Pacific. That the situation is somewhat more complicated for margins of the types rimming the present-day South Pacific (Figs. 3, 15, 16), is demonstrated below.

According to the global stress field induced by the centrifugal force of the spinning Earth (Moody & Hill, 1956; Knetsch, 1965; Moody, 1966; Roland, 1976; Kafka & Kvet, 1988; Zoback et al., 1989; Turnit, 1991d), vertical tectonic planes orientated parallel to longitude are subjected to tension in a W-E direction and to shearing in a diagonal orientation. Arranged in an orientation parallel to latitude/equator, they are subjected to compression in a N-S direction. The forces are strongest close to the equator and become weaker towards the Poles.

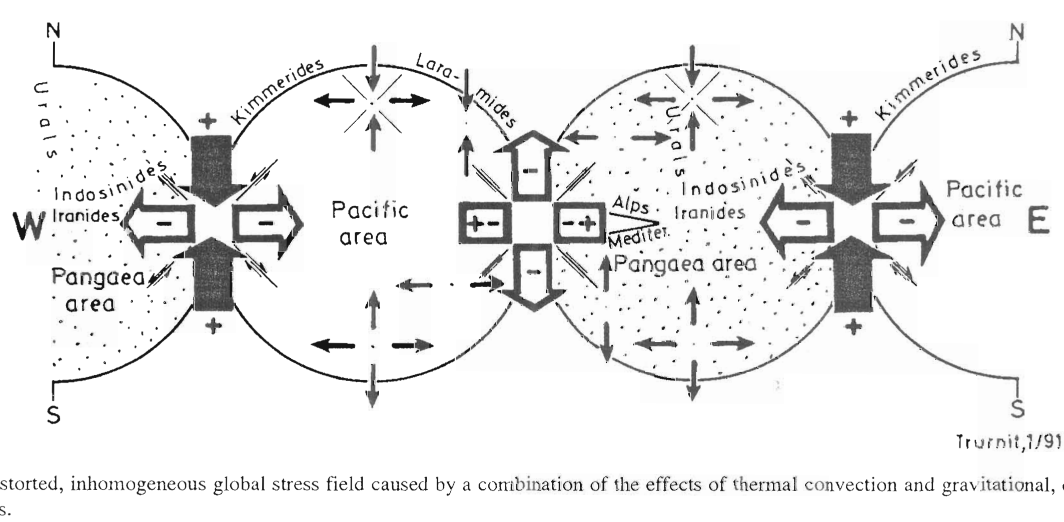


Fig. 5. The distorted, inhomogeneous global stress field caused by a combination of the effects of thermal convection and gravitational, centrifugal and tidal forces.

Plate margins and other plate tectonic planes (e.g. transform faults, fracture zones, etc.) mostly behave in accordance with the centrifugally induced global stress field. The model of the Wilson Cycle combined with the global stress field, should make oceans open with the aid of seafloor spreading in an orientation parallel to longitude and oceans close and their margins collide with the aid of subduction/obduction in an orientation parallel to latitude/equator. The backarc basins of the West Pacific, the Red Sea/Indian Ocean, the Atlantic and the East Pacific Rise are indeed stretched roughly in a N-S direction. The Southern or Circum-Antarctic Ocean with its zones of seafloor spreading forms an exception because of the weakness of the rotating Earth´s centrifugal force in the Polar Regions. The collision zone, along which the Neo-Tethys Ocean to the West of the present-day Pacific Ocean was closed during the Late Cretaceous and Tertiary to become the Alpine collisional mountain belt, is orientated roughly parallel to latitude close to the equator (Figs. 3, 6, 7, 13, 15, 16).

The centrifugally induced global stress field is distorted by gravity and geoid anomalies most probably caused by the convecting upper mantle/crust system, the convesting lower mantle, and the same forces that displace the Pacific basin East relative to the Earth´s crust. As a result, the effects of the centrifugally induced stress field in the West Pacific and to the West of the Pacific are intensified (formation of the collisional mountain belt parallel to latitude/equator and opening of back-arc basins parallel to longitude) and to the East of the Pacific they are weakened, neutralized or slightly reversed (opening of Mediterraneans/Carribeans parallel to latitude/equator and formation of Cordilleran/Andean type mountain belts parallel to longitude) (Isacks & Molnar, 1971; Oxburgh, 1974; Trurnit, 1984c, 1988a, d; 1989a; 1991a, d) (Fig. 5). The pattern of gravity anomalies and the undulations of the geoid (Mörner, 1984; Aubrey et al., 1988; Pirazzoli, 1988; Trurnit, 1991d) should advance across the Earth´s surface with the angular velocity of the lower mantle, the Pacific basin, the ocean states of the Wilson Cycle and the sequences of states of other global tectonic megacycles.

6. *The switch from Late Cretaceous/Early Tertiary compression in N-S direction to Late Oligocene/Miocene/Pliocene extension in the same direction in the stress field of the Alpine-Mediterranean-North African Region* (Figs. 5-7). This reversal of stress could have been effected through rotations of the respective continents or areas through the centrifugally induced global stress field, the orientation of which is aligned with the spinning Earth´s axis and equator. However, major rotations

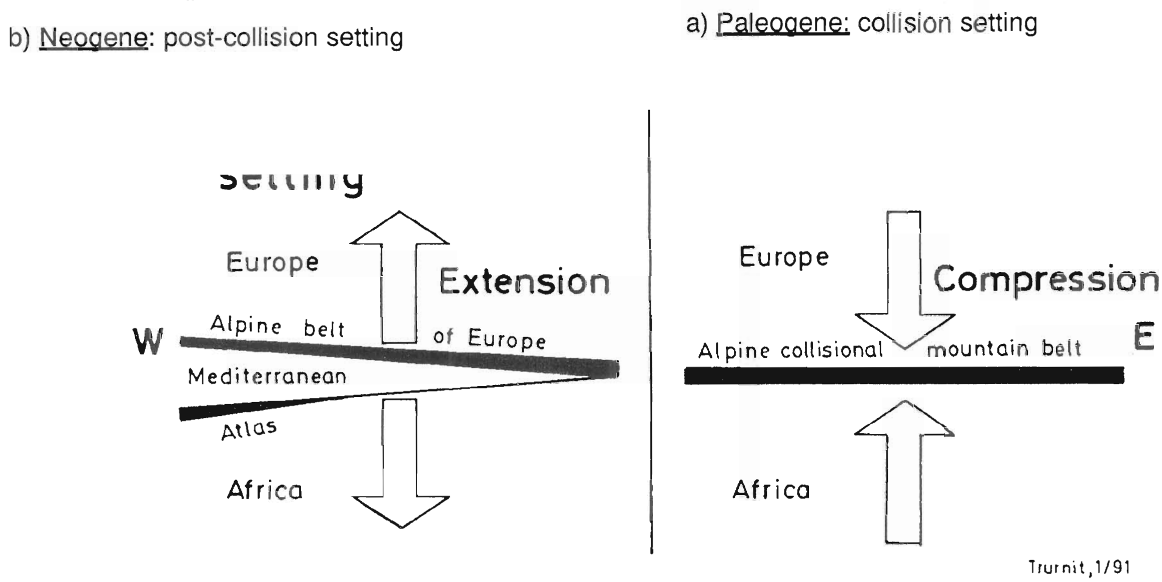


Fig. 6. N-S orientated stress in Europe during. a - the Early and b - the Late Tertiary.

of Africa and Europe during the Tertiary have not been recorded paleomagnetically. A second and more probable explanation would be a westward directed shift of the Alpine/Mediterranean area and its accompanying continents in the North and South, away from the collision setting at the equator to the West of the Pacific and in the eastern part of the Pangaea area, by means of an eastward displacement of the Pacific basin relative to the Earth's crust and continents, towards a more central position and finally western setting in the Pangaea area and to the East of the Pacific, where in the distorted, inhomogeneous global stress field of Figure 5 N-S directed extension prevails. As a result, the Alpine belt was stretched in N-S direction along low-angle detachment faults (tectonic denudation/erosion, extensional unroofing/exhumation and isostatic rise of metamorphic core complexes;

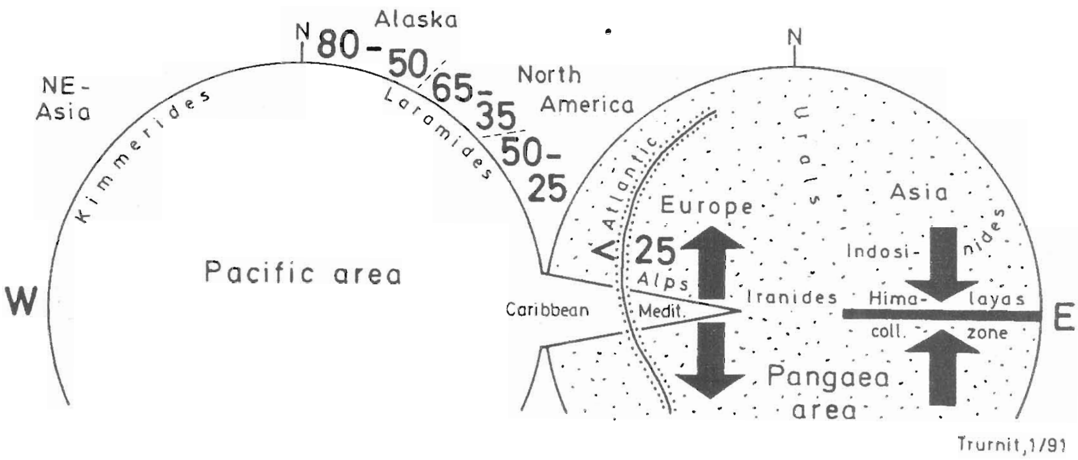


Fig. 7. Beginning and termination of post-collisional, N-S directed Mediterranean/Caribbean type extension in million years as well as eastwards propagation of basin-and-range type extensional tectonics (isostatic rise of metamorphic core complexes along low-angle detachment faults).

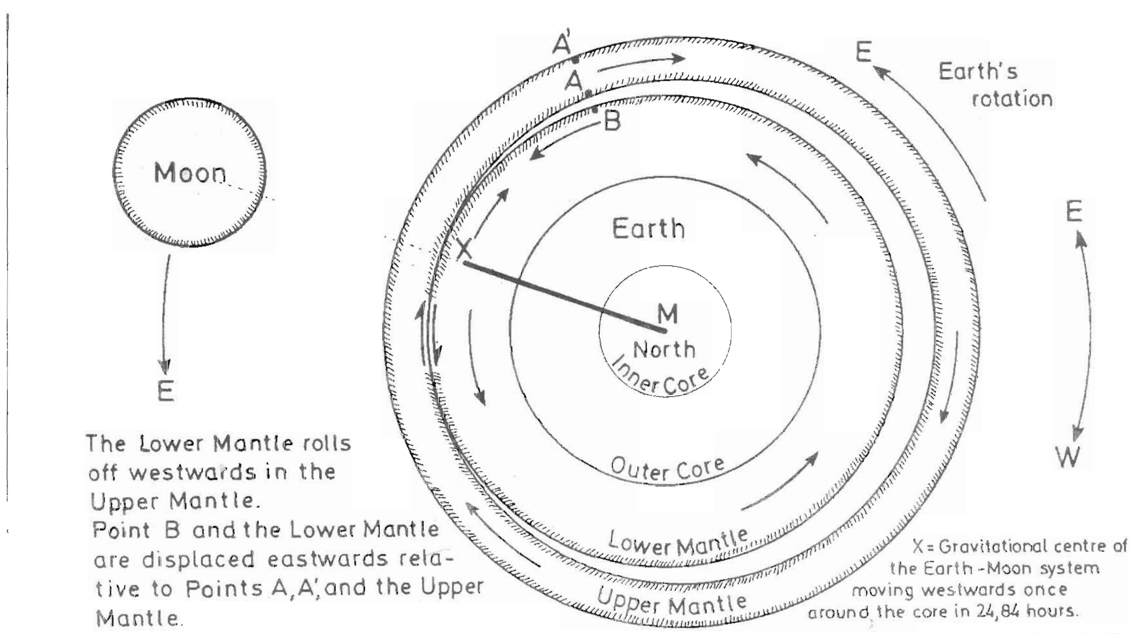


Fig. 8. The Earth depicted in the form of a gigantic hypocycloid gearing; explanation for the eastward displacement of the Pacific basin relative to the Earth's crust (North polar view). Off-centre rotation of the spinning Earth around the gravitational centre of the Earth-Moon (-Sun) system, which moves West through the lower mantle around the Earth's spin axis with the angular velocity of the Earth's tidal bulges (see also Fig. 7 of Trurnit, 1988a).

Wernicke, 1981; Lister et al., 1987; Davis, 1988; Dewey, 1988; Laubscher, 1988; Trurnit, 1988a, b; 1989a, 1991d; Kilias, 1991; Seyitoglu & Scott, 1991) and the Mediterranean area subsided below sea level from Late Oligocene/Early Miocene onwards (Morelli, 1990).

Analogous to the earlier/younger to older/later arrangement of ocean types from East to West through 360° around the globe in the evolutionary sequence of the Oceanic or Wilson Cycle (opening and closing of an ocean) and the collision ages in the Himalayan-Iranian-Alpine-Laramide-Kimmerian-Indosinian-Uralian collisional mountain belt becoming older towards the West, the ages of post-collisional extension in this belt and its forelands (Mediterranean/Caribbean type rifting; basin-and-range type extension; rise of metamorphic core complexes, etc.) should also show an increase towards the West (Fig. 7) (Eisbacher, 1988).

The contribution of tidal forces to global tectonics and plate movements

The eastward displacement of the Pacific basin relative to the Earth's crust and the high probability of its coupling to the angular velocity of the lower mantle and its gravity anomalies, demands a slightly higher angular velocity of the convecting lower mantle around the Earth's spin axis as compared with that of the convecting upper mantle-crust system (Nadai, 1952; Jardetzky, 1954; Bostrom, 1971a, b; 1972, 1976, 1977; Tanner, 1971, 1973; Roeder & Nelson, 1971; Knopoff & Leeds, 1972; Meyerhoff & Meyerhoff, 1972; Nelson & Temple, 1972; Cullen, 1973; Danes, 1973; Gilliland, 1973; Moore, 1973, 1975; Jordan, 1974; Lyalin, 1979; Uyeda, 1983; Miyashiro, 1986; Doglioni, 1991). This presumed behaviour of the Earth appears to be caused by the off-centre rotation of the spinning Earth around the gravitational centre of the Earth-Moon (-Sun) system, which moves West through the lower mantle around the Earth's spin axis with the angular velocity of the Earth's tidal bulges. According to the principle of hypocycloid gearing, the outer circumferences of solid inner shells roll off in the inner circumferences of solid outher shells (kinematics of the hypocycloidal movement), whereby the direction of displacement of the inner shells in relation to the outer shells is opposite to the off-centre rotation and the amount of displacement between neighbouring shells depends on the difference between the radii or circumferences (Fig. 8) (Lehmann, 1979 and personal communication 1984; Trurnit, 1984c, d; 1985a - c; 1986a - d; 1987a - c; 1988a - d, 1989a - e, 1990a - c, 1991a - d).

Besides the possibility of a two-layer, two-shell or superposed mantle model with relatively independent upper mantle-crust and lower mantle convection systems or a model of whole mantle or mantle-wide convection, there are also indications for a combination of whole mantle convection with two-layer convection. In such a configuration, the Pacific crust might belong to the roof of the lower mantle that cools independently at the Earth's surface. In such case, only the oceanic crust from the Intra-Pangaea Oceans (Atlantic, Indic, Arctic, Circum-Antarctic Oceans) in the North and South Pangaea areas might belong to the upper mantle-crust convective system. The difficulty with such a model is the relative movement of the Pacific crust relative to the remainder of the Earth's crust (Pangaea area with the crust of the Intra-Pangaea Oceans). The Pacific crust is disconnected to a large extent from the surrounding continental and oceanic crust of the North and South Pangaea areas along the circum-Pacific ring of subduction zones (with the exception of West Antarctica/Antarctica that rotates clockwise in the sense of the East drift of the lower mantle) and can move relatively independently. Taken as a whole, the of the Pacific Plate crust does not appear to drift East with the Pacific basin in relation to the Earth's crust but (besides rotating counterclock-wise since the Permian/Early Mesozoic) in relation to the Pacific basin moves West or NW in more or less the same direction as the remainder of the Earth's crust. An assumed eastward or southeastward directed jet-stream-like counterflow in the whole-mantle convection cell of the Pacific above the mantle-core boundary (Hawaiian Islands - Emperor Seamounts Chain and other similar Pacific hot spot tracks) would overcome this difficulty and would move the mantle material of the Pacific whole-mantle convection cell as a whole eastwards with the same angular velocity as the Pacific basin (Trurnit, 1986a, 1988a, 1989a, 1991d).

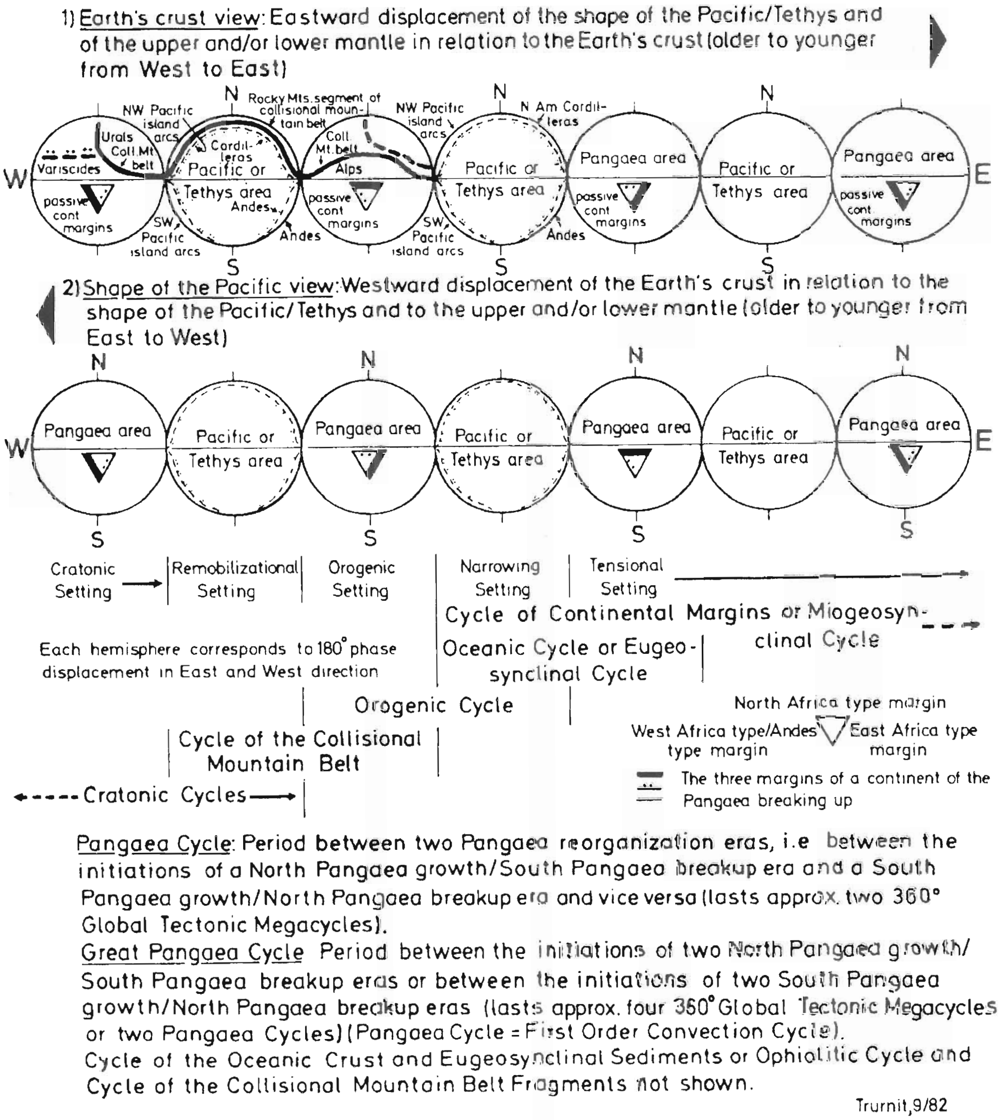


Fig. 9. Evolutionary sequence of the global tectonic megacycles migrating East together with the Pacific basin and the lower mantle relative to the Earth´s crust (upper mantle-crust system) (equatorial view); some 120° clockwise rotation of a continent from the South Pangaea/Gondwana breaking up in the southern hemisphere while passsing through the Antarctica setting during one East drift lapping of the Earth´s crust by the Pacific basin (one West drift lapping of the Pacific basin by the Earth´s crust).

The global tectonic megacycles of the oceans (the continents), the continental margins, the oceanic crust and orogenic mountain belts

The Earth´s history is subdivided by, and the Earth´s crust is subjected to, a series of 360° global tectonic megacycles (Gastil, 1960; Sutton, 1963; Williams, 1981; Stockwell, 1982; Salop, 1983; Trurnit, 1984a-d; 1985a-d; 1986a-d; 1987a-d; 1988a-d; 1989a-e; 1990a-c; 1991a-d; Worsley et al., 1984, 1986; Schmidt & Baumann, 1985; Nance et al., 1986, 1988) arranged in series from East to West from the younger or earlier to the older or later evolutionary states, overlapping each other by 180° each (by definition) and advancing East across the Earth´s crust above gravity anomalies of the lower mantle and with the angular velocities of the Pacific basin and the lower mantle that rotate somewhat faster around the Earth´s spin axis relative to the convecting upper mantle-crust system (Fig. 9): *The Oceanic, Eugeosynclinal or Wilson*

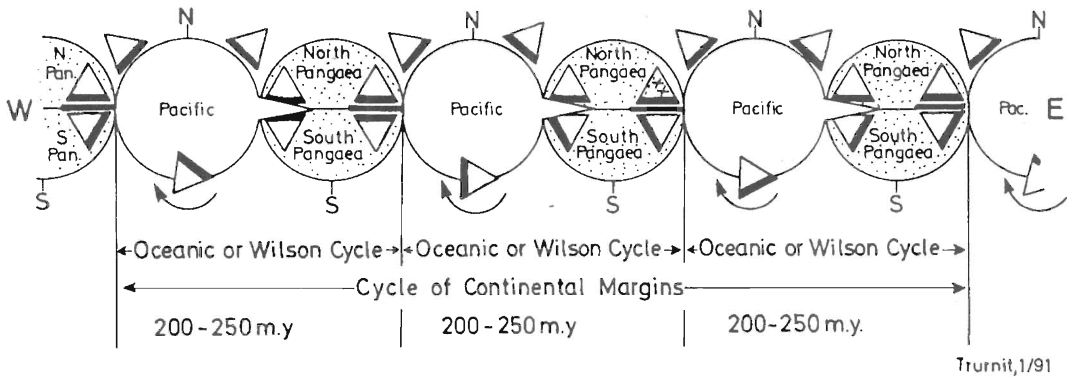


Fig. 10. Distinction between the Oceanic, Eugeosynclinal or Wilson Cycle and the Cycle of Continental Margins or Miogeosynclinal Cycle. On average, the Oceanic or Wilson Cycle lasts one, the Cycle of Continental Margins three 360° East drift lappings of the continent, that the margin is part of, by the Pacific basin, between the rift states in the Pangaea area (parallel to longitude SW-Pacific backarc basins, Red Sea and Atlantic type rift states; parallel to latitude Mediterranean and Caribbean type rift states) and the collision state at the equator to the West of the Pacific.

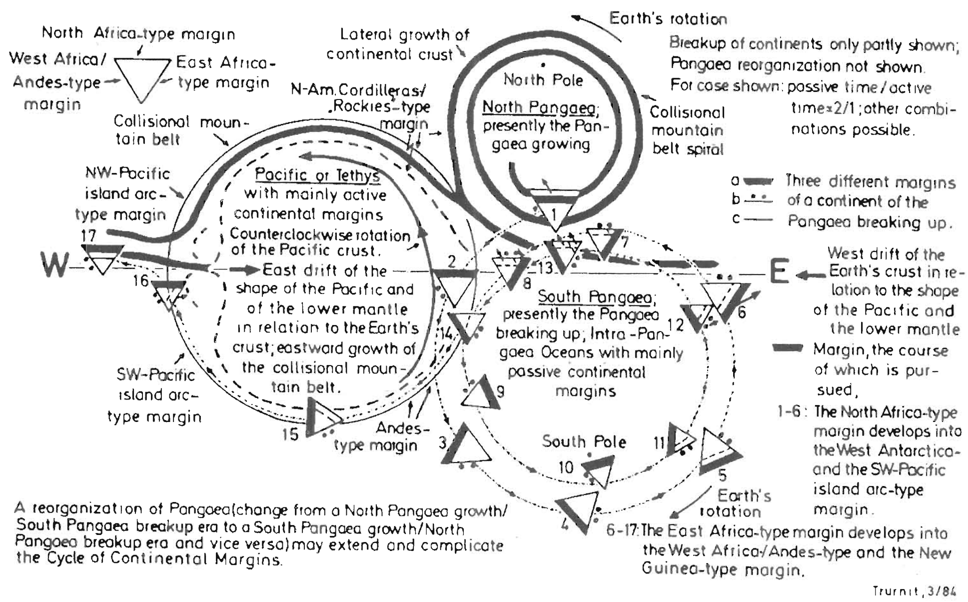


Fig. 11. Triad of the Pacific area, North Pangaea area and South Pangaea area. Sequence of plate tectonic settings a continent from the Pangaea breaking up is passing through on its way westwards around the globe during the Cycle of Continental Margins or Miogeosynclinal Cycle (Nos. 1-16) (combined equatorial, North Pangaea and South Pangaea hemisphere views).

Cycle (some 180° Red Sea/Rift state and Atlantic state with mainly passive continental margins in the Pangaea area; some 180° Pacific state with 90° Cordilleran and Andean type continental margins East of the Pacific, and 90° island arc, microcontinent and backarc basin type continental margins in the West Pacific; terminated with the Collisional/Himalayas state at the equator to the West of the Pacific) (Figs. 2, 3, 9), the *Orogenic Cycle* (some 180° Pacific state; some 180° collisional mountain belt state in the Pangaea area), and the *Cycle of the Collisional Mountain Belt* (180° collisional mountain belt in the Pangaea area; 180° collisional mountain belt either in the northern or the southern hemisphere along the Pacific rim) (Figs. 3, 9). This series is headed in the East by the *Cycle of Continental Margins or Miogeosynclinal Cycle* (Figs. 9 - 11) and is succeeded in the West by the *Cratonic Cycles* (Fig. 9), both types with a length and duration of more than 360°.

One East drift lapping of the Earth´s crust, of the Pangaea areas with intra-Pangaea Oceans or of the

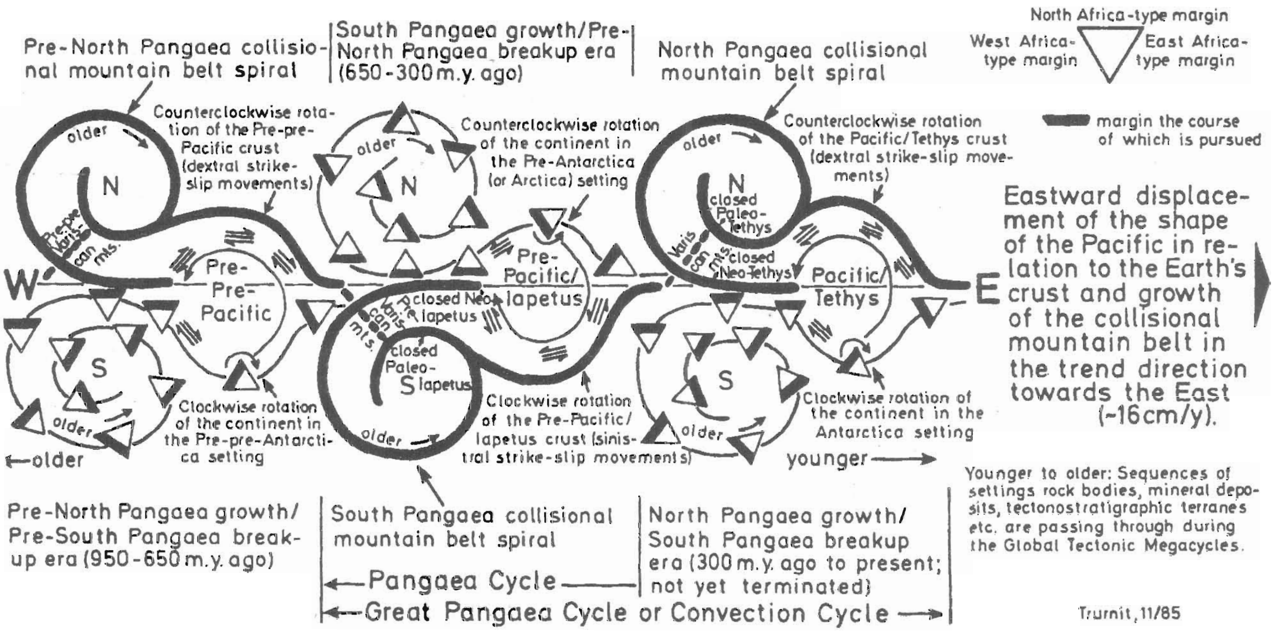


Fig. 12. The Earth´s history appears to be subdivided into alternating North Pangaea growth/South Pangaea breakup eras and South Pangaea growth/North Pangaea breakup eras. In the hemisphere of the Pangaea growing a collisional mountain belt continuously forming behind the eastward migrating Pacific basin winds around a cratonic nucleus in the form of a two-lap spiral. In the hemisphere of the Pangaea breaking up a continent in the Antarctica setting rotates through approximately 120° (clockwise during a South Pangaea breakup era; counterclockwise during a North Pangaea breakup era) and breaks up into several fragments during one East drift lapping of the Earth´s crust by the Pacific basin. Two East drift lappings of the Earth´s crust by the Pacific basins, by the ocean states of the Wilson Cycle, two laps of the collisional mountain belt or two succeeding, not overlapping 360° Global Tectonic Megacycles correspond to one Pangaea Cycle; two Pangaea Cycles make one Great Pangaea or Convection Cycle. Between Pangaea Cycles the collisional mountain belt is interrupted by Europe type continents and by Variscan type mountain belts. The latter form in West Pacific island arc type settings (Fig. 19: Marathon-Ouachita-Alleghenian and European Variscan belt) (combined equatorial, North Pangaea/North Polar and South Pangaea/South Polar views).

continents by the Pacific basin, by the sequence of ocean states of the Wilson Cycle, by the collisional mountain belt or by the lower mantle during the younger history of the Earth lasted approximately 200 to 250 m. y., taken from the differences in age between the overlapping segments of the first completed and older lap (Triassic Indosinian segment of the closed Paleo-Tethys) and the second, still incomplete and younger lap (Tertiary Alpine-Himalayan segment of the closed Neo-Tethys) of the Late Paleozoic-Mesozoic-Cenozoic North Pangaea collisional mountain belt spiral in Asia (Figs. 3, 9) (calculated with 250 m. y.: some 16 cm/y or 0,44 mm/d East drift at the equator and growth of the collisional mountain belt behind the Pacific basin towards the East).

For determining the setting of an individual rock body, ore body, tectonostratigraphic terrane or continent during a specific era of the Past, one has to shift it back East through the sequence of plate tectonic settings it has been subjected to or passed through since that era. For Europe this succession of settings would be: South Asia setting - Tertiary; NE-Asia/NW-Pacific island arc, microcontinent and backarc basin setting - Cretaceous; North America/Cordilleras-Rocky Mountain setting - Jurassic; Europe/Northern Mediterranean setting - Triassic; South Asia setting - Permian, etc.) (Figs. 3c, 9).

Pangaea (all continental crust, past and present) consist of a North Pangaea and a South Pangaea (a North and a South Pangaea collisional mountain belt spiral) roughly distributed in the northern and southern hemispheres, whereby simultaneously and alternately one is always growing while the other is breaking up. Since the Permian a North Pangaea (Laurasia) has been growing and feeding on the last South Pangaea (Gondwana and North America) breaking up simultaneously. During the Late Proterozoic and Early to Middle Paleozoic (approximately from 650 my. to 320 m. y. ago), a South Pangaea has been growing while a Pre-North

Pangaea was breaking up simultaneously (Fig. 12). Lateral growth of a Pangaea and of its continents as well as the cyclical repetition of orogenic states for a certain longitude or continental margin are effected by the joining of the younger lap of a collisional mountain belt spiral to the older, from the equator towards a pole. The North Pangaea area and the South Pangaea area are comparable to a pair of cogs rotating against one another (Figs. 11, 12). However, relative to the Earth's crust, the continents or plates and their margins remain more or less fixed to their longitudes on which they only move polewards and towards the equator (pendular movements), while the Pacific basin moves eastwards between them. Antarctica has moved southwards from an equatorial setting during the Cretaceous/Tertiary along the longitudes of South America only on a grid related to the eastward migrating Pacific basin. On a grid related to the Earth's crust, it has come from a position on the equator to the West of the present-day East Pacific Rise. Periodically recurring growths and disintegrations of Pangaeas have also been proposed by Worsley et al. (1984, 1986) and Nance et al. (1986, 1988).

Approximately two neighbouring (not overlapping) cycles of the above-mentioned 360° global tectonic megacycles, two East drift lappings of the Earth's crust by the Pacific basin (two West drift lappings of the Pacific basin by the Earth's crust) or the growth of some two laps of the collisional mountain belt (a two lap spiral) correspond to one *Pangaea Cycle* between two Pangaea reorganization eras. Two Pangaea Cycles correspond to one *Great Pangaea Cycle or Convection Cycle* (Figs. 12, 21). The Pangaea Cycle is comparable to the 11 year sunspot cycle (magnetic solar cycle), the Great Pangaea cycle to the 22 year sunspot cycle that is antisymmetrical about the Sun's equatorial plane.

Pendular movements, rotations and breakup of continents;the reassembly of Gondwana,of South Pangaea and of its Late Proterozoic to Middle Paleozoic colisional mountain belt spiral

During one East drift lapping of the Earth's crust by the Pacific basin of 200 to 250 m. y. duration the continents of the northern and southern hemispheres move differently, in that their margins pass through different sequences of plate tectonic settings:

In the *hemisphere of the Pangaea growing* (since the Permian the northern hemisphere) the continents always face either the equator or the Pacific

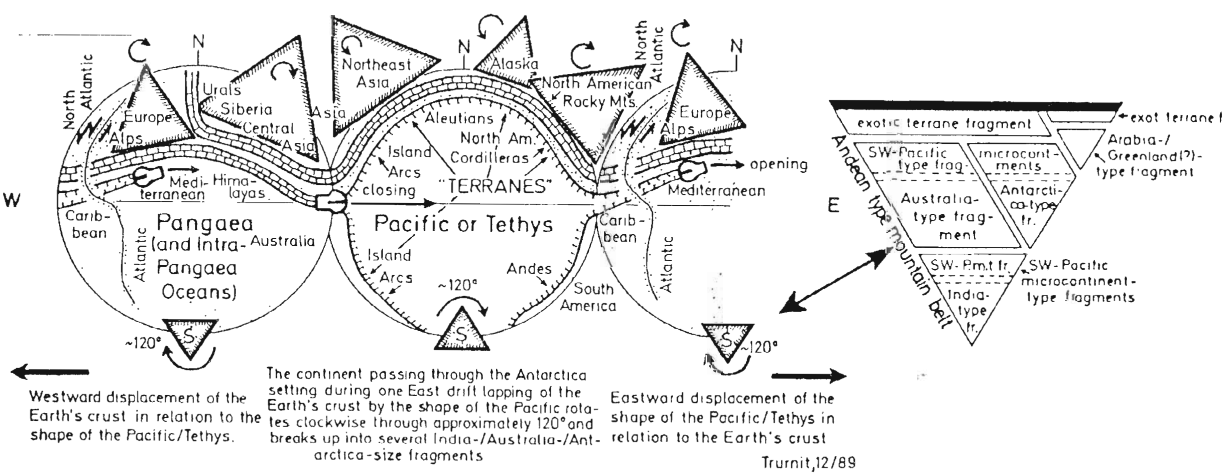


Fig. 13. Rotational and pendular movements (movements between high and low latitudes and alternating clockwise and counterclockwise rotations) of the continents from the Pangaea growing and from the Pangaea breaking up. In the northern hemisphere the continents always face the equator or the Pacific with the same margin. Otherwise, a collisional mountain belt spiral would not form. The continent in the Europe and North America setting rotates clockwise and the continent in the Alaska and NE-Asia setting rotates counterclockwise (opposite rotations during South Pangaea growth eras). The remaining two margins alternate between an Arctic and a North Atlantic type of setting for one Pangaea Cycle. In the hemisphere of the Pangaea breaking up the continents are subjected to pendular movements between high and low latitudes, rotate through approximately 120° in the southern South America and Antarctica settings (clockwise during South Pangaea breakup eras; counterclockwise during North Pangaea breakup eras) and additionally break up into India, Australia and Antarctica type fragments during one East drift lapping of the Earth's crust by the Pacific basin.

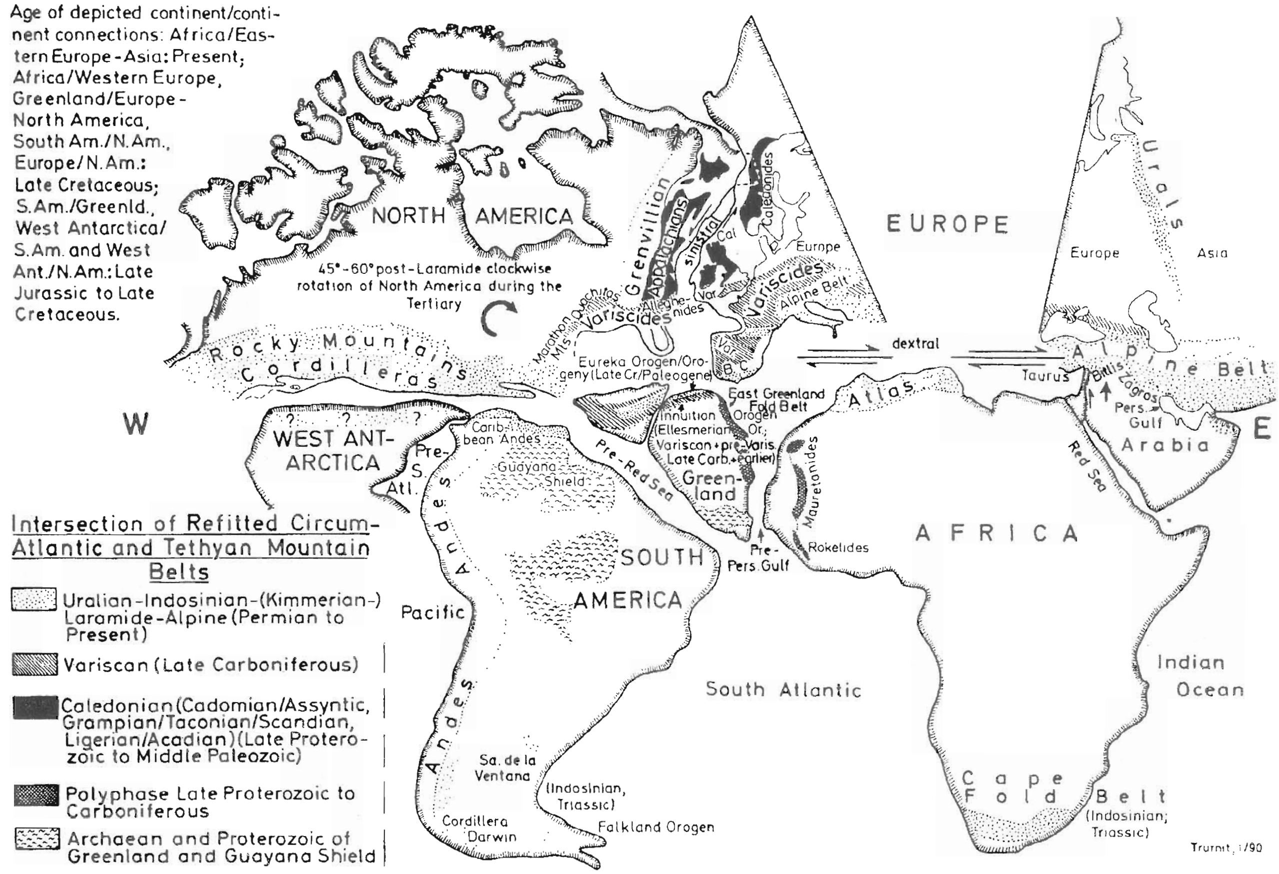


Fig. 14. Late Jurassic to Early Tertiary scenario of the Laramide-Alpine segment of the North Pangaea collisional mountain belt and of the attached continents before and during the initial opening of the Central and North Atlantic Oceans. Both Greenland (the Arabia of South America) and Arabia (the Greenland of Africa) appear to have been disconnected from their respective parent continents in similar plate tectonic settings. The collision between the continents from the northern and southern hemispheres always takes place parallel to the equator in only one setting and that is at the equator to the West of the eastward migrating Pacific basin (see also Fig. 16).

with the same margin (Fig. 13). Otherwise a collisional mountain belt spiral would not form (Fig. 3). As the continents move around the Pacific basin in the North, they are subjected to a pendular movement (alternating clockwise and counterclockwise rotations combined with movements between high and low latitudes). The continents in the Europe and North America settings rotate clockwise (Passerini et al.,1991) after the openings of a respective North Atlantic and respective Mediterraneans and Caribbeans and the continents in the settings of Alaska and Northeast Asia rotate counterclockwise (Fig. 13) (opposite rotations during South Pangaea growth eras in the southern hemisphere). Since the Permian the Cordilleran type margin of the NE-Pacific develops into a NW-Pacific island arc, microcontinent and backarc basin type and both types are forerunners of the collision type margins at the equator to the West of the Pacific. Since the Late Jurassic/Cretaceous, both types are remobilized segments from the older lap of the North Pangaea collisional mountain belt spiral (Figs. 3, 13, 16).

Due to the pendular movements of the continents from the Pangaea growing the eastern margin of North America should not be joined to the northwestern margin of Africa (West Africa connection a major error ! Currently accepted Pangaea reconstruction; Trurnit, 1990a-c; 1991b, d). The remaining two margins of a continent from the Pangaea growing stay passive and alternate between an Arctic and a North Atlantic type of setting. With regard to their opening and in contrast to the relatively simply opened South Atlantic, the Central and

North Atlantic Oceans appear to be the least understood oceans of the Earth (Figs. 13, 14).

In the literature, the Cordilleran/Andean type states East of the Pacific and the island arc type states of the West Pacific, the specific features of which are imposed upon a continental margin passing through these settings before arriving at the collision setting at the equator to the West of the Pacific, are mostly thought of as being separate collision states when their evolution from the point of view of the completed orogen is described, instead of being correctly interpreted as embryonic pre-states of collision (e. g. individual states of the Orogenic Cycle). This wrong conception inhibits understanding the true relations. The Late Jurassic one-sided Nevadan "orogeny" of Western North America should represent the embryonic or pre-collisional NE-Pacific Cordilleran and/or NW-Pacific island arc states along the northern rim of the Mesozoic Pacific or Tethys, to the East, in front and ahead of the two-sided, Late Cretaceous Western North America Laramide collision (orogeny) at the equator to the West of the Late Mesozoic Pacific or Tethys (Figs. 3, 13, 16). Likewise, the Late Silurian/Devonian Acadian "orogeny" should only represent the Cordilleran and island arc states between the Late Ordovician to Early Silurian Taconian collision and the Late Devonian/Early Carboniferous Late Acadian collision (fragments of the South Pangaea collisional mountain belt spiral containing segments from the older and the younger lap; Figs. 12, 16, 17) along the southern rim of the Paleozoic Neo-Iapetus or Pacific Ocean.

In the *hemisphere of the Pangaea breaking up* (since the Permian the southern hemisphere) the continents are also subjected to pendular movements between high and low latitudes and additionally rotate through approximately 120° while passing through the southern South America and Antarctica settings (clockwise during the breakup of a South Pangaea in the southern hemisphere - Permian to Present; counterclockwise during the breakup of a North Pangaea in the northern hemisphere - Late Proterozoic to Middle Paleozoic; a time-lag on the initiation of rotation must be taken into account - Fig. 16), during one East drift lapping of the Earth's crust by the Pacific basin (Fig. 15). The relative movements (strike-slip along transcurrent

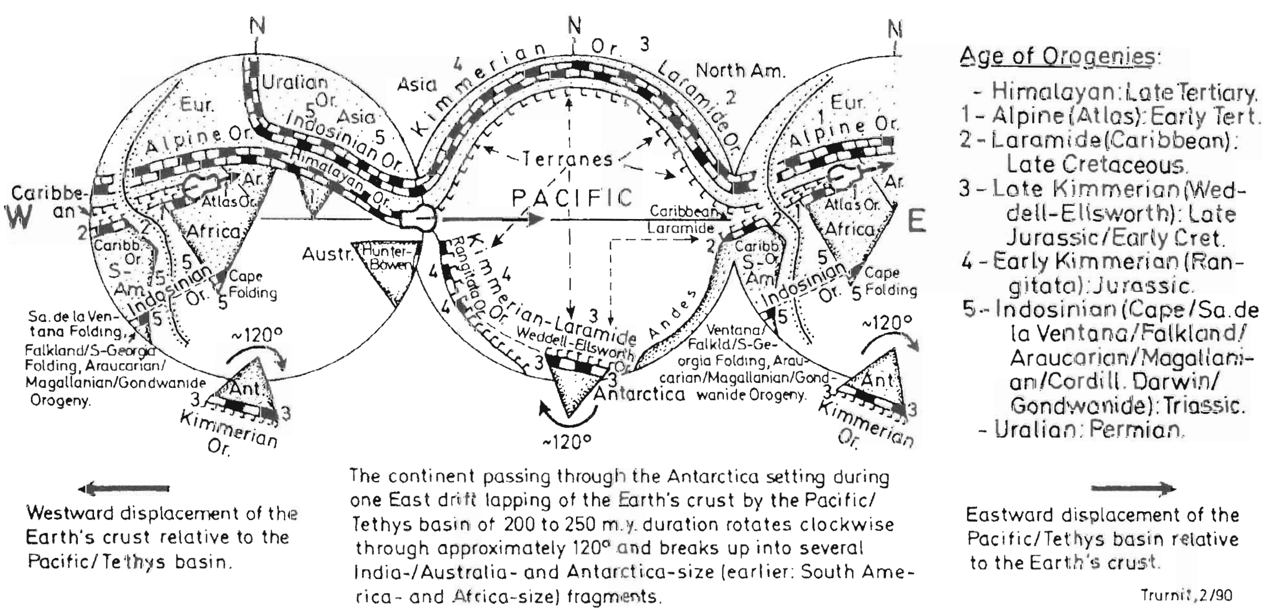


Fig. 15. Fragments of the Late Paleozoic-Mesozoic-Cenozoic North Pangaea collisional mountain belt (northern hemisphere) that circle the Pacific in the South together with the South Pangaea/Gondwana continents. Formation of tectonostratigraphic terranes: Continents of the southern hemisphere which had previously collided at the equator to the West of the eastward migrating Pacific basin against the continents from the northern hemisphere, should after half an East drift lapping of the Earth's crust/the continents by the Pacific basin have disengaged again from the northern continents in the extensional, post-collisional Mediterranean and Carribbean settings to the East of the Pacific, leaving their former northern margins in the form of tectonostratigraphic terranes attached to the northern continents. These later migrate westwards around the Pacific basin in the North, accompanied by post-collisional basin-and-range type extensional provinces (see also Fig. 16).

faults; oblique spreading ?) between the plate in the Antarctica setting and the neighbouring plates of the Earth´s crust during the presently prevailing South Pangaea breakup era are sinistral (dextral during North Pangaea breakup eras in the northern hemisphere; Knopoff and Leeds, 1972; Nelson and Temple, 1972; Dalziel, 1984; Tessensohn, 1984; Trurnit, 1988a, b; 1989a, 1991a, d). In the Antarctica setting a continent or plate should rotate in phase with the East drift angular velocity of the Pacific basin and the lower mantle relative to the remainder of the Earth´s plates.

Since the Permian, the active Andean type margin of the SE-Pacific evolves from a passive West Africa type (i.e. a passive margin that developed or collided against a subduction zone; Dietz, 1963; Dewey, 1969; Dewey & Bird, 1970; Garfunkel, 1975; Burke et al., 1984; Officer and Drake, 1985; Trurnit, 1988a, b; 1989a, 1991a, d), which subsequently develops directly (without passing through the SW-Pacific island arc, microcontinent and backarc basin setting) into an active New Guinea type margin and into the southern thrust zone of the two-sided, bilaterally verging collisional mountain belt (Collision/Himalayas type) forming at the equator to the West of the Pacific (Figs. 3, 15, 16). The South American Andes are only a one-sided or unilaterally verging pre-state or forerunner (former passive West Africa type margin) of the collision type margin.

The SW-Pacific island arc/backarc basin/East Australia and later East Africa type margins in the hemisphere of the Pangaea breaking up (since the Permian the southern hemisphere; during the Late Proterozoic and Early to Middle Paleozoic the northern hemisphere) have evolved from the passive North Africa/southern Mediterranean, passive northern South America/southern Caribbean or passive West Antarctica type margins, that still might carry fragments from the southern parts of the collisional mountain belt (Fig. 13 - 15) (e. g. Atlas, Caribbean Andes of Colombia and Venezuela; West Antarctica; New Zealand - Rangitata orogen; East Australia - New England Orogen; etc.) the main parts of which migrate around the Pacific basin in the form of tectonostratigraphic terranes in the North, together with and as parts of the collisional mountain belt of the North Pangaea growing since the Permian. Should an active Andean type margin in the hemisphere of the Pangaea breaking up be missing, then instead of an active New Guinea type margin, a passive continental margin might also be incorporated into the collisional mountain belt at the equator to the West of the Pacific (e. g. in future the passive margin of Antarctica facing Australia).

During one East drift lapping of the Earth´s crust by the Pacific basin, a West Gondwana, East Gondwana, South America or Africa type continent will break up into several fragments in the southern South America and Antarctica settings (Fig. 3c, 13): e. g. South Pangaea into Laurentia, West Gondwana and East Gondwana; East Gondwana into India, Australia and Antarctica; West Gondwana into South America and Africa; South America in future into the Fireland-Falkland, the Brazilian and the Guayana continents; Africa in future into the South Africa, East Africa and West Africa continents (Figs 13, 16). Generally accepted and visible evidence of these events and these settings only are the separations of South America from Africa since the Lower Cretaceous and of Australia from Antarctica since the Early Tertiary. Apart from these breakups, one of the three margins of a South America, an Africa, an East Gondwana, a West Gondwana or a South Pangaea type continent is scaled off twice during one East drift lapping by the Pacific basin (Figs. 15, 16): Firstly the North Africa/southern Mediterranean type margin during its separation from the collisional mountain belt in the Mediterranean setting. Major parts of this margin remain attached to the collisional mountain belt of the Pangaea growing (tectonostratigraphic terranes). Only relics of the belt stay with the newly formed passive North Africa type margin (Atlas, Caribbean Andes, West Antarctica, New Zealand and parts of East Australia; Fig. 15). Secondly the East Australia/SW-Pacific island arc and microcontinent type margin (former North Africa type). With the SW-Pacific island arcs and microcontinents (e. g. New Zealand, etc.) most relics of the old collisional mountain belt lap are finally integrated into the new or younger lap of the belt. The SW-Pacific island arcs and microcontinents from one continent are pushed North (pushed South during South Pangaea growth eras) into the collisional

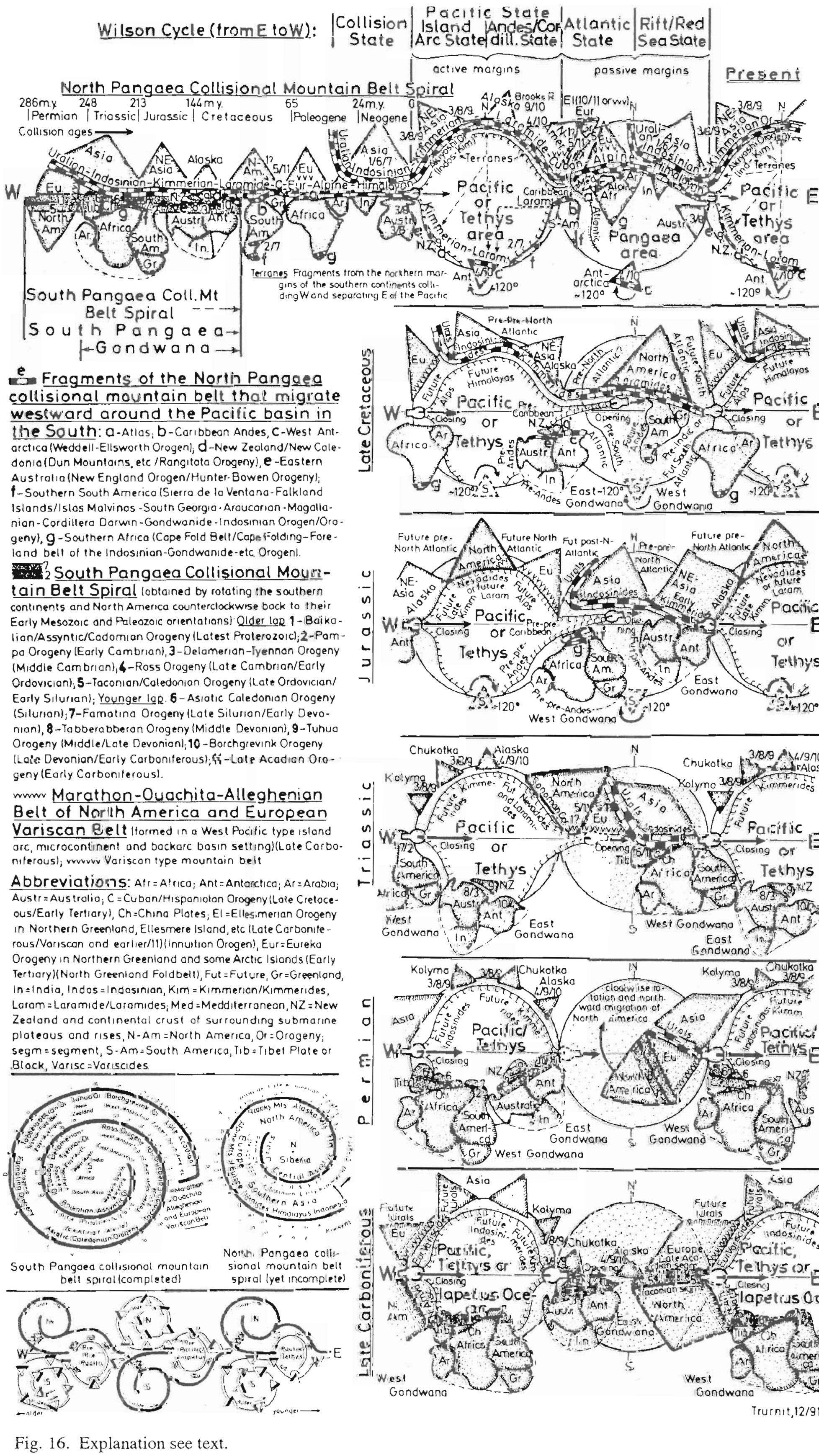


Fig. 16. Explanation see text.

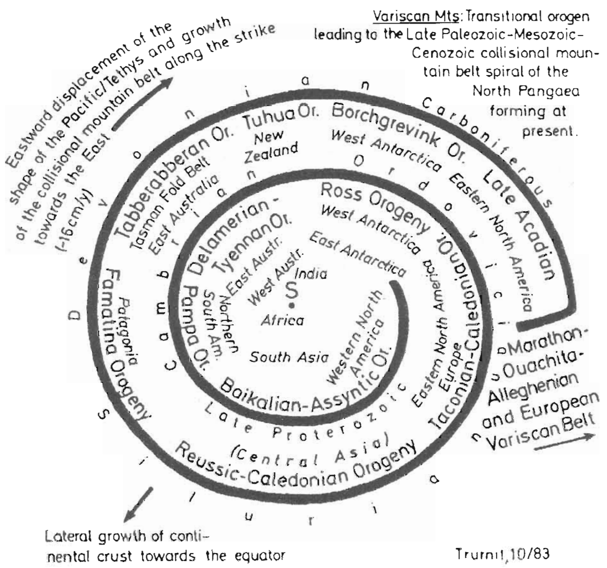


Fig. 17. Collisional mountain belt spiral of the Late Proterozoic and Early to Middle Paleozoic South Pangaea (sequence of orogenies; arrangement of continents; South polar view).

mountain belt at the equator to the West of the Pacific by the succeeding continent: Those of East Africa by India, those of India by Australia, those of Australia in future by Antarctica, etc. Naturally South America did not possess any island arcs when passing through the Australia setting during the Cretaceous. Relative to South America, Africa can be regarded as a former gigantic New Zealand type "microcontinent" and the South Atlantic as a still existing backarc basin between these two West Gondwana continents.

Through reverse or counterclockwise rotations of the southern hemisphere continents and reassembled megacontinents from their present orientation back to their Early Mesozoic and Paleozoic orientations (taking into consideration a time-lag on the initiation of rotation - Fig. 16) one arrives at the correct reconstruction of the Early Mesozoic Gondwana, of the Late Proterozoic to Middle Paleozoic South Pangaea and of its collisional mountain belt spiral (Figs. 16, 17). This South Pangaea and Gondwana reconstruction must satisfy the following requirements:

1. The South Pangaea collisional mountain belt (i. e. the collisions that caused it) has to become progressively older from East to West.

2. The younger lap of the collisional mountain belt spiral has to face the equator, the older lap the South Pole.

3. Analogous to the Permian to present-time North Pangaea collisional mountain belt, the Late Proterozoic to Middle Paleozoic South Pangaea collisional mountain belt had to be orientated during the whole era of its eastwards growth either parallel to the equator or in the southern hemisphere alongside the Paleozoic Pacific or Iapetus Ocean (pendular movements of the South Pangaea continents with the same margins always facing either the equator or the Pacific basin; Figs. 13, 16).

4. All Late Proterozoic to Middle Paleozoic (Early Carboniferous) orogenic fragments which at the present-time are dispersed individually all over the globe, either in the interiors of continents (e. g. Asia) or at a margin of a southern hemisphere continent (e. g. Tuhua orogeny of New Zealand; Pampa and Famatina orogeny fragment of southern South America), have to be accommodated in the reconstructed South Pangaea collisional mountain belt spiral (Fig. 16, 17).

In the top of Fig. 16 the right side depicts the present-time arrangement of continents around the globe and around the Pacific. On the left side, the North Pangaea collisional mountain belt leads from West of the Pacific westwards from the Neogene Himalayas via the Early Tertiary and Mesozoic segments of the belt to the Urals formed during the Permian. The southern hemisphere continents have been reassembled as the East and West Gondwana megacontinents and rotated counterclockwise back to their previous orientations during the respective eras in which they were attached to the northern continents and to the North Pangaea collisional mountain belt, between the collisions at the equator to the West of the Pacific/Tethys and before separating anew, East of the Pacific in the Mediterranean/Caribbean settings. The right side of Figure 16 depicts from top to bottom global continent/ocean distribution scenarios for the Present, the Late Cretaceous, Jurassic, Triassic, Permian and the Late Carboniferous with their respective segments of the North Pangaea collisional mountain belt (becoming less and shorter towards the Past/the bottom).

Due to the rotations of the West and East Gondwana megacontinents and their fragments during the Mesozoic/Cenozoic and the northward migration of Laurentia (North America and Europe welded together during the Late Acadian orogeny/collision in the Early Carboniferous) during the Late Carboniferous and the Permian, in this new Pangaea and Gondwana reconstruction, during the Late Jurassic and Cretaceous, South America and Antarctica only once fitted loosely together with West Antarctica (Antarctic/Palmer Peninsula) and northwestern South America (Caribbean Andes) through southern parts of the Cretaceous segments (Late Kimmerian-Laramide segments) of the older lap of the North Pangaea collisional mountain belt, in much the same way as Africa/Arabia and India do at present and in the future, India and Australia will be connected through southern parts of the North Pangaea collisional mountain belt.

While South Pangaea and its collisional mountain belt spiral were growing during the Late Proterozoic to Middle Paleozoic, the present East Antarctica margin to the East of the Antarctic Peninsula must have been connected to the present-time Arctic margin of North America, and the present western margin of South America (southern and central segments) must have been attached to the present-time northern margin of Australia (New Guinea) (Trurnit, 1984; 1988a, b; 1989a; 1991a, d; see left side of top drawing of Fig. 16). The hypothetical "Pacifica", "Chilenia", or "East Pacific" continent frequently mentioned in Central and South Andean geological literature (Miller, 1970; Nur & Ben-Avraham, 1979, 1983; Dalmayrac et al., 1980; Nur, 1984; Breitkreuz et al., 1988; Ramos, 1988a, b), into which the structural grain of the Coastal Cordillera of Southern Perú and Chile should continue (presumed by some authors to be lost or sunken in the Pacific to the West of South America), in accordance with the top drawing (left side) of Fig. 16, will most likely prove to be East Gondwana (New Guinea; parts of northern India subducted under Asia).

The Appalachians/Caledonides orogenic belt is that fragment from the two-lap South Pangaea collisional mountain belt spiral (Figs. 16, 17) (not having originated subsequently to the closure of an unitarian and single Iapetus ocean, not to mention the closure of a proto-Atlantic Ocean) which migrated North together with North America and Europe (Laurentia, a fragment of South Pangaea !) during the Late Carboniferous and Permian to collide against Asia (Kasakhstania/Siberia) during the Permian to form the Urals, i. e. the first segment of the North Pangaea collisional mountain belt and the nucleus of the future North Pangaea/Laurasia (Fig. 16). The remaining East and West Gondwana megacontinents collided against this Laurasia nucleus up till the Late Jurassic/Early Cretaceous and transmitted a large part of the remaining South Pangaea collisional mountain belt spiral (mainly parts of the younger lap) to the older lap of the North Pangaea collisional mountain belt spiral from the present-day Central and South Asia (Indosinian collision) via East and NE-Asia, Alaska, Western Canada and down to approximately the line of the Columbia River of the north-western United States (Kimmerian-Early Laramide collisions; Figs. 16, 17) (Coney et al., 1980; Churkin & Trexler, 1981; Churkin, 1983; Fujita & Newberry, 1983; Saleeby, 1983; Churkin et al., 1985 Howell, 1985; Howell et al., 1985; Parfenov & Natalin, 1986). Following these two megacontinets, the next continent to collide against Laurasia or North Pangaea at the equator to the West of the Mesozoic Pacific/Tethys was South America during the Late Cretaceous Laramide collision/orogeny. It was the first continent that did not collide with a margin containing South Pangaea collisional mountain belt fragments. From the fragments of the South Pangaea collisional mountain belt spiral that still remain at present with the Gondwana continents in the southern hemisphere (many segments from the older lap; Miller, 1981), some will soon be incorporated via the SW-Pacific island arcs and microcontinents (New Zealand - Middle to Late Devonian Tuhua collision; Antarctica - Late Cambrian to Early Ordovician Ross and Late Devonian/Early Carboniferous Borchgrevink collisions; Figs. 15 - 17) into the younger lap of the North Pangaea collisional mountain belt spiral.

The Appalachians/Caledonides orogenic belt did not result from the closure of a proto-Atlantic Ocean (Wilson, 1966) but represents a fragment from the Late Proterozoic and Early to Middle Paleozoic two-lap South Pangaea collisional mountain belt spiral marking the trail of the eastward migrating Late Proterozoic to Middle Paleozoic Pacific or Iapetus Ocean (Figs. 15 - 17). In this fragment from the South Pangaea collisional mountain belt spiral, the Paleo-Iapetus Ocean was closed by the

Late Ordovician to Early Silurian Taconian collision and the Neo-Iapetus by the Early Carboniferous Late Acadian collision. Paleo- and Neo-Iapetus Oceans have their Mesozoic-Cenozoic equivalents in the Paleo- and Neo-Tethys Oceans (Figs. 3, 12, 13, 15 - 17) (Sengör, 1979, 1985, 1986, 1987).

The sequence of collisions (paroxysms of orogeny) that formed the individual segments of the reconstructed South Pangaea collisional mountain belt spiral (Figs. 16, 17) from the former East to the former West and from the younger towards the older is: Late Acadian (Appalachians/Caledonides of Eastern North America and Western Europe: Early Carboniferous), Borchgrevink (West Antarctica: Late Devonian/Early Carboniferous), Tuhua (New Zealand: Devonian), Tabberabberan (East Australia: Devonian), Famatina (Patagonia and Southern Andes of South America: Late Silurian/Early Devonian), Reussic or Asiatic-Caledonian (South and Central Asia: Silurian), Taconian-Caledonian (Appalachians/Caledonides of Eastern North America and Western Europe: Late Ordovician to Early Silurian), Finnmarkian (Late Cambrian/Ordovician), Ross (Antarctica: Late Cambrian/Early Ordovician), Delamerian-Tyennan (East Australia: Cambrian), Pampa (Patagonia and Southern Andes of South America: Early Cambrian), Baikalian/Assyntic/Cadomian (Central and South Asia, Europe, Eastern North America ?: Late Proterozoic) (Trurnit, 1984c; 1988a, d; 1989a; 1991a, d).

The Late Proterozoic Pre-North Pangaea

A fragment from the pre-North Pangaea collisional mountain belt spiral of Panafrican-Brasiliano age is contained in West Gondwana (Trurnit, 1984a;1988a, b; 1989a; 1991a, d) (Fig. 18). The

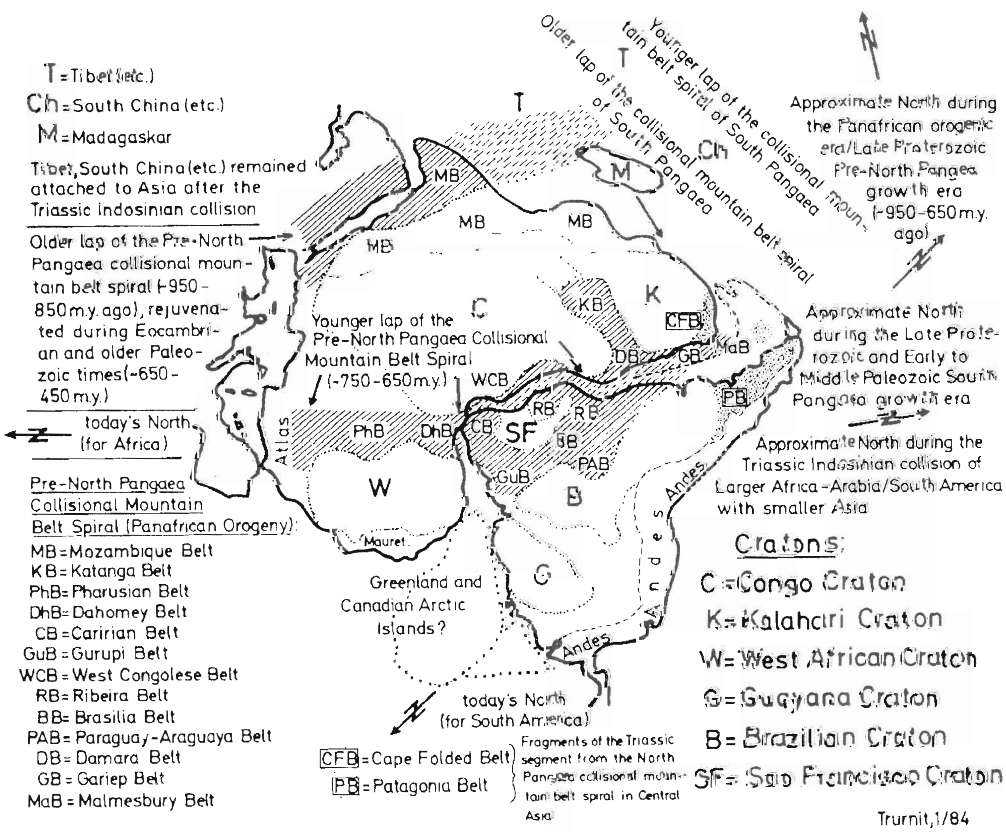


Fig. 18. The South America/Greenland ? /Africa/Arabia fragment from the pre-North Pangaea collisional mountain belt spiral (Late Proterozoic Panafrican/Brasiliano orogenic cycle). Fragments off the East coast of the present-day Africa with segments from both laps of the South Pangaea collisional mountain belt spiral and a fragment from the older lap of the pre-North Pangaea collisional mountain belt spiral remained attached to the then smaller Asia after the Triassic Indosinian collision (closing of the Paleo-Tethys). Asia at present displays a chaos of mountain ranges from five laps (only counting those of the last 950 m.y. and not taking into consideration Variscan-type mountain belts) belonging to three different collisional mountain belt spirals (Figs. 3, 12, 16, 17).

Pharusian-Dahomey-West Congo-Brasiliano-Mantiqueira-Ribeira-Paraguay/Araguaya-Kaoko-West Damara-Gariep-Malmesbury segment of the Pre-North Pangaea collisional mountain belt spiral in Eastern South America and Western Africa during the pre-North Pangaea growth era (approximately from 950 m. y. to 650 m. y. ago) was most probably facing the equator and belongs to the younger lap, the Hijaz-Mozambiquian segment in Eastern Africa and Arabia was facing the North Pole and belongs to the older lap. Following the Triassic Indosinian collision between the then smaller Asia and the formerly more extended southeastern/eastern margin of the then greater Africa/Arabia (part of West Gondwana), parts of the Panafrican Mozambiquian segment of the pre-North Pangaea collisional mountain belt spiral (older lap) and of the former East African fragment from the last South

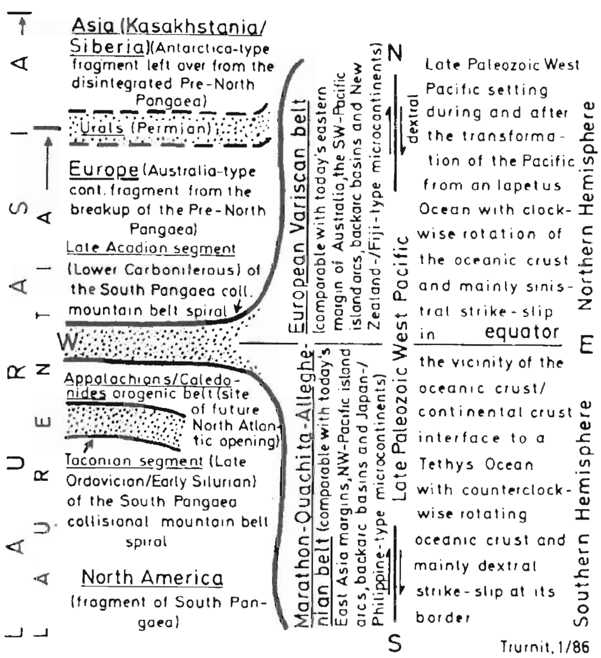


Fig. 19. The formation of the southern North America Marathon-Ouachita-Alleghenian and European Variscan belt in a West Pacific island arc, microcontinent and backarc basin setting during and after the Late Paleozoic era of reorganization from a South Pangaea growth/Pre-North Pangaea breakup to a North Pangaea growth/South Pangaea breakup. The European Variscan belt and Europe link the end of the South Pangaea collisional mountain belt (Late Acadian segment) with the first segment of the North Pangaea collisional mountain belt (the Urals). For simplicity the Taconian and Late Acadian segments are depicted as separate belts and not superposed on another as they in fact are. Since the Early to Middle Paleozoic Europe served as foreland for the Caledonides, the Variszides, the Urals and the Alpine Mountain chains.

Pangaea collisional mountain belt spiral (both laps; Figs. 16 - 17) remained attached to the Early Mesozoic Asia which at present displaysa chaos of mountain ranges from five laps (only counting those of the last 950 m. y. and not taking into consideration Variscan type mountain belts) belonging to three different collisional mountain belt spirals (Figs. 3, 12, 16 - 18). Strike-slip movements along the trends of the belts and the agglomeration and mixture of tectonostratigraphic terranes during the growth of the individual spirals have additionally contributed to the chaos (Trurnit, 1984c; 1988a, d; 1989a; 1991a, d).

The formation of Variscan type mountain belts in West Pacific type island arc, microcontinent and backarc basin settings

During eras of reorganization from a South Pangaea growth/North Pangaea breakup to a North Pangaea growth/South Pangaea breakup and vice versa (i. e. every 400 to 500 m. y.; Figs. 12, 21) the two-sided, bilaterally verging collisional mountain belt formed parallel to latitude/equator at the equator to the West of the Pacific is interrupted by one-sided, unilaterally verging Variscan type mountain belts that form parallel to longitude in a pre-collisional SW- and NW-Pacific island arc, microcontinent and backarc basin setting (Fig. 19).

The Late Carboniferous scenario for the European and North American Variscan mountain belt (Trurnit, 1988a, c, d; 1989a, c; 1991a, d) is a reflected image of the present West Pacific setting with the South Pangaea collisional mountain belt spiral having just been completed (Late Acadian collision between Europe and North America; Figs. 16, 17) and the Pacific in a state of reorganization from a Middle Paleozoic Neo-Iapetus Ocean with a clockwise rotating Pacific/Neo-Iapetus Plate (type of the present Pacific Plate) to a Late Paleozoic and Early Mesozoic Paleo-Tethys Ocean with a counterclockwise rotating Pacific/Paleo-Tethys Plate (Pacific Plate type of the present-time Pacific Ocean). In the southern hemisphere, the southern North American Marathon-Ouachita-Alleghenian belt was arranged from South to North up to the equator in a setting comparable to the present-time NW-Pacific island arcs, microcontinents and back-

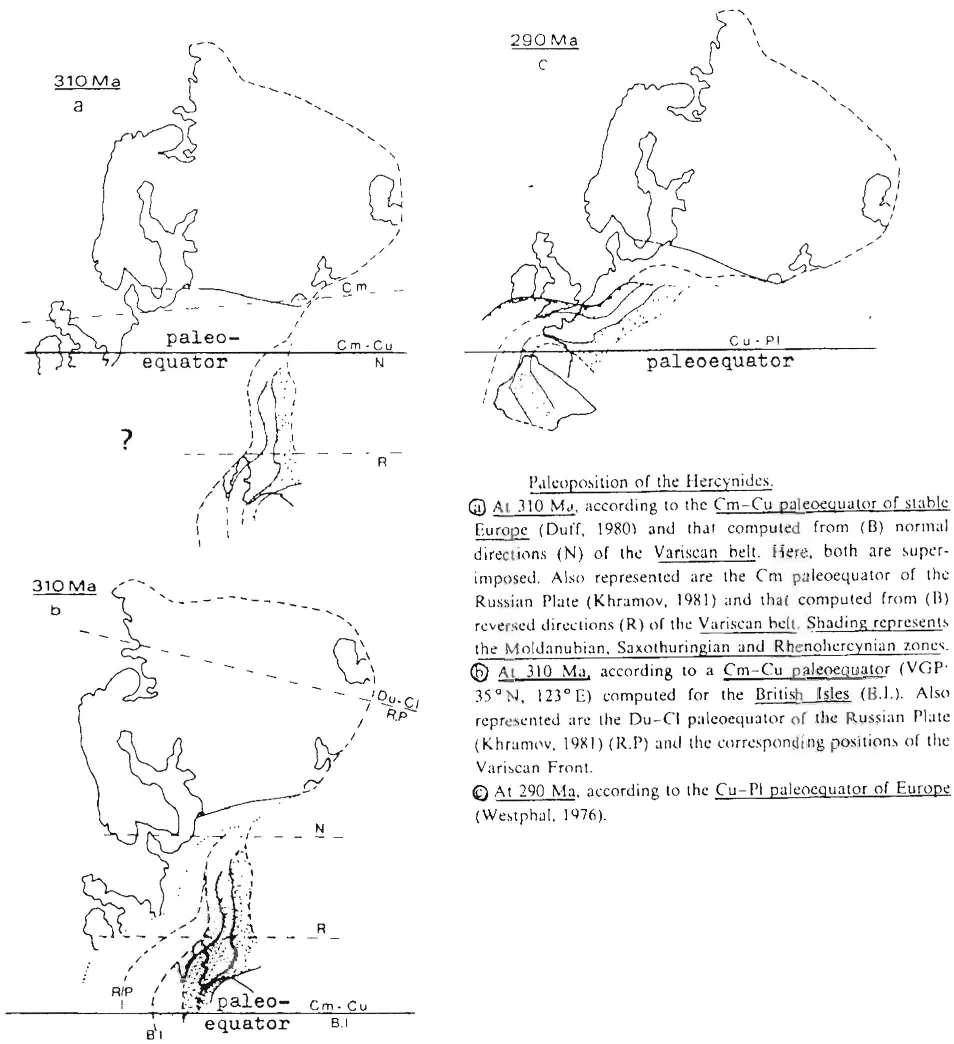


Fig. 20. Paleopositions of the European Variscan belt during the Late Carboniferous according to Edel (1987, Fig. 3): 310 m.y. ago Europe might have been positioned either to the North (a) or to the South of the equator (b); 290 m.y. ago Europe lay at the equator and had started to rotate clockwise. Shaded are the Rhenohercynian, Saxothuringian and Moldanubian zones of Kossmat (1927) (see also Fig. 16: Late Carboniferous and Permian).

arc basins (mirror image symmetry). In the northern hemisphere the European Variscan belt was arranged from the equator towards the North in a setting comparable to the present SW-Pacific island arcs, microcontinents and backarc basins (mirror image symmetry).

Dextral transcurrent faults running more or less parallel with the trend of the Variscan mountain belt were caused by the northward migration of Laurentia (North America and Europe; Fig. 19) during the Late Carboniferous and the Permian (Edel, 1987) and before termination of the collision between Laurentia and Asia (Figs. 16, 20a-c) (initiation of the North Pangaea collisional mountain belt with the formation of the Urals). Europe and the Variscan belt connect the end of the South Pangaea collisional mountain belt (Late Acadian segment) with the first segment of the North Pangaea collisional mountain belt (the Urals; Figs. 3a, 9, 12, 14, 16, 17, 19). During and subsequent to the formation of the Urals during the Permian, Laurasia rotated clockwise in the northern hemisphere (Figs. 16, 20a-c) (Edel, 1987). The Marathon-Ouachita-Alleghenian belt of southern North America and the European Variscan belt were separated during the Late Mesozoic/Tertiary opening of the North

T.P. Trurnit : Present plate tectonic setting of the arabian plate

Atlantic Ocean along the Appalachians/Caledonides segments of the South Pangaea collisional mountain belt spiral (Fig. 16).

Late Mesozoic and Early Tertiary setting of the Greenland Plate compared to the present-day setting of the Arabia Plate

The North America paradox (Atlantic versus Pacific margin) has possibly resulted from a major error in North and Central Atlantic continent reassemblies that inhibits advancement in Historical Geology and Global Tectonics: The "West Africa Connection" (assumed former connection between the eastern margin of North America and the western margin of North Africa). Since Wegener (1915, 1929), Choubert (1935), Du Toit (1937), Bullard et al. (1965) and others reassembled the continents surrounding the present-day Atlantic Ocean into Pangaea, the single Late Paleozoic/(Early Mesozoic ?) supercontinent, in the absence of an alternative model, no one has ever challenged their axiomatic reconstruction apart from minor adjustments (e. g. Pitman & Talwani, 1972; Emery & Uchupi, 1984; Sheridan & Grow, 1988; Ziegler, 1988; etc.). While the former fit between Africa and South America is beyond doubt and that between Europe and North America only needs readjustment, the origin of Greenland is still most obscure (Lefort, 1989) and the presumed former connection between Eastern North America and Western North Africa requires critical reappraisal. An alternative model now exists (Trurnit, 1983a-c; 1984a-d; 1985a-d; 1986a-d; 1987a-c; 1988a-d; 1989a-e; 1990a-c; 1991a-d. Its relevance for the history of the opening of the Central and North Atlantic Oceans, for the presumed northward migration of Greenland during the Tertiary as well as for the future opening of a Red Sea-Gulf of Aqaba-Levant Fault (post-North Atlantic; rift propagation towards the North) (Bushara, 1989; Cisternas et al., 1989; Philip et al., 1989) and the northward drift of the Arabian Plate is outlined below (Figs. 1, 13, 14, 16, 18).

Geodynamically, a "West Africa Connection" appears improbable for the following reasons:

1. According to the inhomogeneous, distorted global stress field induced by the combined effects of thermal convection and gravitational, centrifugal and tidal forces (Fig. 5) (Trurnit, 1991d), a collision between continents from the northern and southern hemispheres only takes place parallel and close to the equator and only behind and to the West of the eastward migrating Pacific basin (Figs. 3c, 13, 15, 16). During their formation the Late Cretaceous Laramides collisional mountain belt of present-day Western North America must therefore have been orientated parallel to the equator and North America subsequently been rotated clockwise through 45° to 60° into its present orientation during the Tertiary (Figs. 14, 16). The Caribbean Andes and possibly Cuba and Hispaniola must be southern fragments from the Laramide orogen (Figs. 15, 16).

2. In the extra-Pacific or Pangaea area the continents from the northern and southern hemispheres are connected to each other always and only through a collisional mountain belt which is orientated more or less parallel to equator/latitude and which has collision ages between 0 and up to 100 m. y. (Figs. 3, 9 - 16). Connections between northern and southern hemisphere continents without a separating, comparatively young collisional mountain belt do not exist and have never existed.

3. According to the distorted, inhomogeneous global stress field (Fig. 5) the post-collisional disengagements between the continents from the northern and southern hemispheres (North and South Pangaea; Laurasia and Gondwana, etc.) take place only parallel to latitude/equator and only to the East and ahead of the eastward migrating Pacific basin in the extensional Mediterranean and Caribbean settings (Figs. 3c, 5-7, 10, 13, 15, 16). The disconnecting rift ocean forms more or less parallel to the trend of the orogen/collisional mountain belt, either in the midst of the belt, shifted somewhat equatorwards in the orogen or in the foreland positioned on a continent of the Pangaea breaking up. During their early evolution, neither the western passive margin of North Africa nor the eastern passive margin of North America formed as margins of such a parallel to equator Mediterranean type ocean (tongue of the Pacific) (Figs. 3c, 13, 15-16).

4. Subjected to the Coriolis Force, North America must have rotated clockwise during its northward migration during the Tertiary (Figs. 14, 16).

5. Due to migration northwards of the Pacific Plate (Trurnit, 1986a, 1988a) (fig. 11) the present-day western (Late Cretaceous southern) margin of North America must have been drawn northwards (San Andeas fault system) and as a result, the North American continent rotated clockwise once the continent had arrived East of the Pacific basin and of the Pacific Plate, during the Middle to Late Tertiary (Figs. 13, 14, 16).

In the light of the above, it appears impossible that the eastern margin of North America was formerly attached to the western margin of North Africa. Geodynamic considerations support the conclusion that the western margin of North America, presently orientated in a NW-SE direction alongside the NE-Pacific, collided during the Late Cretaceous Laramide orogeny with the northern margin of South America (Caribbean Andes, Cuba, Hispaniola) close and parallel to the equator and to the West of the Late Mesozoic Pacific/Tethys. During the post-Laramide era North America should have rotated clockwise through 45° to 60° from the Early Tertiary onwards into its present-day orientation (Figs. 13, 14, 16), the eastern margin of North America should have been positioned in more northerly latitudes and until Early post-Laramide times, should have been connected to the western margin of Europe but not to the western margin of North Africa ("West Africa Connection" a major error !) (Fig. 14). Numerous paleomagnetic measurements mainly on Tertiary magmatic rocks of Western North America confirm a clockwise rotation of the continent through 45° to 60° during the Tertiary (not explainable alone by local block rotations caused by dextral strike-slip between the western margin of North America and the Pacific Plate, by oblique collisions of terranes or by coastwise transport of tectonostratigraphic terranes along the convex margin of Western North America; Beck, 1976, 1980; Kissel & Laj, 1989). The future detection of suspected sinistral movements between the eastern margin of North America and the Mid-Atlantic Ridge (oblique spreading; ballbearing-like rotations of microplates along the ridge - Hey et al., 1985; Passerini et al., 1991; slip along major faults off the coast - Sheridan & Grow, 1988) (Figs. 13,16) and a reinterpretation of the seafloor spreading magnetic anomalies in these realms might assist in clarifying the North America paradox.

In addition to the presumed clockwise rotation of North America during the Tertiary, a high probability exists that Greenland was formerly connected to Northeastern South America and Northwestern Africa (Figs. 13, 14, 16, 18). The Late Cretaceous/Early Tertiary Eureka Orogen positioned at the northern margin of Greenland and on the Canadian Arctic Islands, possibly fits in between the Late Cretaceous Laramide Caribbean Andes/Cuba/Hispaniola Orogen in the West and the Early Tertiary Alpine Atlas Mountains of North Africa in the East (southern parts of the Late Cretaceous/Early Tertiary Laramide-Alpine collisional mountain belt). The East Greenland Fold Belt (Early Paleozoic orogeny) and the polyphase Mauretanides-Rokelides Belt of West Africa (Late Proterozoic, Early to Middle Paleozoic) (Williams, 1984; Barker & Gayer, 1985; Deynoux et al., 1985; Dallmeyer et al., 1987) (Figs. 14, 16) should be one and the same orogenic belt. Analogues to the Permian and Triassic faunal and floral assemblages from East Greenland (e. g. Cape Stosch, etc.) exist in European Russia but also in Gondwana Regions such as the Salt Range of Pakistan, the Indian Himalaya, Madagascar and Western Australia (Haller, 1971; Birkelund et al., 1973;Birkelund & Perch-Nielsen, 1976; Balme, 1979; Hallam, 1986b). Climatic conditions indicating a northward migration of Greeland during the Tertiary (coal, petrified forets, subtropical and tropical fauna and flora) are mentioned by Miall (1984, 1985), Steel et al. (1985), McMillan (1986) and Francis & McMillan (1987). Greenland (the Arabia of South America ?) could have drifted North since post-Laramide, Early to Middle Tertiary times (Pelocene/Eocene Thule volcanism) into its present-day position from a plate tectonic setting which corresponds to the present-day setting of Arabia (the Greenland of Africa ?) to the South of the Tertiary Alpine collisional mountain belt (Figs. 1,13-16, 18).

Due to migration eastwards of the individual states of the Global Tectonic Megacycles around the globe (e. g. the ocean states of the Oceanic, Eugeosynclinal or Wilson Cycle, the states of the Orogenic Cycle and of the Cycle of the Collisional Mountain Belt; Trurnit, 1984c, 1988a, d; 1989a;

1991a, d; Figs. 9-12, 16), relative to a distinct plate tectonic setting of an ocean, a continent or continental margin, a future or later evolutionary state at the Earth´s surface is always depicted in a setting simultaneously developed further to the West and a past or earlier state in a setting simultaneously occurring further to the East. Concurrently with the future opening of a post-North Atlantic Ocean in Eastern Europe or between Europe and Asia, Arabia might probaly continue to force its way North through the Alpine Belt and the Russian Platform of Eurasia (Figs. 1, 16). It should in future arrive at a plate tectonic setting comparable to that presently occupied by Greenland (Figs. 14, 16) (Trurnit, 1990a-c; 1991b).

In the foreland, to the South of the North Green-

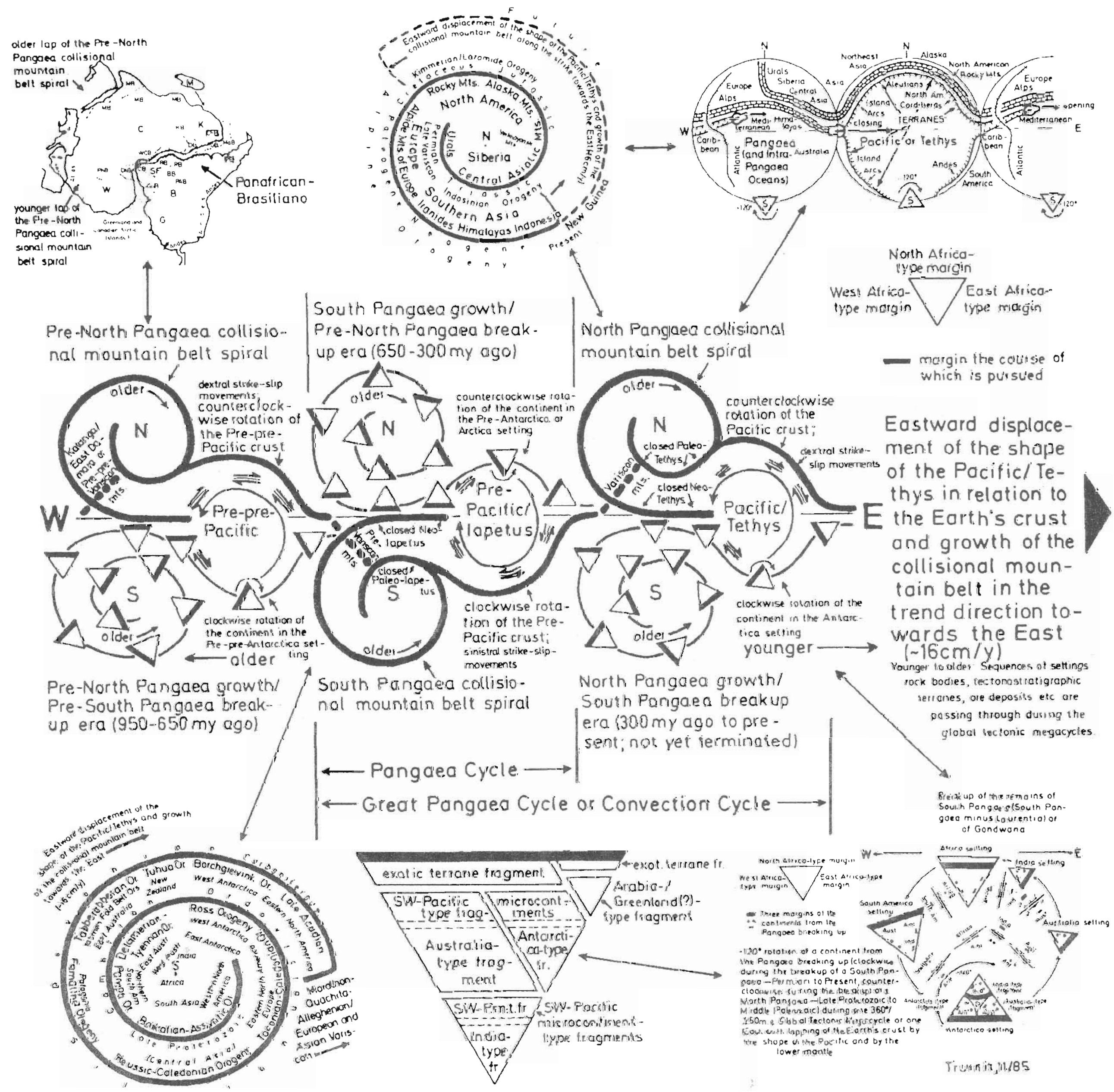


Fig. 21. The Great Pangaea Cycle or Convection Cycle: Two successive Global Tectonic Megacycles of 360°/200 to 250 m. y. duration (Oceanic Cycle or Wilson Cycle followed by the Cycle of the Collisional mountain belt, etc.), two laps of the collisional mountain belt spiral or two East drift lappings of the Earth´s crust/the continents by the Pacific basin and the ocean states of the Wilson Cycle correspond to one Pangaea Cycle; two Pangaea Cycles amount to one Great Pangaea or Convection Cycle. Pangaea Cycle and Great Pangaea Cycle are comparable to the 11 year and 22 year sunspot cycles (magnetic solar cycles).

land Fold Belt (Innuition Orogen; Ellesmerian Orogeny; Devonian to Late Carboniferous - Variscan and earlier; Eureka Orogen/Orogeny: Late Cretaceous/Early Tertiary) hydrocarbon occurrences could be expected in quantities comparable with the foreland deposits of the Himalayan-Alpine-Laramide-Kimmerian (Atlas-Caribbean Andes-West Antarctica-SW-Pacific microcontinents) Belt in Burma, Bangla Desh, India, Pakistan, Arabia, North Africa, Northern South America, (East Antarctica ?; SW-Pacific backarc basins - Lord Howe Rise/Chalenger Plateau) and California.

Conclusions

Oceans do not simply open and close periodically above approximately the same longitudes and latitudes with the same margins always bordering the same expanding and later contracting ocean (Wilson, 1966). Opening and closing of an ocean are in fact steered by the lower mantle rotating somewhat faster relative to the convecting upper mantle-crust system (tidal forces; principle of hypocycloid gearing) and the Pacific basin, i. e. the sequence of ocean states of the Wilson Cycle, migrating East relative to the upper mantle-crust, the Pangaea areas and the continents as a result, with one full turn around the globe completed in 200 to 250 m. y., in that the identity or state of an ocean in the West is slowly but constantly transferred to the neighbouring ocean in the East. The sequence of ocean states of the Wilson Cycle is just one in a series of sequences of other cycles (Oceanic, Eugeosynclinal or Wilson Cycle; Cycle of Continental Margins or Miogeosynclinal Cycle; Orogenic Cycle; Cycle of the Collisional Mountain Belt; Cycle of the Oceanic Crust and Eugeosynclinal Sediments or Ophiolitic Cycle; Cratonic Cycles; Pangaea Cycle; Great Pangaea or Convection Cycle), the sedimentological, magmatic and tectonic signatures of which are imposed on one another over and over again (Figs. 2, 9-12, 16, 21).

This new concept demonstrates that throughout the more recent history of the Earth (since the beginning of the Proterozoic at least), the general ocean/continent distribution pattern should always have been similar to that of the Present (with the exception of Pangaea reorganization eras during Variscan type mountain building). Relative to a generally fixed East to West and younger/earlier to older/later arrangement of ocean types, the continental margins and continents have always passed West through definite and similar sequences of plate tectonic settings (Oceanic or Wilson Cycle, Orogenic Cycle, etc.) between their birth in a rift in the Pangaea area and their death in the collision zone at the equator to the West of the eastward migrating Pacific basin or area. The arrangement of ocean types of the Present, of the present-time passive and active continental margins, of the Himalayan-Alpine collisional mountain belt between South Asia and Europe, and further on via the Laramides of Western North America, Alaska, NE-Asia, East Asia, South and Central Asia to the Urals, that exhibits a westward directed sequence of gradual evolution and ageing (geosynclinal and orogenic models) represents an actualistic scenario for the pre-, syn-, and post-collisional evolution of older collisional mountain belts, vanished oceans, continents, continental margins and the general continent/ocean distribution pattern of the Past.

Due to migration eastwards of the individual states and the sequences of states of the global tectonic megacycles around the globe, relative to a distinct plate tectonic setting of an ocean, a continent or continental margin, a future or later evolutionary state at the Earth´s surface is always depicted in a setting simultaneously developed further to the West (that of the Arabian Plate in the Greenland setting; that of the passive West Africa type margin in the setting of the active Andean type) and a past or earlier state in a setting simultaneously occurring further to the East (that of the Greenland Plate in the Arabia setting; that of the active Andean type margin in the setting of the passive West Africa type). Concurrently with the future opening of a post-North Atlantic Ocean in Eastern Europe or between Europe and Asia, Arabia (the Greenland of Africa ?) might probably continue to force its way North through the Alpine Belt and the Russian Platform. It should in future arrive at a plate tectonic setting comparable to that presently occupied by Greenland (the Arabia of South America ?). After having disengaged from Asia in the near geological future, Europe will rotate clockwise as North America has done during post-Laramide times since the Eocene/Oligocene. A West Indian Ocean will become a South Atlantic, a Red Sea-Gulf of Aqaba-Levant Fault a Central/North Atlantic-Baffin Bay and a Persian Gulf-Mesopotamian Plain a North Atlantic. Like the ocean states of the Wilson Cycle, plate tectonic characteristics move eastwards (the character of the Greenland Plate to the Arabian Plate, etc.) with the angular velocity of the Pacific basin and of the lower mantle.

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Geochemistry and crustal P-T polymetamorphic path of the mantle-derived rocks from the Pohorje area (Austrides, Eastern Alps, Slovenia)

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Abstract

These eclogites and metaultrabasites have been recently studied by Hinterlechner-Ravnik et al. (1989, 1991a, b) from the petrographic, mineral chemical and geothermobarometric points of view. The aims of the present paper is that of outlining and interpreting the main geochemical features of these rocks and integrate these new results with those supplied by the literature.The eclogite lenses may present the metamorphic products of fragmented ocean floor basalts with N/MORB affinity, while the major "metaultrabasic" body a layered gabbro body. The whole complex within which these rocks occur may represent an old tectonic mélange, deeply reworked by the Variscan and Alpine metamorphisms. Eclogites and metaultrabasites record a three-stage crustal evolution:(i) an older, dry, high-temperature stage; (ii) a later, lower-pressure granulitic stage; (iii) a younger, epidote-amphibolite overprint. The latter recrystallization may be Alpine, while the main part of the metamorphic history is pre-Alpine.

KEY WORDS : eclogites, ultrabasites, geochemistry, polymetamorphism, Austrides, Eastern Alps.

Introduction

Eclogites, metabasites and metaultrabasites occur in a number of localities of the Austridic basement in the Eastern Alps (see Hinterlechner-Ravnik et al., 1991, for the references).

The rocks considered in the present paper make up variously sized bodies in the Austridic basement of the Pohorje area (Fig. 1), which is the most south-eastern part of the Austridic basement of the Eastern Alps. Important Alpine tectonic lines border the Pohorje crystalline block to the west (southern prolongation of the Lavanter Line) and to the north (phyllonites in contact with the Miocene molasse); to the east and to the south thePohorje crystalline block subsides, buried by the young, Quaternary and late Tertiary, sediments of the Pannonian Basin. Detailed data on the Pohorje area are given in Hinterlechner-Ravnik (1971, 1973, 1982, 1988) and Hinterlechner-Ravnik & Moine (1977), from which most of the general information given in the following section is taken. The eclogites and metaultrabasites have been recently studied by Hinterlechner-Ravnik et al. (1989, 1991a, b) from the petrographic, mineral chemical and geothermobarometric points of view. The aim of the present paper is that of outlining and interpreting the main geochemical features of these rocks and integrate these new results with those supplied by the above three quoted papers.

Geo-petrographic outlines

The tonalitic Pohorje massif occupies the central part of the Pohorje crystalline block. It is surrounded by metamorphic rocks of various types, in which four main rock complexes have been distinquished:

1) an augengneiss-rich complex, which is the deepest part of the whole rock sequence; the augengneisses have rhyolitic to rhyodacitic composition, and are interlayered with biotite-muscovite paragneisses;

2) an overlying marble-rich complex, in which marbles, amphibolites,quartzites and schists are repeatedly interlayered;

3) an overlying eclogite-rich level, in which prevailing micaschists and paragneisses include numerous small eclogitic and metaultrabasic lenses, as well as related abundant amphibolites; the major metaultrabasic ("serpentinite") body occurs in the lower part of the eclogite-rich level, and it also includes numerous eclogite lenses; southwards it is cut by a fault and partly buried by the young sediments of the Pannonian basin;

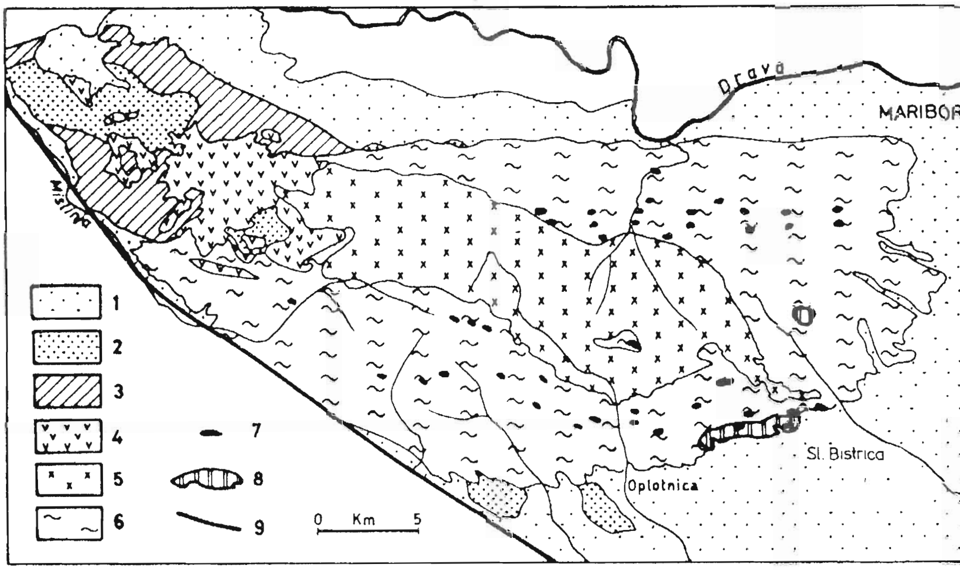

Fig. 1. Geological sketch of the Pohorje Mountains (taken from Hinterlechner-Ravnik, 1982). 1 - Quaternary and Late Tertiary covers, 2 - Permo-Mesozoic cover, 3 - Silurian-Devonian cover, 4 - Dacites, 5 - Granitoid pluton, 6 - Austridic micaschists, paragneisses and related rocks, 7 - Eclogites, 8 - Metaultrabasites, 9 - Main tectonic boundary.

4) finally, in the uppermost part, a paragneissic complex bearing minor rock intercalations; the almandine-amphibolite facies mineral assemblages formerly prevailing in this paragneissic sequence are retrogressed into greenschist facies assemblages, and the paragneisses grade to phyllonites due to Alpine deformations.

Most of the above mentioned rocks in the whole metamorphic sequence are B-tectonites. Blastomylonitic bands occur sometimes in different complexes, in particular at the base of the eclogite -rich level, which could be interpreted as a thrust rock sheet.

Only minor parts of the Pohorje metamorphic rocks are to be referred to progressive greenschist facies, considering that the main part of them (including the rocks in which eclogites and metaultrabasites occur) bear mineral assemblages related to the almandine-amphibolite facies.

The country rocks of the eclogites and metaultrabasites display at least a two-stage metamorphic development.

The older stage (believed to be early Hercynian or "Caledonian"), experienced significantly high temperature (> 600 °C) and pressure (> 6 kb) values. With reference to Fig. 2, this estimation is based on:

(i) the reaction: muscovite + quartz → Al silicate + K feldspar + H_2O,

(ii) the fact that the Al-silicate phase is kyanite and that it coexisted with staurolite,

(iii) the occurrence of foliated pegmatitic and aplitic veins and bands, which are locally abundant in the eclogite-rich level in the Pohorje area (Hinterlechner-Ravnik, 1988), and were also reported in the Saualpe eclogites (Manby et al., 1988).

The younger stage (Alpine in age) is related to thermic conditions corresponding to the boundary between the greenschist facies and the lower amphibolite facies (A in Fig. 2). This is suggested because of the common occurrence of post-tectonic chloritoid in the above mentioned phyllonites, and the local appearance of post-tectonic staurolite in the micaschists of the underlying eclogite-rich level.

Eclogites and metaultrabasites usually make up conformable lenses, boudins and bands, which are tectonically related to their present country-rocks. Eclogites are up to 1 km long and some 10km thick, but usually smaller. The major metaultrabasic body is 5 km long, and 200 - 1000 m thick. The country-rocks of the eclogites and metaultrabasites are amphibolites and micaschists. At least part of the amphibolites are completely retrogressed eclogites, as the field relationships and petrographic features (including relics) indicate. As mentioned above, eclogites also occur within the major ultrabasic body.

Eclogites, coronitic to symplectitic eclogites,amphibolitized eclogites and related layered amphibolites can be distinguished in the field, as well as light and dark varieties of eclogites. Kyanite-zoisite bearing eclogites also occur, and are interpreted as old plagioclase cumulates.

Harzburgites and dunites are reported to have been the main protoliths of the serpentinite body (Hinterlechner-Ravnik, 1982, 1988). Compositional layering is rather common, as well as lenses, dykes and veinlets of peculiar composition (e.g. garnet peridotites and coarse-grained veinlets of clinopyroxene).

All rock types between almost unaltered eclogites and deeply symplectitized eclogites have been

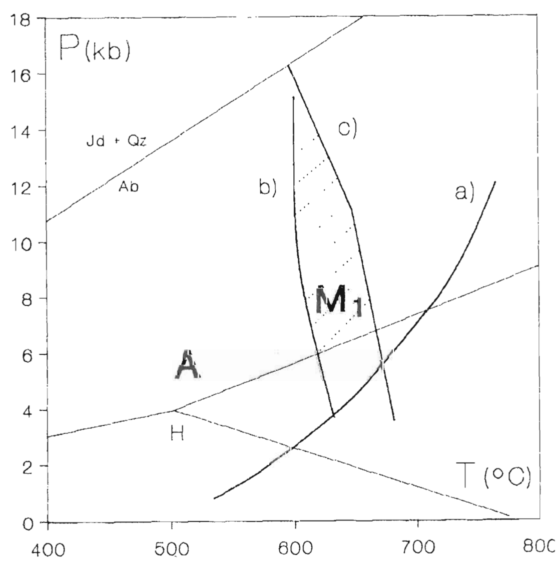

Fig. 2. P-T location of the main pre-Alpine metamorphism (M_1) and the Alpine overprint (A). H - Al_2SiO_5 triple point (Holdaway, 1971). a = Ms + Ab + Qtz $\rightarrow$ Al_2SiO_5 + Bt + Kf + liq, b = beginning of melting in H_2O - saturated metapelites, c = upper stability boundary of staurolite. (All curves redrawn from Koons & Thompson, 1985).

observed, as well as more or less serpentinized coronitic metatroctolites, garnet peridotites, garnet pyroxenites, forsterite-tremolite schists and deeply serpentinized ultrabasites. Related amphibolites also occur, as mentioned before, and can be considered the end-product of the eclogite alteration.

Geochemical features

Table I shows the representative chemical analyses of the main rock types. In general, the contents of SiO_2, TiO_2, K_2O, Rb, Nb and ΣREE are low, those of MgO, Sc, Co and V are relatively high, whereas the values of Al_2O_3 fall within a large range. The derivation of these metamorphic rocks from basic and ultrabasic igneous protoliths having a moderate cumulitic character turns out to be very clear from these chemical data consistently with petrographic data. However, effects of a moderate chemical mobilization (e.g. a loss of SiO_2, and gain of K_2O, Rb and Sr) during metamorphism may also be detected in them.

TAB. 1

Sample	142	142h	244	255	262a
SiO_2	39.62	42.15	49.97	46.07	44.93
TiO_2	0.04	0.22	0.41	0.10	0.12
Al_2O_3	5.94	26.52	15.08	15.41	15.33
Fe_2O_3*	9.71	5.48	5.75	5.58	8.16
MnO	0.11	0.08	0.11	0.08	0.14
MgO	30.60	6.79	10.14	18.58	17.76
CaO	3.57	14.57	13.58	12.41	11.28
Na_2O	0.19	1.20	3.64	0.94	0.86
K_2O	0.01	0.39	0.03	0.01	0.01
P_2O_5	0.08	0.11	0.08	0.07	0.08
LOI	9.85	1.60	0.21	0.32	1.35
Total	99.72	99.11	99.00	99.57	100.02
Mg #	86.19	71.05	77.74	86.83	81.17
Cr	2360	165	1363	1992	1372
Ni	1210	267	182	493	354
Co	123	52	108	123	96
Sc	15	16	44	26	25
V	43	121	218	86	95
Cu	83	11	16	70	84
Zn	51	24	42	34	39
K	83	3238	249	83	83
Rb	-	17	-	7	8
Ba	17	86	6	-	21
Sr	20	528	43	56	136
Ga	8	12	11	6	-
Nb	< 5	< 5	6	< 5	< 5
Zr	9	16	18	6	7
Ti	240	1319	2458	600	719
Y	1	10	10	4	4
La	0.32	1.77	0.70	0.64	0.35
Ce	0.82	5.01	3.17	2.63	2.27
Nd	0.28	2.76	0.80	0.90	0.74
Sm	0.49	1.01	0.69	0.66	0.69
Eu	0.04	0.68	0.18	0.16	0.19
Gd	0.30	1.10	0.70	0.60	0.62
Dy	0.14	1.32	1.43	0.58	0.66
Er	0.24	0.97	1.13	0.47	0.51
Yb	0.10	1.06	1.13	0.35	0.41
Lu	0.03	0.13	0.16	0.03	0.04

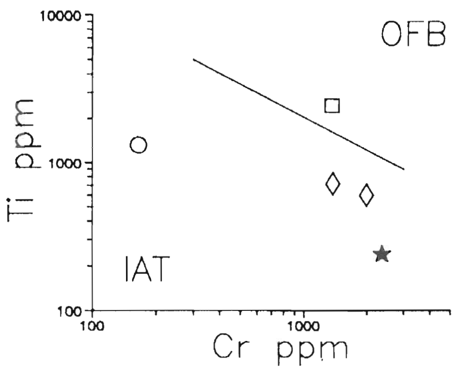

Fig. 3. Locations of the representative points in the Ti vs. Cr diagram. Symbols: square = eclogite, circle = amphibolite, double triangle = troctolite and pyroxenite, star = garnet peridotite. The eclogite plots in the field of the ocean floor basalts (OFB). IAT = island arc tholeiits.

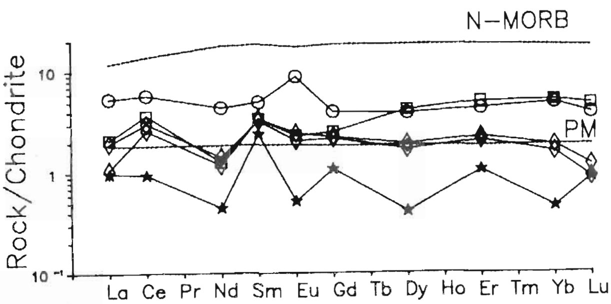

Fig. 4. REE patterns of Pohorje rocks and N-MORB (Hofmann, 1982), chondrite values after Nakamura (1974). Square = eclogite, circle = amphibolite, double triangle = troctolite and pyroxenite, star = =garnet peridotite, PM = P-MANTLE (Hofmann, 1988).

In particular, the garnet peridotite (column 1 in table 1) shows the lowest Al_2O_3 and the highest MgO contents. The spinel amphibolite (col. 2 in table l) is very rich in Al_2O_3, poor in Cr and displays a positive Eu anomaly (Fig. 4): plagioclase played an important role in the pre-metamorphic (magmatic) fractionation. Al_2O_3 content is moderate in the eclogite (col. 3 in table l), which plots in the Cr-Ti field of the ocean floor basalts (Fig. 3) and in the Zr/4-Y-2Nb field of the P-MOR basalts (Meschede, 1986); this rock is the only one in which Nb occurs in a detectable amount (i.e. >5 ppm). Finally, metagabbros display chemical features which are consistent with troctolite (col. 4 in table l) and pyroxenite (col. 5 in table l) protoliths, for which relatively high Sc and Co contents may suggest pyroxene (and perhaps olivine) cumulus.

As concerns rare earth elements, the value of ΣREE ranges between 2.76 and 15.81, being lower than 5xchondrite in every case. The ΣREE pattern (Fig. 4) of the spinel amphibolite shows a positive Eu anomaly, that of the garnet peridotite displays some irregularities. However, disregarding the garnet peridotite, the ΣREE patterns are similar to that of the N-MOR basalts, and intermediate between the latter one and that of the P-Mantle (Hofmann, 1988).

A comparison between the Pohorje metabasites and some Austridic eclogites and metabasites of the Eastern Alps is shown in Fig. 5 and 6, in which the patterns of the basalts and the P-Mantle are also shown for reference. The Sm anomaly of the garnet peridotite is to be pointed out (Fig. 5), as well as the contrasting behaviour of Sr (Fig. 6) perhaps related to a synmetamorphic chemical mobility. However, the similarities among the patterns of the Austridic rocks is very clear, as well as the analogies of all of them with the N-MORB pattern. These similarities may suggest a common origin of all these Austridic metabasites in the frame of a MOR magmatism.

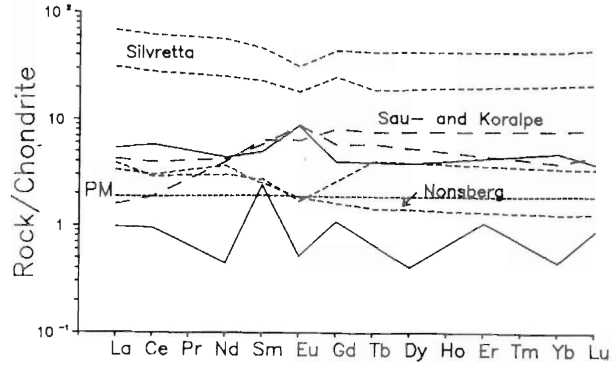

Fig. 5. Comparison between Pohorje rocks (solid lines), the Silvretta (Maggetti et al., 1987), Sau- and Koralpe (Miller et al., 1987) and Nonsberg metaultrabasites (Morten & Obata, 1990). PM = P-MANTLE.

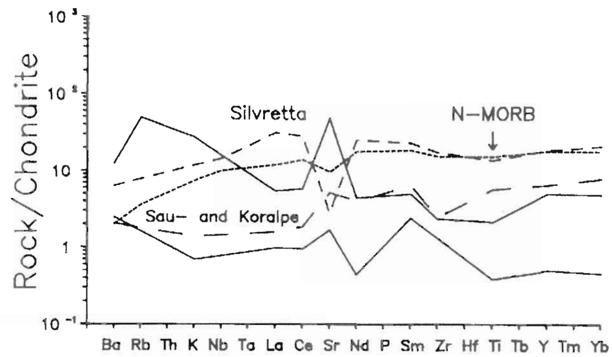

Fig. 6. Spider diagram (Thompson, 1982), symbols as in Fig. 5.

Petrography and petrology

The basic petrographic features of the major rock types are given below. Readers can find more data and details in Hinterlechner-Ravnik et al. (1991 a, b).

a) The *less altered eclogites* are essentially medium to coarse-grained, bimineralic rocks, consisting of omphacite and pyrope-almandine garnet. The grain size can be uneven within the same rock sample. Kyanite and zoisite are sometimes present. Very thin reaction rims occur commonly around almost all mineral grains. Garnet makes up the largest, euhedral crystals in most of these rocks. Omphacite occurs as prismatic, colourless or slightly greenish crystals, which are usually free of inclusions. Kyanite either makes up scattered idioblasts randomly oriented and distributed in the rocks or, in special cases, it is intimately associated with small omphacite crystals.

b) The common effects of the *eclogite alteration* are:

(i) symplectite rims around the omphacite crystals and symplectite aggregates and patches, within which corroded clinopyroxene and kyanite relics may occur, as well as zoisite idioblasts and poikiloblasts. These symplectite aggregates and rims can be referred to the reactions:

$$\text{Na - Cpx + Grt} \rightarrow \text{Cpx + Pl} \quad (1)$$

and

$$\text{Na - Cpx} \rightarrow \text{Cpx + Pl;}$$

(ii)plagioclase + corundum + spinel pseudomorphs; these consist of small prismatic crystals of An-rich plagioclase, the core of which is crowded by worms-like corundum crystals ± green spinel. These pseudomorphs can be related to the reactions:

$$\text{Ky + symplectite} \rightarrow \text{Pl+Cor} \quad (2)$$

$$\text{Cr - Ky + symplectite} \rightarrow \text{Pl + Cor + Spl} \quad (3)$$

The reactions Ky + Zo + Qtz → An and Ky + +Zo → An + Cor may also be mentioned;

(iii) zoisite idioblasts and poikiloblasts (in the special sample 142E);

(iv) green amphibole + plagioclase rims around the garnet crystals; the amphibole consists of Fe-pargasitic hornblende, barroisite and Mg-taramite (Leake, 1978);

(v) pargasitic-edenitic hornblende crystals.

The appearance of the above-mentioned amphiboles (rims and aggregates, respectively under points iv and v) can be referred to the reaction:

$$\text{Pl + Cpx + Grt} \rightarrow \text{Amph + Czo (or Zo)} \quad (4)$$

Among the above described products of the eclogite alteration, the symplectite and the Plag + + Cor + Spl pseudomorphs after kyanite certainly represent an early and dry stage of the post eclogitic evolution: their appearance, indeed, is directly related to the destabilization of the eclogitic mineral phases. A further later evolutional stage is represented by the appearance of amphibole and zoisite: these minerals, indeed, developed from the products of the early alteration stage due to access of water.

c) The *coronitic metatroctolites* are medium-grained rocks, in which the original granular panxenomorphic texture is still recognizable, nothwithstanding the important development of coronas. Disregarding the serpentinization, three crystallization stages can be clearly recognized.

The oldest stage is the magmatic crystallization. Olivine, pyroxene, plagioclase, brown spinel and magnetite are the mineral phases related to this stage. The crystals of all these minerals are commonly affected by coronitic alterations, which record a second crystallization stage. The magmatic pyroxene crystals commonly consist of Al-diopside containing exsolution lamellae of spinel (and sometimes corundum).

Along the olivine-plagioclase interfaces, both minerals are systematically surrounded (and sometimes replaced) by a zoned, polymineralic corona which consists of:

- an inner, orthopyroxene zone characterized by radiate texture;
- an outer, clinopyroxene zone characterized by radiate symplectitic texture; clinopyroxene displays small corundum exsolution lamellae and drops;

- an intermediate, thin zone of pale-green spinel, which systematically separates the two pyroxene-zones from each other.

The above described polymineralic corona is related to the reaction:

$$Ol + An \rightarrow Opx + Spl + Cpx. \quad (5)$$

According to Hinterlechner-Ravnik et al. (1991b), this alteration probably records the initial step of the burial of these rocks towards deeper levels. This interpretation is supported by the occurrence, within the magmatic plagioclase crystals, of numerous omphacite blebs (up to 59 % Jd, which is the highest Jd content found in all these rocks during the present research) and kyanite blebs. The omphacite inclusions indicate high-pressure induced destabilization of magmatic plagioclase, of the type:

$$Plag(An) \rightarrow Omp \quad (6)$$

A greenish amphibole (edenitic hornblende, sometimes actinolitic hornblende) also occurs locally, as a record of later, water-controlled, lower-grade alterations.

The metamorphic evolution of these rocks came to an end during a further, low-temperature, hydration stage, which was responsible of the extensive serpentinization.

d) The *garnet peridotites* are extensively serpentinized (antigorite). Only few relics of the original, probably magmatic, mineral assemblage survived; they are clinopyroxene crystals and roundish olivine crystals. Pyrope garnet phenocrysts, the shape of which is elongated parallel to the layering, could represent the product of a later, subsolidus reaction. At least two alteration stages, in addition to the serpentinization, are very clearly recorded.

The oldest alteration effects occur as multiphase reaction rims systematically located at the garnet-olivine interface. They have been referred by Hinterlechner-Ravnik et al. (1991b) to the reaction:

$$Ol + Grt \rightarrow Cpx + Opx + Spl \quad (7)$$

A later alteration stage is represented by: (i) the appearance of a greenish hornblende and gedrite + chlorite + talc at the clinopyroxene-garnet interface and within the orthopyroxene + clinopyroxene + spinel aggregates replacing garnet, and (ii) the very thin, highly refringent fibrous reaction rims occurring at the garnet-amphibole interface (too thin for being analysed).

e) The *garnet-pyroxenites*, which are here the most uncommon rock type, display a well developed coronitic texture. The original (magmatic?) clinopyroxene crystals have a dusty aspect and are brown in colour. They are frequently surrounded by a complex zoned corona, the inner zone of which consists of clinopyroxene and the outer zone of garnet. These coronas have been related by Hinterlechner-Ravnik et al. (1991b) to the destabilization of the original pyroxene according to the following reaction:

$$Cpx + Pl \rightarrow Grt + Cpx \quad (8)$$

The garnet zone in these coronas is rimmed and penetrated by a weakly coloured amphibole (edenitic or tremolitic hornblende). Colourless amphiboles also occur as relatively large nematoblasts which replace the old, dusty pyroxene crystals, and are either associated with the newly formed clinopyroxene of the coronas or included within the garnet zones of the coronas.

f) The *forsterite-tremolite (anthophyllite) schists* are well foliated metaultrabasites characterized by the abundance of tremolite and the occurrence of Mg-chlorite. The forsterite crystals are largely replaced by serpentine. The peculiarity of these rocks

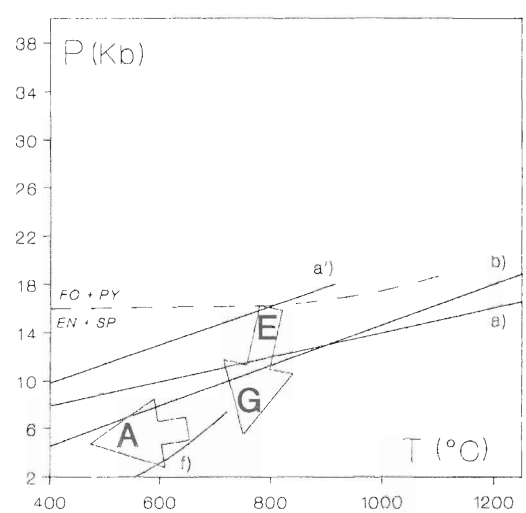

Fig. 7. Schematic P-- path of the Pohorje eclogites, from the eclogitic stage (E) to the symplectitic (G) and the epidote-amphibolitic (A) stages. a = Jd + Qtz (Jd = 16 %), á = Jd + Qtz (Jd = 37 %), b = Zo + Ky $\rightarrow$ Pl + H_2O (Nitsch, 1972), f = epidote out (Apted & Liou, 1983).

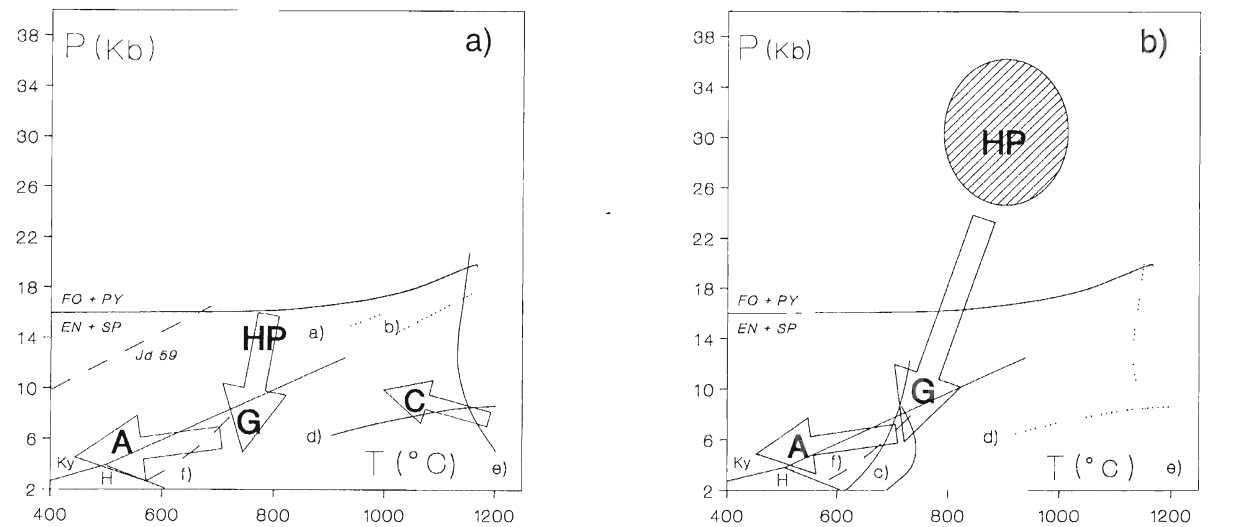

Fig. 8. Schematic P-T path of the Pohorje coronitic metatroctolites (Fig. 8a) and garnet peridotites (Fig. 8b), from the high-pressure (HP) stage to the granulitic (G) and the epidote-amphibolitic (A) stages. C represents the coronitic stage. a = Grt + Ol ➤ Pxs + Spl (Jenkins & Newton, 1979), b = Grt + Ol ➤ Pxs + Spl (MacGregor,1970), c = field of Amph-Anth-Chl peridotites (Jenkins, 1981), d = An + Fo ➤ Cpx + Opx + Spl (Obata, 1976), e = approximate location of the mantle peridotite melting curve (Wyllie, 1977), f = Epidote-out (Liou, 1973; Apted & Liou, 1983).

is therefore the substantial lack of anhydrous mineral alterations, and the abundance of hydrated mineral phases.

The P-T polymetamorphic path

The above presented textural and petrological data indicate a polymetamorphic history of the Pohorje eclogites and metaultrabasites. Taking into consideration the results obtained by Hinterlechner-Ravnik et al. (1991a, b), to whom readers are referred for more details, this polymetamorphic history is outlined below.

As concerns the Pohorje eclogites, three petrogenetic stages have been recognized in them:

1) an eclogite stage, during which the omphacite + garnet and omphacite + garnet + kyanite assemblages formed (+ zoisite + rutile + apatite); the jadeite content in the analysed omphacite crystals ranges between 16 and 37%;

2) a *symplectite stage*, during which the plagioclase + diopside (reaction #1) and plagioclase + corundum ± spinel (reactions #2, #3) assemblages formed;

3) an *epidote-amphibolite stage,* during which amphiboles and plagioclase + new zoisite (reaction #4) crystallised from the minerals of the symplectite stage.

The T-P conditions estimated by Hinterlechner-Ravnik et al. (1991a) for each of these stages are sketched in Fig. 7. Considering these geothermobarometric estimates, the symplectitic stage has been called "granulitic stage" by the above mentioned authors. It was also pointed out that the P-T conditions assigned to the granulite stage are similar to those estimated for the pre-Alpine main metamorphism in the surrounding metapelites (compare the location of G in Fig. 7 and Ml in Fig. 2);

As concerns the *Pohorje metaultrabasites,* the following crystallization stages are recorded:

1) a *magmatic stage,* represented by the following mineral assemblages: Ol + Pl + Cpx + Spl (±Opx); Ol + Cpx; and probably Cpx + Pl;

2) a *coronitic stage,* mainly recorded in the coronas, related to the above mentioned reaction #5;

3) a *high-pressure stage,* suggested by the above reaction # 6;

4) a *lower-temperature hydratation stage,* represented by the crystallization of amphiboles and chlorite;

5) *the serpentinization stage.*

The P-T conditions estimated by Hinterlechner-Ravnik et al. (1991b) are shown in Fig. 8a and b. The coronitic stage is related to cooling under isobaric or, better, increasing pressure conditions at about 8 kb (arrow C in Fig. 8a).

It is worthy to recall that the high-pressure stage (Hp in Fig. 8a) is recorded by the omphacite armoured relics included within the plagioclase crystals of the coronitic metatroctolites, and by the original stability of the garnet-olivine assemblage in the garnet peridotites. These findings indicate high pressure conditions (consistently with the above mentioned kyanite inclusions), the maximum values of which can be estimated on the basis of the omphacite composition (59% Jd).

The multiphase rims occurring in the garnet peridotites at the olivine-garnet interface (reaction #7) is reasonably related to a decompressional evolution, as sketched in Fig. 8b by means of the HP G path. For the coronitic metatroctolites, this path is partially speculative; however, it is consistent with the fact that clinopyroxenes in these rocks are commonly symplectitic Al-diopsides, omphacite occurring only locally as a relic phase.

A lower temperature, water-controlled, further evolution followed, producing mineral alterations which can be classified as *epidote-amphibolite stage*. Path (A) referred to this stage corresponds to the obvious mineral alterations which can be expected within the ambit of this physico-chemical context at shallow crustal levels.

An early cooling stage under probably increasing pressure conditions is only recorded in the coronitic metatroctolites (path C in Fig. 8a). If this important peculiarity is disregarded, the schematic polymetamorphic P-T path turns out to be similar in all the Pohorje metabasites and metaultrabasites (compare Figs. 2, 7 and 8), and records the main steps of their crustal history.

Concluding considerations

The above reported data and interpretations concerning the Pohorje eclogites and metaultrabasites suggest some further remarks on the history of these rocks and their implication in the regional geology of this part of the Austrides in the Eastern Alps.

1. The various types of the Pohorje eclogites and metaultrabasites are intimately associated in the field: in fact, besides the numerous, isolated, eclogitic lenses and boudins of relatively small size, a big "metaultramafic" body exists in which all above described rock types are intimately associated, in such a way that they cannot have experienced a significantly different metamorphic evolution.

2. The chemistry of all these rocks clearly indicates the subcrustal origin of all Pohorje metabasites and metaultrabasites. The patterns of REE display an affinity with N-MOR basalts, and strong similarities with similar Austridic rocks of the Eastern Alps. In the Pohorje area the eclogitic lenses and boundins may represent the metamorphic products of fragmented ocean floor basalts, while the major "metaultramafic" body may represent a partly dismembered, layered gabbro body. The whole complex within which these mantle-derived rocks occur may represent an old tectonic melange, deeply reworked by the Variscan and the Alpine metamorphisms.

3. The Pohorje eclogites and metaultrabasites record a complex crustal evolution consisting of three main metamorphic stages (Figs. 7 and 8a, b):

(i) a dry, high-pressure recrystallization, during which the protoliths altered into eclogites, and high-pressure mineral assemblages replaced the magmatic minerals in the ultrabasites, under high temperature conditions *(high-pressure stage);*

(ii) a later, lower pressure, metamorphic recrystallization, during which the high pressure mineral assemblages were altered under dry, lower-temperature conditions *(granulitic stage);*

(iii) a further, lower pressure/lower temperature metamorphic recrystallization under water controlled conditions *(epidote amphibolitic stage).*

4. From the geological point of view, the above three stages can be defined as follows:

(i) the older, high-pressure stage implies a progressive burial into deep and hot crustal levels, an early stage of which seems to be recorded in the coronitic stage of the metatroctolites (C in Fig. 8a);

(ii) the successive, granulitic stage corresponds to uprising towards shallower, but still hot and dry crustal levels (G in Figs. 7 and 8a, b);

(iii) the younger, epidote-amphibolitic stage (A in fig. 7 and 8) records a further uprising stage, during which the rocks reached shallow, cooler and water-bearing levels.

5. The age of the Pohorje eclogites and metaultrabasites is unknown. However, some unquestionable chronological constraints can be pointed out:

(i) the epidote-amphibolitic stage in the eclogites and in the metaultrabasites (A in Figs. 7 and 8) matches fairly well the P-T conditions of the Alpine metamorphic overprint recorded in the country-rocks (A in Fig. 2); but it also roughly fits the T-P conditions of the Variscan metamorphism (Purtscheller & Sassi, 1975; Bögel et al., 1979; Becker et al., 1988);

(ii) the P-T conditions of the pre-Alpine metamorphism recorded in the country-rocks (M_1 in Fig. 2) are consistent with those of the above outlined granulitic stage (G in Figs. 7 and 8);

(iii) bands and veins of well-foliated leucocratic gneisses occur within the large ultrabasic body, locally crosscutting its layering. They are identical to the above mentioned aplitic and pegmatitic pre-Alpine gneisses occurring as veins and bands within the surrounding metapelites.

These three facts consistently point to a pre-Alpine age for most of the history of the Pohorje eclogites and metaultrabasites. Whether this pre-Alpine history is significantly related to the "Caledonian" geodynamic event remains matter of speculation, which can be clarified only on the basis of radiometric geochronologic analyses of the Pohorje eclogites and metaultrabasites.

Acknowledgements. The present research was carried out in the frame of the IGCP Project No. 276, with the support of the "Centro di Studio per i Problemi dell' Orogeno delle Alpi Orientali" (CNR, Padova) and the "Geoloski Zavod Ljubljana". Financial support by Italian MURST (40%) is also acknowledged. Thanks are due to G. Franz (Berlin) who critically read an early version of the sections concerning petrography and metamorphic petrology and evolution.

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Stratigraphy of Silurian in the Korabi zone (Albania)

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Silurian sediments of the Korabi zone are mostly represented by little metamorphosed schists containing mainly graptolitic fauna.

The first data on the existence of Silurian to Devonian sediments of the Korabi mountains have been reported by Melo (1969). Later Crinoidea (Scyphocrinites) (Pinari, 1972; Quirici & Kodra, 1982; etc.) and also Graptolites (Nasi et al., 1973) have been found.

Results of prospecting and geological mapping of Albania in 1:200 000 scale but namely the stratigraphic (Xhomo et al., 1987) and biostratigraphic investigations (Pashko et al., 1988) contributed to the stratigraphic clarification of Paleozoic sediments in this zone.

Investigations of Graptolites in the Silurian beds of the Korabi zone allowed (differentiation of distinct) stages and also the majority of graptolitic zones. These new biostratigraphic results are of great importance for the entire Dinarides - Albanides - Hellenides.

The Korabi zone is divided into three units: the Kollovozi, Muhurri (Caje) and Mali i Korabit units. Silurian beds have great extension and thicknees in the Muhurri (Caje) unit. Stratigraphic studies are based on Graptolites. In the Kollovozi unit, the extension and thicknesses of Silurian beds are limited. The biostratigraphic subdivision is based on the investigations of Conodonts and Crinoidea, whereas in the Mali i Korabit unit, the black shales of Silurian age are devoid of distinquishable fossils but occur underneath the schist-carbonate formation of Lower Devoniau age (Xhomo et al., 1985; Meco, 1988).

Silurian in the Muhurri unit

Biostratigraphic investigations allowed to distinguish Lower and Upper Silurian beds. In general, the Lower Silurian sediments occur over aleuritic and quartzose shales with quartzite (Buzemadhe, Muhurr, etc.) or over sericite schists with lenses of silex. Silurian sedimentation was followed by black schists associated with limestone and containing Tentaculites (Bulac, Perroi i Miravecit, Katundi i Ri) as well as Devonian conodonts.

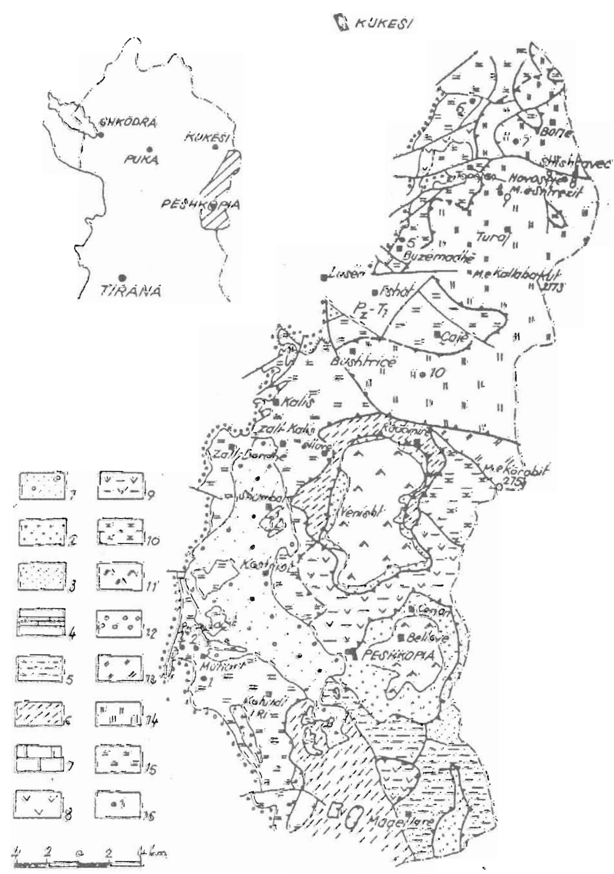

Fig. 1. Geographic and structural situation of the investigated sector. 1 - Pliocene to Quarternary, 2 - Maastrichtian flysch, 4 - Globotruncana limestones, 5 - Lower Cretaceous flysch, 6 - Tithonian - Cretaceous flysch, 7 - neritic Senonian limestone, 8 - volcano-sedimentary formation (Triassic ?), 10 - terrigenous facies with limestone (Triassic ?), 11 - evaporite, Permian to Triassic, 12 - conglomerate, Permian to Triassic, 13 - Paleozoic formations (Mali i Korabit unit), 14 - Paleozoic formations (Kollovozi unit), 15 - Paleozoic formations (Muhurri unit), 16 - situation of Silurian profiles.

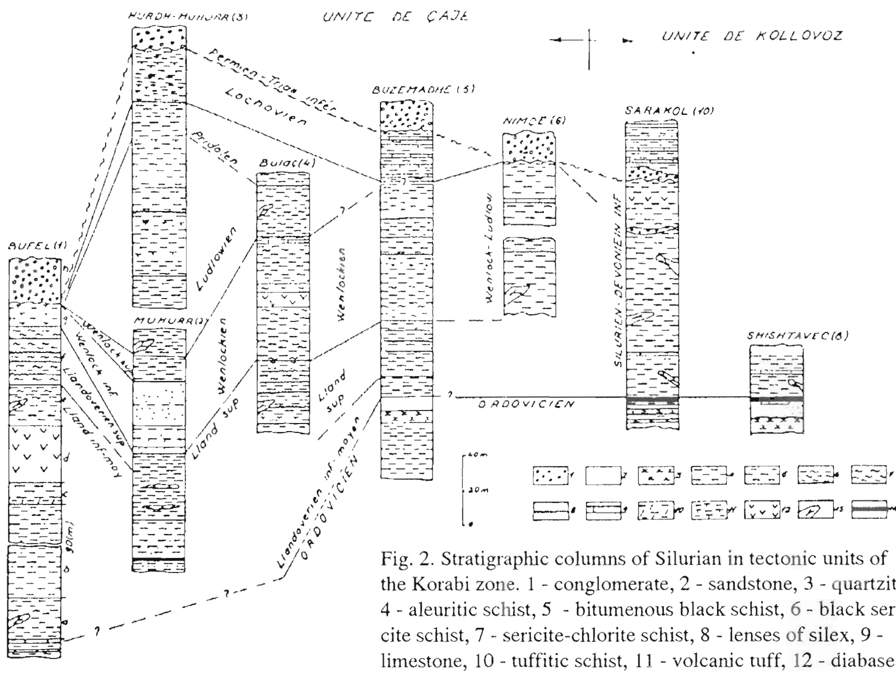

Fig. 2. Stratigraphic columns of Silurian in tectonic units of the Korabi zone. 1 - conglomerate, 2 - sandstone, 3 - quartzite, 4 - aleuritic schist, 5 - bitumenous black schist, 6 - black sericite schist, 7 - sericite-chlorite schist, 8 - lenses of silex, 9 - limestone, 10 - tuffitic schist, 11 - volcanic tuff, 12 - diabase, 13 - minette, 14 - iron ore.

Lower Silurian (Llandoverian - Wenlockian)

The Lower Silurian black schists (are extend) from Zogaj and Muhurri on the south (to) Nimce in the north. It is possible also to observe more detailed sections of these sediments at Buzemadhe, Bufel, Bulac and elsewhere (fig. 2).

At Buzemadhe it was possible to describe in detail the lowermost levels of Llandoverian by the means of Graptolites. Sediments of Lower Silurian age occur over sericite schists containing, at the top, a stratified and microfolded quartzite of probable Ordovician age. The profile is composed, from the base to the top of the following units:

- 13 - 14 m of fine-grained quartzitic sandstone passing into aleuritic shale,
- 120 m of black carbonaceous sericite schist with thin layers of silex (in the lower part), banks of sandstone and aleuritic schists.

In the first levels of schists over the sandstone the following Graptolites have been found: *Climacograptus (Monoclimacic)* sp., *C. gr. scalaris, Monograptus (Monoclimacis)* sp. (cf. *crenularis*) dating the Lower and Middle Llandoverian.

Higher, within the indicated Graptolites, Pashko recognized *Petalolithus (Petalolithus) elongatus, P. (P.)* cf. *tenuis, Monograptus (Monograptus) priodon priodon, M. (Pristiograptus) gr. nudus, M. (P.) nudus variabilis, M. (Stenograptus) crispus, M. (Spirograptus) turriculatus, M. (S.) turriculatus minor, M. (S.) proteus proteus, M. (S.) proteus curvus, M. (S.) spiralis contortus, M. (Demirastrites)* cf. *pragensis, Diversograptus capillaris,* an assemblage characterizing the Upper Llandoverian.

It is possible that the upper part of the schists where still no Graptolites have been found, represents the Wenlockian.

At Bufel a more complete profile is present containing from the base to the top:

a) 25 - 30 m thick sericite schist and aleurite shale of dark grey colour containing small lenses of biomicrosparitic limestone, lenses of silex as well as sandstone to conglomerate banks. In the upper part of schists badly preserved Graptolites have been discovered. Lamprophyre (minette) bodies are present amidst the sericite schists.

b) 90 m of aleuritic and arenaceous schists with clear rhythmicity.

c) 120 m of sericite schist with a horizon (4 m) of reddish limestone at the base.

d) 3 m of albitic diabase generally very chloritized and carbonatized.

f) 30 - 35 m of graphitic sericite schists alternating with thin layers of silex.

Graptolites are abundant and well preserved in the schists. Pashko distinguished there: *Climacograptus* sp. *(gr. scalaris), Monograptus (Monograptus)* cf. *lobiferus* characterizing Lower and Middle Llandoverian. Also the *Monograptus (Monograptus) priodon priodon, M. (M.)* cf. *veles, M. (M.)* cf. *sedwicki, M. (Pristiograptus) nudus, M. (Monoclimacis) griestoniensis, M. (Streptograptus) crispus, M. (Spirograptus) spiralis contortus, M. (S.) tullbergi tullbergi, M. (S.) tullbergi spiraloides, M. (S.), sp. (gr. planus), Rastrites* cf. *linnaei, Diversograptus ramosus, D.* cf. *capillaris, Retiolites (Retiolites) geinitzianus angustidens, R. (Pseudoplegmatograptus) longispinus* which are characterizing Upper Llandoverian.

g) 12 - 13 m of sericitic-chloritic and tuffitic schists in which the following Graptolites were discovered: *Monograptus (Pristiograptus) gr. dubius, M. (Monoclimacis) vomerina hemipristis, Cyrtograptus (Cyrtograptus) murchisoni, Retiolites (Retiolites) geinitzianus* dating Lower Wenlockian.

h) conglomerates of Permian to Lower Triassic age.

Lower Silurian beds of Llandoverian and Wenlockian age are in evidence in the Muhurri region. In black sericite schists with thin layers of silex, we discovered the following Graptolites: *Climacograptus sp., Monograptus (Monograptus) veles, M. (Pristiograptus) sp.* (cf. *nudus*), *M. (P.) dubius initialis, M. (P.)* cf. *prantli, M. (Monoclimacis) griestoniensis, M. (Spirograptus)* cf. *tullbergi, Diversograptus* sp., etc. which point to the Upper Llandoverian.

At Perroi i Bulacit, in a reversed sequence, the oldest stratigraphic levels are represented by carbonaceous black and in places tufitic sericite schists with thin silex layers. In this level we have collected the following Graptolites: *Monograptus (Monograptus) priodon priodon, M. (Pristiograptus) dubius initialis, M. (P.) tenuis, M. (Monoclimacis) vomerina, M. (Spirograptus) spiralis spiralis, Diversograptus ramosus, D. capillaris, Retiolites (Pseudoplectograptus) obesus, R. (P.) obesus hexagonalis, R. (P.) longispinus, R. (Stomatograptus) grandis,* which are characteristic of the Upper Llandoverian.

Higher, the following were found: *Monograptus (Monograptus) priodon praecedens, M. (Pristiograptus) gr. nudus, M. (P.) dubius initialis, M. (Monoclimacis) flumendosa, M. (M.),* sp. *(gr. vomerina), Retiolites (Retiolites) geinitzianus geinitzianus,* dating the Wenlockian.

Magmatic rocks represented by gabbrodiabese and lamprophyres were found in the superior part of Wenlockian schists.

Nimca is located in the extreme northern margin of the Lower Silurian beds with Graptolites. In this place one can observe the sericitic black schists intruded by gabbrodiabases and monzonites. These schists contain the Graptolites *Monograptus (Monograptus) gr. priodon, Cyrtograptus* sp., *Retiolites (Retiolites) gr. geinitzianus* which indicate Wenlockian age. Over the Wenlockian beds, after an open interval, the black schists continue with a layer of *Scyphocrinites* sp. containing limestone belonging to the Ludlowian.

The Upper Silurian (Ludlowian, Pridolian)

Sediments of the Upper Silurian are present in the Muhurri (Caje) unit at Hurd-Muhurr, Bulac, Nimce, etc.

In Hurd-Muhurr the sequence shows a series of black sericite schists with intercalations of silex and with lamrophyre (minette) bodies. A tuffitic admixture is interbedded in the upper part.

The lower boundary of the Ludlowian is marked by the appearance of *Graptolites Monograptus (Pristiograptus) nilsoni, M. (Prist.) bohemicus bohemicus, M. (Prist.) dubius, M. (Pernerograptus) colonus, Retiolites macilentus.* Higher, there is a large amount of *M. (Sactograptus) chimaerae.*

The pridolian creates the upper part of the sequence (together with the Lochkovian part) and it is recognized by the appearance of *M. formosus* and, higher on, of *M. perneri.*

At Bulac the Upper Silurian schists crop out in a very complicated tectonic situation. In the investigation of Graptolites, it was possible to distinguish:

- the Ludlowian in the lower part with *Monograptus nilssoni, M. (Prist.) bohemicus bohemicus, M. chimaerae, Retiolites macilatus* whereas. In the upper part, the *M. scanicus, M. chimaerae* etc. are present.

- the Pridolian constituted the uppermost part of the profile being defined *by Monograptus formosus, R. macilentus and Linograptus.*

Silurian beds are indicated by black sericite schists in the area of Lojme-Topojan, but Graptolites were not found there. A level (1.5 - 2.5 m) of biomicrosparitic compact limestone occurs in schists with numerous remnants of Crinoidea, namely of the *Scyphocrinites* genus. In limestones at Lojme, Crinoid *Cyclocharax fasciatus* has been discovered characteristic of the Upper Silurian and the Conodonts *Hindeodella equidentata, Ozarkodina denckmani* etc. which have broader stratigraphic value.

Silurian in the Kollovoz unit

Silurian sediments of the Kollovoz unit have limited extent. They occur in the Kollovoz, Shtrez, Sorokol etc. mountains and are of small thickness without Graptolites.

In the profile of Sorokoli Mount, over a level of iron ores of chlorite-oxidic type, the following levels have been identified:

- 25 - 30 m of bituminous sericite shists with bodies of lamprophyres (minette).
- 1.7 - 2.5 m of platy biomicrosparitic limestone containing remnants of Crinoidea, Ortoceratides and Conodonts.
- 90 m of sericite schists, in places aleuritics pierced by lamprophyre bodies and also by gabbrodiabases in the upper part.

Similar beds occur in the Kollovoz and Shtrez Mts. and are represented by schists with crinoidic limestones. The Crinoidea of the limestone beds are of the *Scyphocrinites* genus and revealed Conodonts of the apparatus *Kockellela variabilis* which is a typical form of the Upper Ludlowian. The summital part of the limestones yielded the Conodonts *Ozarkodina excavata excavata, O. remscheidensis remscheidensis* but also *Ancyrodelloides* cf. delta and other elements of *Ozarkodina confluens* which are testimony to a Lochkovian age.

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On the different transitions between preflysch and flysch sediments of Devonian in SW Bulgaria

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Abstract

In Kraishte (SW Bulgaria) there is a transition from moderately deepwater carbonate sediments, which passes through chert-clayey to flysch sediments, while southeast of Sofia the transition is from deepwater clayey deposits, passing through chert - to flysch deposits. The overlapping of the two types of sections of the turbidite facies D, C, E, A, F, B is a specific characteristic. The differences are related to the paleogeographic and paleogeodynamic conditions: the Kraishte section (I) illustrates a zone next to a front-arc, while the section (IV) northeast of Sofia illustrates the conditions in the proximity of high back-arc.

Differentiated Devonian sediments crop out along the southern slopes of the Western Balkan Mountains and in Kraishte (West Bulgaria). These sediments are mainly Upper Devonian (± Lower Carboniferous ?) flysch deposits (Spassov, 1973; Tenchov & Yanev, 1987; Yanev, 1991). The transition to the flysch sediments (s. s) is realized above different Lower-Middle Devonian basal sediments which are carbonate rocks in the southwest and clayey sediments in the northern part of the area considered. The transition itself also exhibits lateral particularities.

I. A level, about 60 cm thick, of extraclastic limestone (2), lies above more than 50 m of wave-like and banded limestones with marls and shale interbeds and several beds of lydites (1) in the southwest (e. g. in the Dubiche-Mureno section - Fig. 1-I). Then 14 m of greyish-beige-whitish shales with thin lydite interbeds follow (3), and with a transition, 26 m of thin light-coloured siliceous rocks (4). These are covered by 6 m of shales (5), 25 m of light-coloured banded cherts (6) and 2 m of thin-layered alternation of shales and lydites (7). The section continues with 37 m of yellowish-grey fine-bedded shales (8). An alternation, 30 m (9) lies above them, consisting of interbedding of shales and fine-grained quartzitized sandstones (beds - 1-2 up to 8-10 cm); 8 m (10) with sandstones (63 %); 12 m of thin-bedded alternation (11), within which shales slightly dominate in the cases of certain thicker sandstone layers (25 - 30 cm). Then comes a packet (12) of about 39 m of sandstones and strongly subordinate shales (with single transitions up to siltstones). The sandstones are medium- and thick-bedded, sometimes up to 2.8 - 6.5 m, of massive structure, with rare fine-pebble fragments and some characteristics of fluxoturbidites. The alternation of this interval bears some definite similarities with the deposits of hummocky type. The last interval prior the flysch s. s. is characterized by a packet alternation of the siltstone-shale laminites (12 % : 48 %) and about 40 % sandstones.

In the Tsarichina-Sivil section (Fig. 1-IV) above the greenishgrey shales with dark spots (1) and the shales containing fine siltstone admixture (2) 5 - 6.5 m of shales, siliceous shales and fine-silty shales start (3). Then come 33.5 m of bedded cherts with an admixture of phyllites and fine shale interbeds. At least two massive sandstones fluxoturbidite packets (19.5 and 6.5 m) lie above them, partially by interfingering. The paleotransport data indicate flows from NE to SW. The packets in interval 4 joint laterally with cherts (62.20 m) (Fig. 1-III). The next packet (5 - 9) of mudstones (74.25 m) is of a contourite fabric at varing ratios of the rocks: 5 - 5 m of alternating shales and siltstones (9:1); 6 - 19 m shales and cross-wavy siltstones (7:3); 7 - 11.8 m predominating shales and a smaller amount of siltstones; 8 - 27 m predominating shales in a rhythmic alternation with siltstones (10:1); 9 - 11.5 m shales with a very small amount of siltstones interbeds. Then comes a packet (10 - 13) containing two large sandstone fluxoturbidite packets (14 and 10 m) with a separating flysch-like

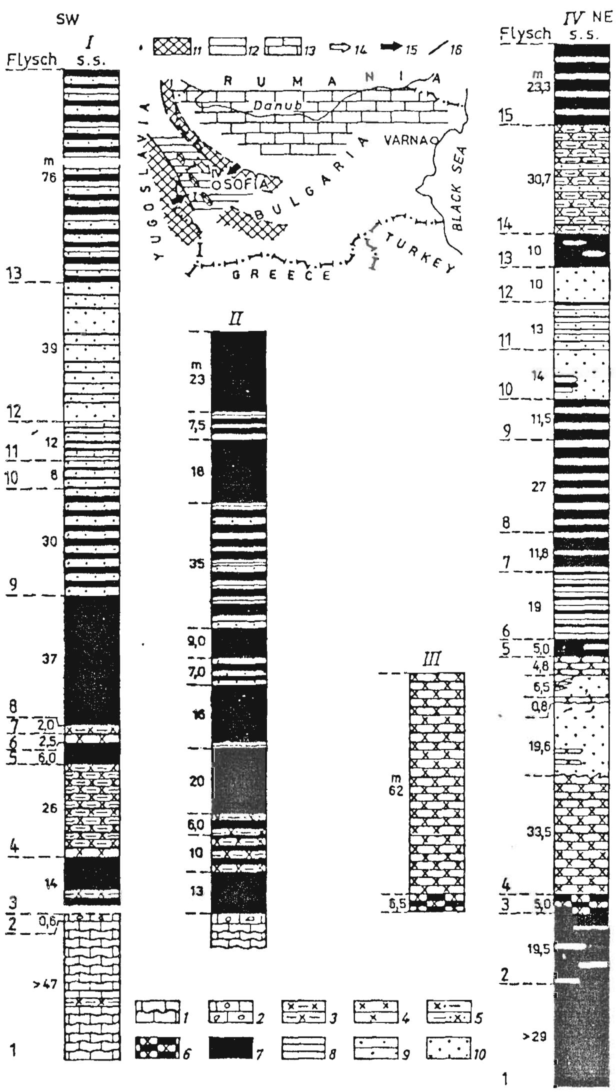

Fig. l. 1 - knotted limestones, "wellenkalck" - type limestones, 2 - extraclastic limestones, 3 - lydite, 4 - siliceous rocks, chert, 5 - siliceous rock, siltic, 6 - siliceous argilite, 7 - argilite, 8 - silstone, 9 - sandstone, 10 - sandstone massive (fluxoturbidite), 11 - dry land, 12 - hemipelagic sedimentation in intraarc trough, 13 - carbonate platform, 14 - longitudinal turbidity current, 15 - transversal turbidity current.

interval (rhythmic alternation of 60 % sandstones, 15 % siltstones, 25 % shales) and a cover (13) of 10 m of shales with lenses of siltstones (5 %). Lydites (74.7 %), sandstones (20.9 %), and shales (4.4 %) alternate in interval 14. The pre-flysch part of the section terminates with 23 m of laminated shales and silty shales (15), above which flysch s. s starts by a transition.

The sections studied, their rock and fossil peculiarities allow to perform facies reconstructions and in this way the restoration of the geologic evolution in concrete areas is possible. The following stages may be noted according to the intervals in column I for Kraishte: 1 - hemipelagic sedimentation in a moderately deep water area from a large marine basin; 2 - inception of differentiation of the basin, movements of different signs, in certain places up to washing out inclusive; 3 - deepening of the concrete area up to depths where the carbonates are unstable; 4 - the deepening is intensified, tectonic fractures occur, along which hydrothermal siliceous solutions penetrate - reprocesses on the background of a pelitic normal sedimentation; 5-8 - dominating pelitic sedimentation in a basin plain at a pulsatory supply of siliceous substance along deep fractures (5-7): 9-11 - sedimentation within the basin plain combined with distal parts of longitudinal turbidite flows - monotonous fan fringe turbidites, contourites, etc. (of facies E, D and C according to Mutti and Ricci-Lucchi, 1972); 12 - at the background of contrast movements during formation of an island-arc system and inter-arc trough the area belongs to an uplifted area - in the zone of slope accumulation near the shelf and in the previous band between the shelf and the basin plain; 13 - deepening of the area (within the basin plain), supply of clastics with distal turbidites by longitudinal flows and episodic transversal proximal flows (sandstones of facies B); 14 - formation of the system of basin plain, external, medium and internal fan and migrating channels (facies C and D, episodically A, F, sometimes B). In the Sofioter Balkan Mountains the following evolution is reconstructed (Fig. 1-IV): 1 and 2 - pelagic and hemipelagic clayey sedimentation in deep-water basin; late influence of slightly rising remote feeding land; 3 - inception of transformation of the area with tectonic fractures, rising hydrotherms and biogenic mobilization of siliceous substance; 4 - rearrangement of the sedimentary basin-inception of a rather contrasting discutegration - with unstabilized high island-arc and deept troughs at tectonic activity and increased seismicity. When the clastic accumulation zone is near, the environment favours the formation of fluxoturbidites; 5-9 - inclusion of the area in the sedimentary plain by a further decrease. The marginal parts of longitudinal turbidite flows reach here (facies E and D at the dominating facies

G); a tendency to a gradual recession of the zone of formation of the gravity flows; 10-13 - adjacent formation of an island-arc and a physicogeographic environment of a typical fan. The sandy sediments are related to the internal fan (in the meaning given by Mutti & Ricci-Lucchi, 1972), the other packets - with an interchannel area. The repetition of the fluxoturbidite is due to migration of the main channel; at the end of the stage - shifting of the fan; 14 - superimposed silica sedimentation over the basin plain related to solutions, occurrence of deeper fractures of the contact between the sedimentation area and the island-arc; 15 - hemipelagic pelitic sedimentation in a basin plain with an access of fine derivatives of grained material (facies G with turbidite layers and nepheloid layers).

The differences in the evolution of the two depositional areas are obvious. The role of the siliceous rocks and fluxoturbidites in the northern area is considerably greater. There, the movements were more contrasting and more intensive. The sequences studied exhibit the intermittent and uneven course of the processes of transformation of the formerly large deepwater basin included in the system of island-arcs and troughs. Judging by the paleogeographic and paleogeodynamic fabrics (Yanev, 1991) the Kraishte section (I) illustrates the processes occurring in the area adjacent to the frontal arc, while the Balkan Mountains section (IV) illustrates the typical evolution of a zone adjacent to a higher back-arc. The restored Upper Devonian reliefs are the first elements of the Variscan orogen, the final shaping of which occurred in the Balkan Peninsula during the collision stage of the area during Carboniferous and Permian.

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