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IGCP Project No. 276 Newsletter No. 2

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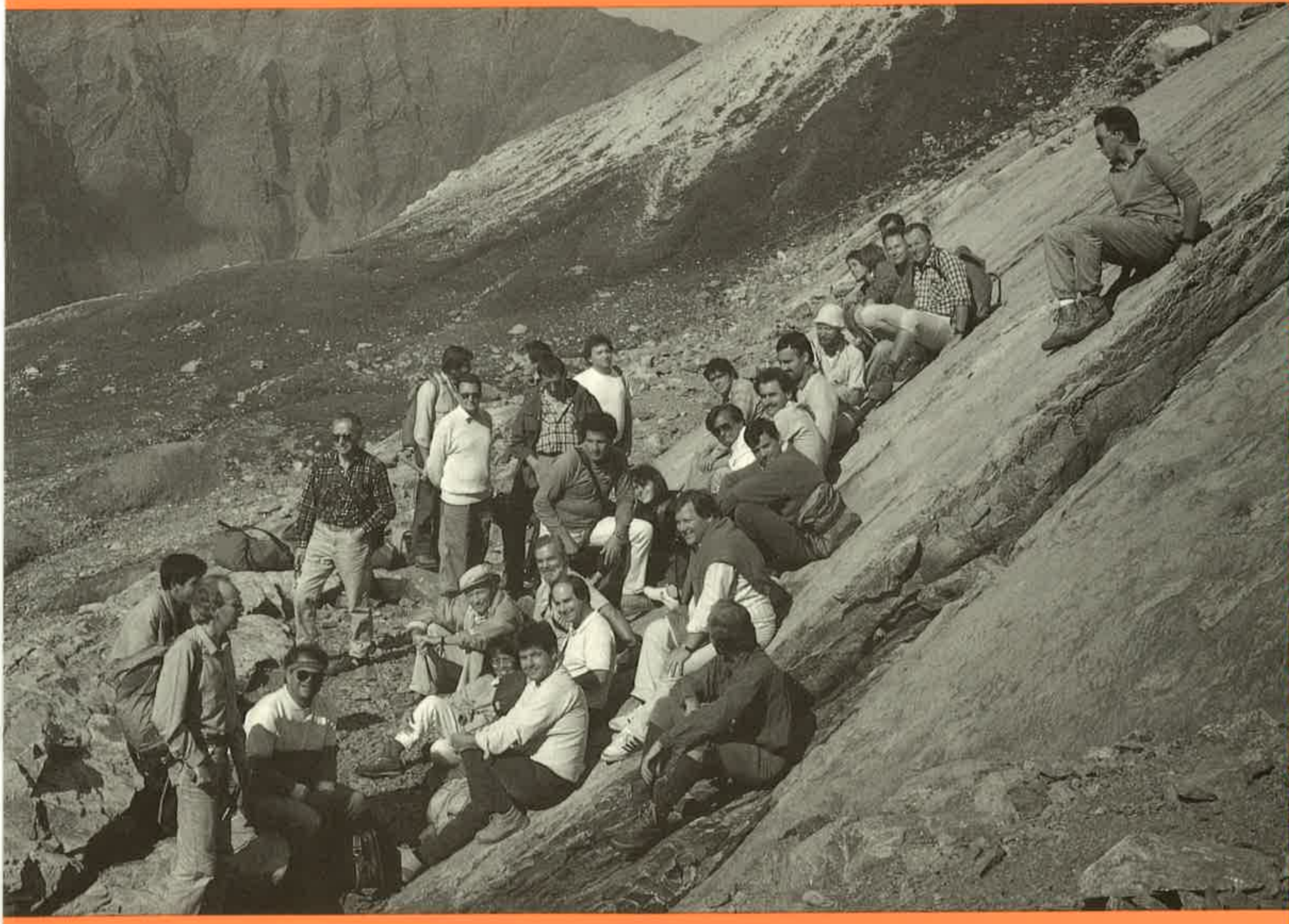

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TABLE DES MATIERES

AVANT-PROPOS

La 2ème réunion du projet PICG 276, a été organisée à Lausanne du 3 au 9 septembre 1989. Elle a rassemblé une soixantaine de participants venant de 19 pays. A cette occasion un cours d'introduction à la géologie des Alpes centrales et occidentales a été mis sur pied avec l'aide de nos collègues des Instituts de géologie et de minéralogie de Lausanne et de l'Institut de minéralogie de Fribourg, collègues qui se sont également dépensés sans compter pour diriger l'excursion de trois jours et demi dans les Alpes valaisannes. Un compte rendu de cette réunion a été publié dans Episodes (1990), vol. 12/2, p. 111-112.

Ce mémoire de géologie est une publication commune du Musée de géologie de Lausanne, des Instituts de Géologie et Minéralogie de l'UNIL et du Projet 276 du PICG dont il constitue la deuxième publication (Newsletter 2). Il présente une collection de travaux remis aux éditeurs à la suite de la réunion de Lausanne en 1989. Les derniers manuscripts ont été reçus en juin 1990, et les versions revisées en automne 1990.

Ces travaux couvrent un très large secteur géographique qui va de la chaîne des Andes jusqu'à la Mongolie intérieure en passant par le Rif, la Calabre, la Corse, les Alpes orientales, les Carpathes, les Balkans, les Taurides et les Pontides, les chaînes arabiques et le Caucase.

Toujours dans l'optique du projet 276 du PICG, les divers domaines de recherches de Sciences de la terre sont abordées dans ce mémoire:

- les problèmes d'évolution géodynamique et tectonique par Belov, Benmakhlouf & al., Minzoni, Sengor et Shao;
- des événements et phénomènes sédimentaires par Ebner et al. et par Ianev;
- des problèmes de métamorphisme par Arkai et par Rudakov, de plutonisme par Bonin, de volcanisme par Schlaegel-Blaut & al;
- des problèmes de corrélations et de reconstructions entre Andes et Téthys par Baldis & al..

En conclusion, nous ne pouvons que souligner la grande vitalité du projet IGCP 276, vitalité due en grande partie à son animateur, le Prof. D. Papanikolaou de l'université d'Athènes. Mes collègues P. Thélin et G. Stampfli ont partagés avec moi les tâches de revision et d'édition de ce mémoire et nous ne saurions terminer cet avant-propos sans remercier les Institutions qui par leur appui financier en ont permis l'impression: l'Académie Suisse des Sciences Naturelles, le Musée de Géologie de Lausanne et les Instituts de Géologie et de Minéralogie de l'UNIL.

Nous remercions également A. Jeanrenaud qui a fait la saisie informatique des textes et E. Thibault qui a entré toutes les corrections et préparé la mise en page.

Pour le comité d'édition: A.Baud

ALPINE REGIONAL METAMORPHISM OF THE DIFFERENT TECTONIC DOMAINS IN THE HUNGARIAN PART OF THE PANNONIAN BASIN

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Abstract

Although the Alpine tectonic of the domains ("microplates") forming the Carpatho-Pannonian realm has been extensively studied for many decades, hardly any attention has been paid to the Alpine regional metamorphism in the given area.
Considering the presence and characteristics of the Alpine regional metamorphism, the following types can be distinguished from NW to SE:
1) mostly retrogressive, low- to medium-grade, medium thermal gradient overprint in the Lower Austroalpine Nappe (Sopron Mts.);
2) plurifacial (polyphase) blueschist/greenschist facies metamorphism of the Mesozoic of the Penninic Window (Kőszeg-Rechnitz Mts.);
3) low (250^{o}C) temperature event overprinting the Hercynian high thermal gradient, low-grade Paleozoic of the Upper Austroalpine Nappe (Little Plain Graz Paleozoic);
4) no signs of Alpine metamorpnism (Transdanubian Midmountains Subunits);
5) variable (low to intermediate or intermediate to high thermal gradient prograde regional alteration ranging from the diagenesis through the anchizone up to the glaucophanitic greenschist or greenschist facies (central Hungarian, Bükk and Aggtelek-Rudabanya Subunits);
6) mostly retrograde and cataclastic, low-temperature metamorphism in certain parts of the pre-Hercynian (?) - Hercynian polymetamorphic basement (Tisza Unit). In certain zones intermediate thermal gradient, anchi-, epizonal Late Paleozoic - Mesozoic sequences were overthrusted into this relatively stable unit.
The pre-Hercynian(?) and Hercynian consolidation of the different fragments seems to be only one (not decisive) factor among those determining the regional distribution of pressure - temperature - time relations of the Alpine metamorphism in question.

1. INTRODUCTION

Due to the Alpine horizontal displacements different crustal blocks originated from various parts of the Tethyan realm, and thus having no geologic (genetic) connections, got in close geographic relations contacting each other along main tectonic lineaments in the Pannonian Basin. Although the important stratigraphic and tectonic features as well as the approximative original positions ("affinities") of these blocks are relatively well known, the details of their complicated movements are poorly understood, however numerous facial, paleogeographic and paleotectonic models have been elaborate recently to explain the present situation (KOVACS 1982, KAZMER & KOVACS 1985, BALLA 1988, HAAS 1987, etc.).
The aim of the present paper is to compare the characteristics of the Alpine and pre-Alpine (first of all Hercynian) regional metamorphic events which affected the crustal blocks forming the Hungarian part of the Pannonian Basin. With this comparison we intend to clarify whether there exists any correlation (and if any, what kind of) between the pre-Alpine metamorphic, magmatic consolidation and the intensity (grade) of the Alpine regional metamorphism of the blocks given. In the present state of investigations it was not possible to make an extensive regional correlation, for - unfortunately - hardly

any data exist on the eventual Alpine metamorphism of the neighbouring areas of these areas has been studied intensively for many decades.
These shortcomings can be explained mainly by methodologic causes: the Alpine metamorphism might have been of low-temperature character in general in the given areas, and in addition to the classical metamorphic petrologic (mineral paragenetic, microstructural) methods, the application of other (illite crystallinity, white mica b_0 geobarometric, vitrinite reflectance) methods is also needed to the up-to-date evaluation of incipient metamorphism. The following short review is based on the systematic research using all of these as well as other (geothermometric and geobarometric) methods.

2. GEOLOGIC AND TECTONIC OUTLINES

Fig. 1 demonstrates the main tectonic units (blocks or domains) and their separating borders (lineaments characterized mainly by horizontal displacements) of the pre-Tertiary basement of Hungary, compiled after FÜLÖP, DANK & al. (1987) with strong simplifications. From NW to SE the following tectonic domains can be distinguished.
The crystalline rocks of the Sopron Mts. belonging to the Lower Austroalpline Nappe System is separated by the so called Répce Line from the basement of the Little Plain correlated with the Graz Paleozoic forming a part of the Upper Austroalpine Nappe System.
The Transdanubian Midmountains Subunit (Bakony-Drauzug) showing Austro- and South Alpine affinities is located between the Raba and Balaton Lineaments. According to the recent interpretations (see KOVACS 1982) the Central Hungarian (or Igal) Subunit located between the Balaton and the Central Hungarian (Zagreb-Hernad) Lineaments represents a complicated tectonic zone, along which Bükk and the South Gemeric Subunits moved horizontally from the Internal Dinarides to their present positions, now forming the innermost parts of the Western Carpathians.

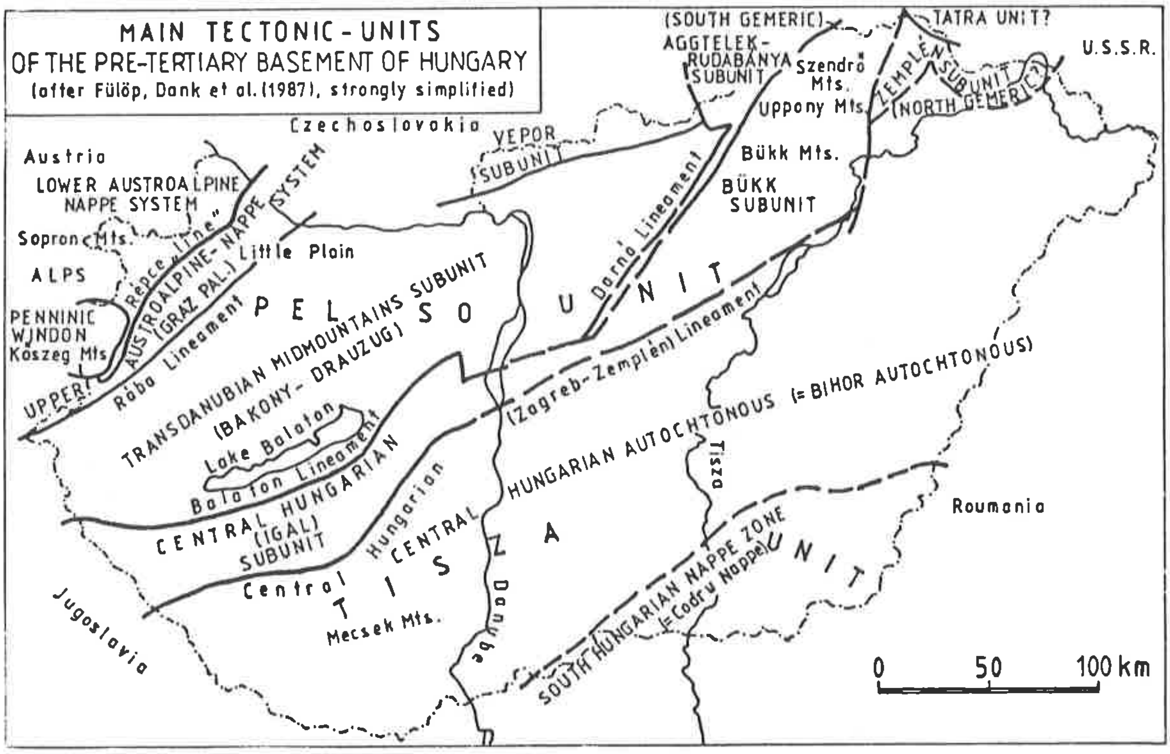

Fig. 1: Main tectonic units of the pre-Tertiary basement of Hungary (after Fülöp, Dank et al. 1987, strongly simplified)

South of the Central Hungarian Lineament the Tisza Unit (Tisia "microplate") can be found. While the Pelso Unit ö (containing the Transdanubian Midmountains, Central Hungarian, Bükk and South Gemeric Subunits) shows clear South Alpine and Dinaric affinities, the Tisza Unit (which can be found south of the Pelso Unit at present) is interpreted as a part of the nothern (European) border of the Tethyan realm.
In the northern and northeastern parts of Hungary the soutern ends of the Vepor and Tatra Units are to be found.

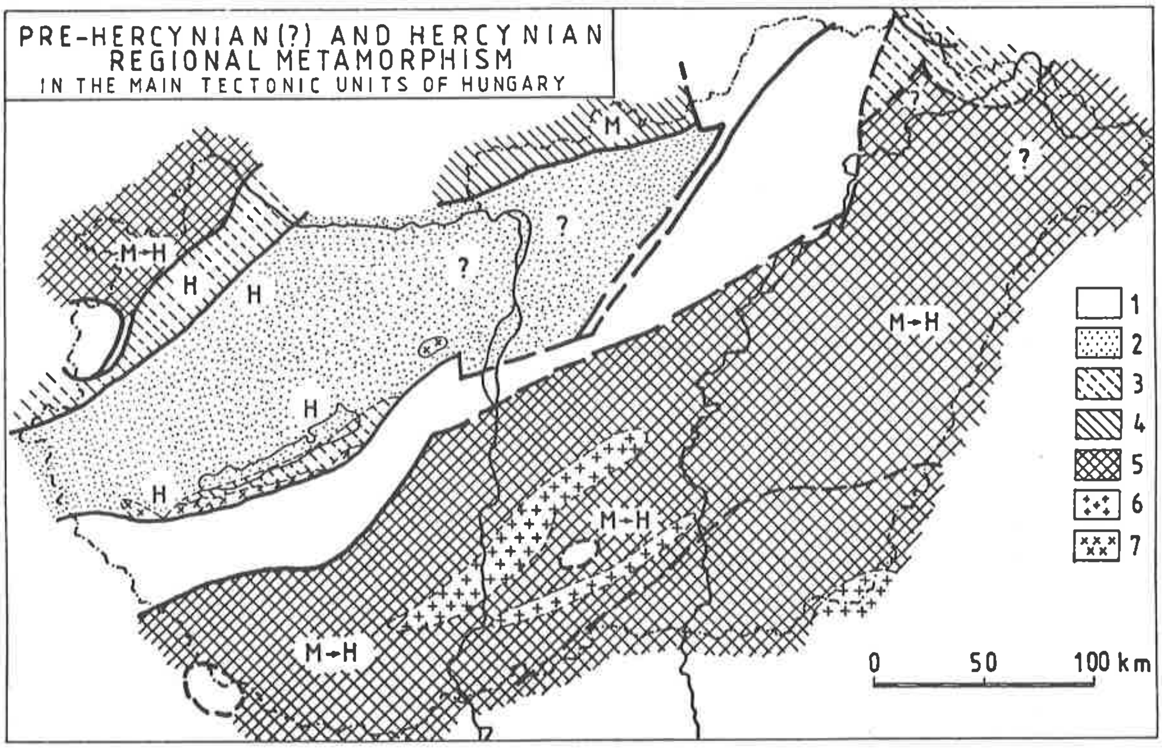

Fig. 2: Pre-Hercynian and Hercynian regional metamorphism in the main tectonic units of Hungary. Legend: 1 - non-metamorphic, 2 - sub-greenschist (very low-grade) facies including zeolite, prehnite-pumpellyite, pumpellyite-actinolite and prehnite-actinolite facies, 3 - grenschist facies, 4 - amphibolite facies (medium thermal gradient), 5 - medium thermal gradient amphibolite facies overprinted by high thermal gradient event, 6 - syn-kinematic granitoid magmatism, 7 - post-kinematic granitoid magmatism

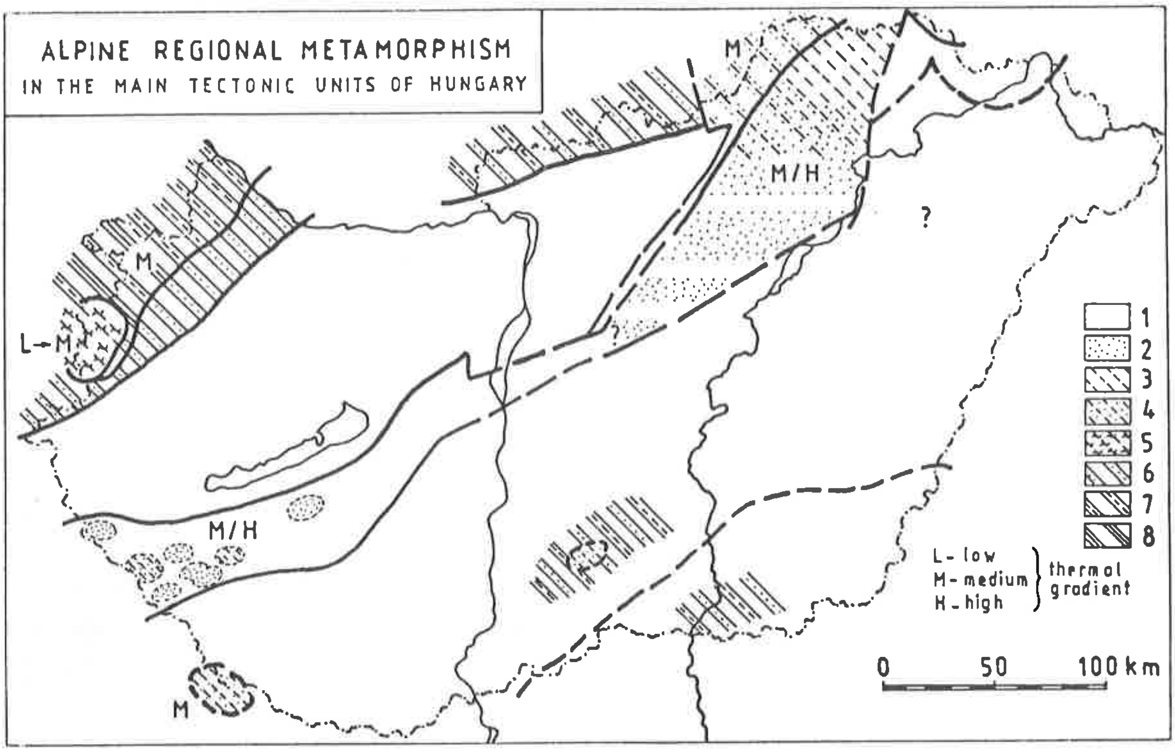

Fig. 3: Alpine regional metamorphism in the main tectonic units of Hungary. Legend: 1 - non-metamorphic, 2 - sub-greenschist facies, 3 - greenschist facies, 4 - transitional sub-greenschist - greenschist facies, 5 - polyphase metamorphism: glaucophane schist metamorphism overprinted by a greenschist facies event, 6, 7 & 8 - Alpine overprints (polymetamorphism sensu stricto) in the pre-Alpine metamorphic basement: 6 - overprint in sub-greenschist, 7 - in greenschist, 8 - in medium pressure amphibolite facies

3. COMPARISON OF ALPINE AND PRE-ALPINE REGIONAL METAMORPHIC EVENTS

On the next pair of figures (figs 2 and 3) we compare the main features of the Alpine and pre-Alpine (first of all Hercynian, partly perhaps pre-Hercynian) regional metamorphic events. The syn- and post-kinematic granitoids are also depicted. The transitional features (e.g. transitional sub-greenschist - greenschist facies), or the polyphase or plurifacial character (within a single tectonocycle) or the comparing the polymetamorphism sensu stricto (developing in different tectonocycles) can also be illustrated with various combinations of symbols.
Two maps striking contrast can be observed between the spatial distribution and intensity (grade) of the Alpine and pre-Alpine regional metamorphic events. The Alpine metamorphism proved to be significant especially in domains lacking pre-Alpine metamorphism.

(I) In Fig. 2 the empty parts of the map indicate those units where the signs of the pre-Alpine metamorphism are not known. These areas can be subdivised into the following types:
(IA) zones of Mesozoic (and oceanization) built up naturally only by Mesozoic (and perhaps younger) formations. This type can be found in Western Hungary in the Kőszeg-Rechnitz Mts. belonging to the Penninic Unit (Window). According to KOLLER (1978, 1985) and LELKES-FELVARI (1982) the Eo-Alpine, low thermal gradient glaucophane schist facies event was overprinted by a medium thermal gradient greenschist facies in the Meso-Alpine phase (showing an example of polyphase or plurifacial metamorphism).
(IB) An other type of the domains without any signs of pre-Alpine regional metamorphism is characterized by rather thick Paleozoic and Mesozoic sequences. In these units the increasing mobility of the sedimentary basins braking up carbonate platforms, the sinking of the basins, the formation of flysch-like sediments, and certain epirogenetic movements indicate the effects of the hercynian cycle. Hercynian regional metamorphism could not be proved (or more precisely: could not be distinguished from the Alpine one). These facts suggest that if there was any recrystallization during the Hercynian cycle, it must have been of burial type, and its temperature should not have exceeded the grade of the Alpine (Cretaceous) metamorphism ranging from very low-grade. In certain cases (aborted) rifting in the Triassic and Jurassic, subduction in the Upper Jurassic and Lower Cretaceous, and later on, collision type regional metamorphism are the main features of the Mesozoic evolution in these domains.
This type of domains can be further subdivided:
(IB1) in the first group the basement of the Paleozoic-Mesozoic sequences is unknowm. Examples of areas are the South Gemeric (Aggtelek-Rudabanya) and the Bükk Subunits.
In the South Gemeric (Aggtelek-Rudabanya) Subunit the grade of the Eo-Alpine regional transformatio on the Permian - Triassic - Jurassic sequences varies from the diagenesis through the anchizone up to the chlorite zone of the greenschist facies (or in certain part of the Meliata Series, South Gemer up to the glaucophanitic greenschist facies), depending on the paleotectonic position. Medium or transitional medium/low thermal gradients are characteristic for this realm (ARKAI & KOVACS 1986).
In the Bükk Subunit (Bükk, Uppony and Szendrö Mts.) also Eo-Alpine regional metamorphism was proved (ARKAI 1973, 1983). The grades of transformation change from the diagenetic zone through the anchizone up to the biotite isograd of the greenschist facies, the metamorphic thermal gradients vary from high to intermediate, reaching the medium value only locally.
Similar (partly diagenetic, mostly anchizonal, rarely epizonal (= greenschist facies chlorite zone), usually medium thermal gradient metamorphism was proved recently in the Late Paleozoic and Triassic of the Central Hungarian (Igal) Subunit, too (ARKAI 1988). In addition to the Eo- Alpine (Cretaceous) metamorphism, traces of Meso-Alpine thermal event can also be demonstrated (ARKAI & BALOGH, in preparation).
(IB2) Considering an other type of tectonic domains, the Alpine prograde metamorphic, Mesozoic rocks are found in tectonic contact with the pre-Hercynian(?) - Hercynian basement, having been mostly overthrusted on it by young (Tertiary) movements. The anchi-, epizonal, medium thermal gradient Mesozoic of the Barcs-West area (Tisza Unit, Drava Basin, SW-Hungary) seems to be an excellent example of this type (ARKAI 1989). The anchi-, epizonal formations in the Danube-Tisza Interfluve may represent similar conditions, but the paleotectonic connections in this case are totally uncertain.
Considering the stratigraphic, paleogeographic and other characteristics, all of these subunits mentioned above in group (IB) are genetically closesy related to the internal Dinarides.
Their present positions are interpreted by horizontal block (or "microplate") movements acting mostlsy during the Meso-Alpine phases (KOVACS 1982, BALLA 1988, HAAS 1987, KAZMER & KOVACS 1985).

(II) The second main group of the tectonic domains comprises those units where the Early Paleozoic basement was affected by very low- to low-grade, high thermal gradient

Hercynian regional metamorphism. Considering the Alpine Metamorphic evolution of these areas, two types can be distinguished:
(IIA) In one case (Little Plain basement between the Répce Line and the Raba Lineament) Mesozoic formations are unknown. The Hercynian, high thermal gradient, mostly greenschist facies (chlorite zone) metamorphism of the Paleozoic (Silurian? - Denovian?) formations was overprinted by an Eo-Alpine ((Cretaceous), very low-grade (ca.<250^{O}C) thermal event, similary to the Graz Paleozoic of the Upper East Alpine Nappe System (FLÜGEL & al. 1980, FRANK 1983, FRANK & al. 1987). In addition to the former lithostratigraphic (BALAZS 1975, FÜLÖP 1980) and metamorphic petrological (ARKAI & al. 1987) evidence, the new isotope geochronological data obtained on white mica fractions by K-Ar method (ARKAI & BALOGH 1989) also support the correlation of the Little Plain basement to the Graz Paleozoic.
(IIB) In case of the other type (in the Transdanubian Midmountains or Bakony - Drauzug Subunit) post-orogenic erosional unconformity separates the mostly anchizonal Early Paleozoic basement from the overlying non-metamorphic Late Paleozoic and Mesozoic. The grade of the high thermal gradient Hercynian regional metamorphism is generally anchizonal (sub-greenschist facies, see LELKES-FELVARI 1978, 1987, in the later case in ARKAI & LELKES-FELVARI 1987), but in the viccinity of the Balaton Lineament (towards SW) the greenschist and the amphibiolite facies (staurolite isograd) were also reached (Fig.2, ARKAI 1987a). Along the southern border of the unit (Balaton Lineament) the Hercynian post-kinematic, collisional type granitoid intrusions caused contact metamorphic overprints. The non-metamorphic (only medium-diagenetic) character of the Alpine structural level can be explained by the horizontal movement of the Bakony - Drauzug Subunit, thus escaping eastwards the main compressional zones of the Eastern Alps ("continental escape", see KAZMER & KOVACS 1985).

(III) The third group of the tectonic domains contains those units, the basement of which is characterized by strong, amphibolite facies, generally polymetamorphic Hercynian - pre-Hercynian(?) recrystallizations, and widespread syn- and sporadically late-kinematic Hercynian granitoid magmatism (mainly involving the Tisza Unit, the Vepor Subunit and the Sopron Mountains). These units may be considered as the "most consolidated" fragments of the Pannonian Basin. The generalized stages of their metamorphic evolution are as follows: Early Hercynian (eventually pre-Hercynian(?)) medium thermal gradient, amphibolite facies metamorphism followed by a Hercynian high thermal gradient, spatially strongly varying, partly prograde (anphibolite facies), generally retrograde, locally cataclastic overprint. In the thermal centres or axes of this event syn- and late-kinematic granitoid bodies were formed (for more detailed characterization of the metamorphic evolution see LELKES-FELVARI & SASSI 1981, SZEDERKENYI 1984, ARKAI 1984, 1987b, ARKAI & al. 1985, CSEREPES-MESZENA 1986, NUSSZER 1986, SZILI-GYEMANT 1986, and for that of the granitic magmatism see BUDA 1985). In these units the Mesozoic and late Paleozoic formations are either lacking (in the Sopron Mts. and the Hungarian part of the Vepor Unit, N-Hungary) or are non-metamorphic (Tisza Unit). Alpine progressive metamorphism has not been prouved so far. The metamorphic effects of the Alpine cycle were restricted to the rocks of the Hercynian - pre-Hercynian(?) basement.
(IIIA) The metamorphic sequences of the Sopron Mts. belonging to the Lower Austroalpine Nappe System show the signs of Eo-Alpine overprint characterized by medium thermal gradient, generally greenschist, maximally low-temperature amphibolite facies (LELKES-FELVARI & al., 1984).
(IIIB) In the southern part of the Vepor Unit Alpine (Cretaceous) retrograde (greenschist and sub-grenschist facies), partly cataclastic (mylonitic) overprint can be extrapolated based on the work of CAMBEL & KORIKOVSKY (1986).
(IIIC) In the Hercynian - pre-Hercynian(?) basement of the Tisza Unit ("Tisia microplate") traces of retrograde Alpine overprint are also known, especially in the central part of the
Danube-Tisza Interfluve (Kiskunhalas, Szank, Jaszszentlaszlo) as well as in the southern part of the Great Plain (Ferencszallas, Deszk, etc.). Locally, the retrograde overprint

was preceeded by mylonite formation. The temperature of recrystallization must have been low (sub-greenschist and greenschist facies). Uncertain data may indicate medium thermal gradient amphibolite facies recrystallization of possible Alpine age in the southern areas (SZÉDERKENYI, 1984). The effects of the Hercynian and eventual Alpine retrograde alterations have not been separated systematically so far: a detailed mineral paragenetic and isotope geochronologic study of this question is needed. According to our present (mostly hypothetic) opinion, the Alpine metamorphism in the Tisza Unit might not be widespread and significant: it might be restricted to the close environments of the main Alpine thrust zones.

4. CONCLUSIONS

Considering the different tectonic domains of the Hungarian part of the Pannonian Basin, it can be stated that there is only a very loose connection (negative correlation) between the pre-Alpine metamorphic, magmatic consolidation and the Alpine regional metamorphism.

Alpine, progressive regional metamorphism could be proved in the sequences deposited in the basins formed by Mesozoic rifting or by mobilization of Paleozoic and Mesozoic epicontinental platform and shelf areas which had not been affected by Hercynian regional metamorphism. The Alpine Progressive metamorphism of the Late Paleozoic and Mesozoic formations reached maximally the biotite isograd (ca. 400 - 450^{o}C) of the grenschist facies at the present erosional level. Its thermal gradient varied between medium and high (in certain zones intermediate: medium-low) values.

There is no clear, unequivocal correlation between the grade of the Hercynian - pre-Hercynian(?) metamorphism and that of the Alpine overprint. In certain terrains characterized by polymetamorphic amphibolite facies conditions, the temperature of the Alpine overprint might reach even the low temperature part of the amphibolite facies (Sopron Mts., Lower Austroalpine Nappe System). On the other hand, certain hardly consolidated, anchizonal terrains (e.g. the Transdanubian Midmountains Subunit) escaped the Alpine metamorphic effects.

Not as much the metamorphic-magmatic consolidation, but the spatial relation of the given tectonic domain to the Alpine tectonometamorphic zones seems to be the main factor controlling the regional distribution and grade of Alpine metamorphism. The negative correlation between pre-Alpine consolidation and Alpine Metamorphism may be of indirect type, in so far as the subduction- and/or collision-type structures generating regional metamorphism, as well as the rift zones connected to ocean floor and later on to burial metamorphism might have been preferentially formed in the least resistant parts of the lithosphere. On the other hand, however, the tectonometamorphic recycling of the older, more or less consolidated crustal fragmnents in the Alpine belt has also been a common, but of course, not everywhere valid phenomenon, which disturbed the theoretical correlation mentioned above.

Future investigations should aim the determination of exact ages (and reconstructed positions) of the Alpine regional events, the grades (temperatures) and thermal gradients of which are more or less known in the domains. Detailed metamorphic petrological and isotope geochronologic data are needed also from the surrounding areas. Thus, we hope, we may promote the interpretation of the Tethyan realm, completing the paleofacial, paleogeographic reconstructions with metamorphic reconstruction, which might offer a more complete picture on later stages of evolution.

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PROBLEMS IN PERI-GONDWANA LOWER PALEOZOIC RECONSTRUCTIONS BETWEEN ANDES AND TETHYS

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Abstract

The platform outline of Gondwana from the Andes to the Paleo-Tethys displays several coincident characteristics that lead to this first attempt for a Lower Paleozoic platform sedimentation, intra-Ordovician unconformities, molasse basins location, and several associated biostratigraphical constrains. The early Paleozoic connection between these two regions through the Iapetus and the western sequences and their geodynamic considerations. A shallow platform, with different facies developments seems to be a dominant feature for the peri-Gondwana lower Paleozoic domains. Also, a consistant, radially-distributed, clastic supply is interpreted for many peripheral-located terrigenous sequences along the Gondwana shield borders.
The Paleo-Tethys reconstruction theories strongly relies on the understanding of the NW African sequence correlation, and the integrated Gondwanaland megacontinental structure developments and geotectonic control.

1. INTRODUCTION

The study of the Lower Paleozoic peri-Gondwana platforms have not been developed intensively enough to establish its continental extent and to detect similar events in the Andean and the Paleo-Tethys region. This is a first attempt to do so, and is the first contribution from IGCP Project 276 in order to achieve a sequence of events from where to start building up some more detailed research.

2. PERI-GONDWANA RECONSTRUCTIONS

The existing literature has based the Lower Paleozoic correlation problems between the Mediterranean and western South American regions on two separate assumptions:
- The relationship among Southern Europe and Northern Africa (including NW Africa) and terranes of similar ages in Laurentia; linking them to the concept of the Iapetus Sea.
- Several correlation attempts between the western South American platforms and the Appalachian region eventually extending the hypothesis to some European fragments.
Both tendencies accept the link through Iapetus, but there is not a well established definition for the western peri-Gondwana platforms.
Three main integrations sketchs are presented, leading to a solution approach:
- The south American peri-Gondwana reconstruction (fig. 1).
- The Andean-Paleotethys peri-Gondwana integration (fig.2), and,
- A group of Lower Paleozoic, distant sedimentary sequences, for comparison.
The reconstruction for early Paleozoic times in South America is specially related with the Cambrian-Ordovician sequences, considering the locations with a continuous Vendian-Cambrian depositional history. The Gondwana platform includes cratonic fragments, folded belts, and basement recycled sequences with ages from Archean to Upper Proterozoic (fig. 1). The areas with Cambrian-Ordovician molasse-type, sedimentary cover have been outlined. The southernmost basin, located on the Uruguay-Brazil border region are Lower Cambrian in age (550 M.A.) (LEMOS & PINTO OLIVEIRA, 1983).

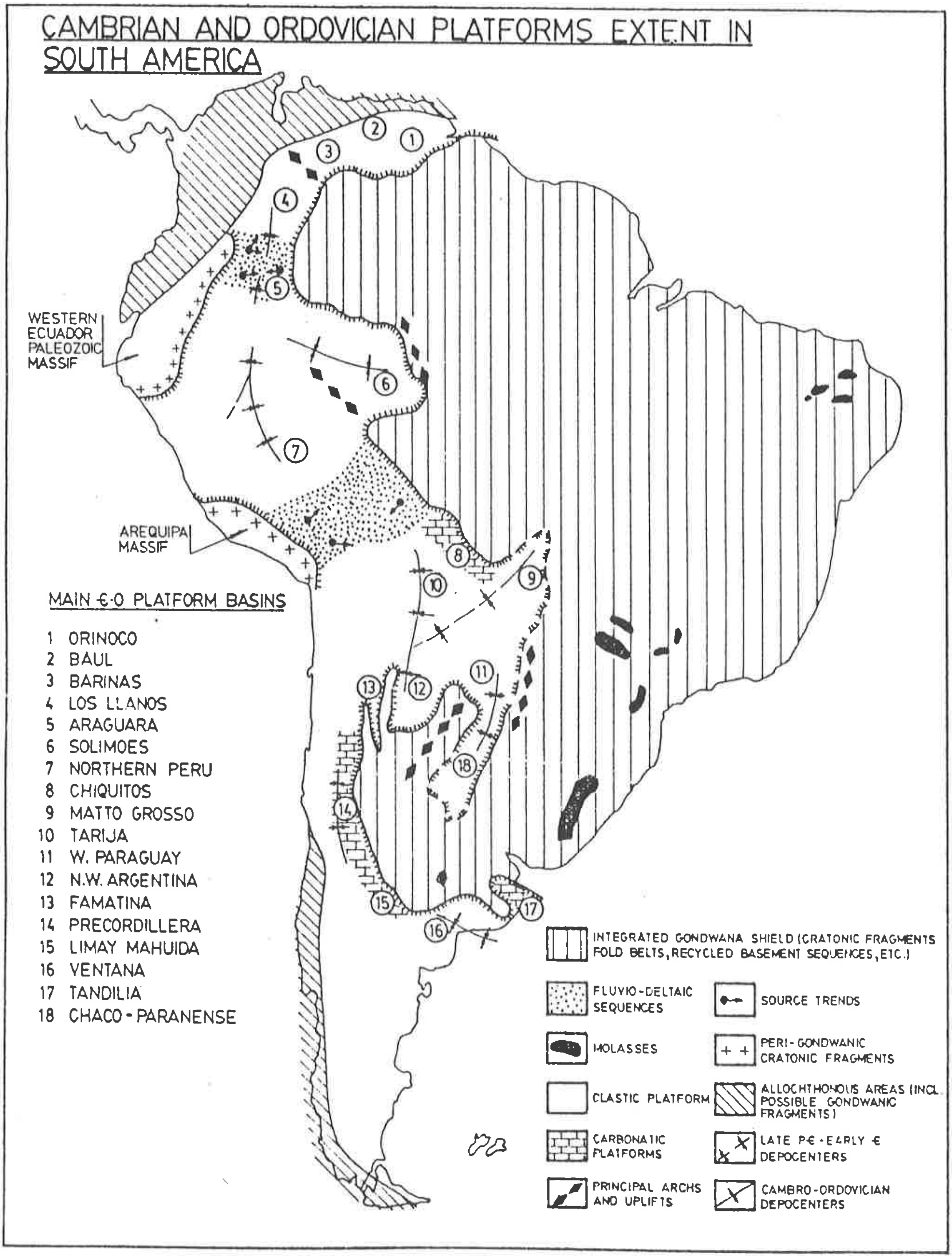

Fig. 1: Cambrian and Ordovician platforms extent in South America.

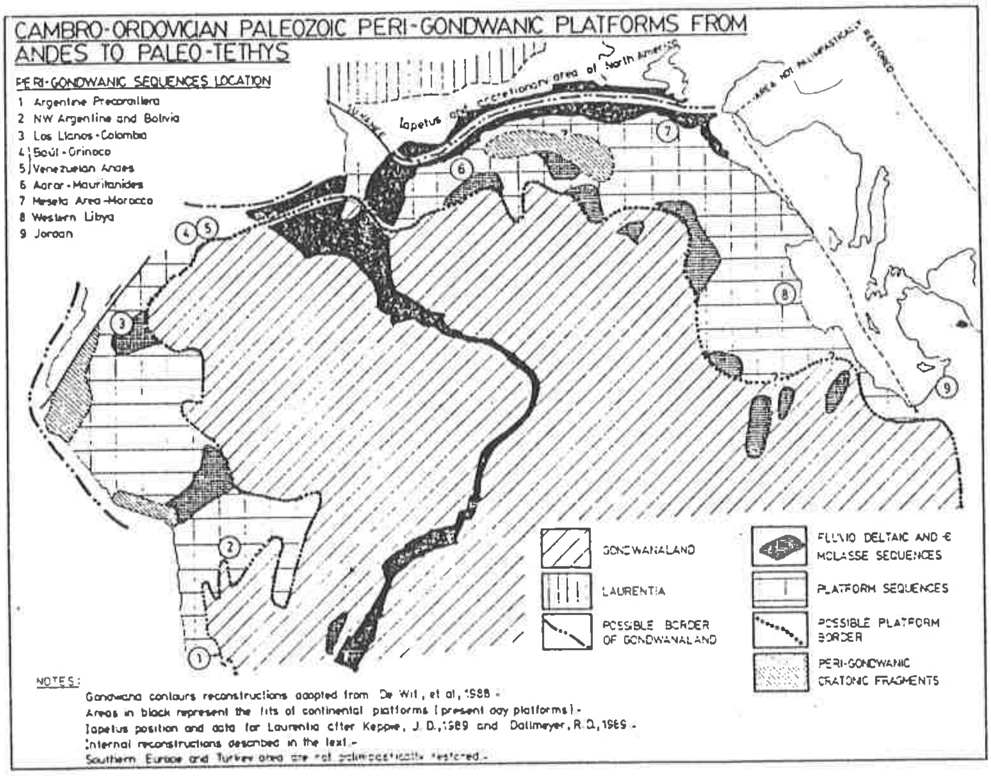

Fig. 2: Cambro-Ordovician Paleozoic Peri-Gondwanic platforms from Andes to Paleo-Tethys.

The peri-Gondwana reconstruction is mainly based on BALDIS & al. 1985 and BALDIS, 1988. The Cambrian and Ordovician platforms extends only along the present-day Andean Arc, but also over some of the Proterozoic central massifs. It is partially obstructed by two sectors of continental sedimentation which probably correspond to transversal ridges, north and south of the Amazonas Basin. The entire basin characteristics display evidences of controlling external margins, west of the Amazonas (BALDIS, 1988); at the Tarija Basin, along the NW of Argentina (OMARINI & al., 1988) and with marginal rift characters in the Precordillera (BALDIS & al., 1982) (RAMOS & al., 1986).
From North to South the South American platform comprises: The Central area of Venezuela to Colombia border area, where three main Cambrian-Ordovician basins are detected: The Orinoco basin (Lower Cambrian); the Baul Basin (Lower Ordovician), and the Barinas basin (Cambrian-Ordovician) all three have a shore line over the Guyana Massif and their platform area sloping to the present Caribbean region. The Barinas ridge separates the Los Llanos basin in Colombia from the terrigenous, continental sediments of the Araguara basin to the South. The Colombian basins possibly extends continuously from Middle Cambrian to Middle Ordovician, toward the south, they are controlled by a possible cratonic fragment remnant that continues through SW Colombia, Ecuador as an onshore massif to the Peru and the northern section of the present-day Amazonas basin, two Upper Cambrian and Lower Ordovician depocenters develop. The older is located in northern Peru and displays and Andean strike; the second corresponds to the Solimoes basin which displays a ESE depocenter orientation. Both basins are controlled by a cratonic marginal ridge and by an intrabasinal uplift. A highly possible Cambrian platform aperture to the Pacific is inferred.

The Arequipa and the Central Brazil massifs control Cambrian and Ordovician continental sediments in central Peru and Northern Bolivia. South of the Titicaca lake parallel to the NW of Argentina a variable platform develops, a NE trending depocenter, crossing the entire NW area including the Chaco, which corresponds with turbidites, limestones, and platform sandstones of late Precambrian to early Cambrian age named Puncoviscana Formation (OMARINI et al., 1988). The sedimentation is continuous from the Vendian to the Tomontian. A possible change in the orientation of the depocenter is assumed by middle-lower Cambrian. The North-South Andean trend depocenter has strong rift characteristics presenting a lateral extension regime until middle Ordovician with a later (Upper Ordovician-Silurian) compressive event, completed by intracontinental subduction and associated models (DAMM et al., in press). Toward the east, the Lower Paleozoic bounds the southern extreme of the Central Brazil and Bolivian cratons, with calcareous platforms facies in Chiquitos and Matto Grosso which probably correspond to lateral facies of the Puncoviscana sequence. Three regional "embayments" developed on Paraguay, Southern Bolivia and Northern Argentina. They correspond to intracratonic basins with SSW-trending depocenters in the Chaco-Paranense basin and the semi-marginal Famatina basin. This last acted as a possible active orogen associated to crustal thinning and Ordovician plutonism.

The Precordillera basin on West Argentine continues southward through the Limay Mahuida basin, and it probably interfingers with the Ventana and Tandilia basin in Southern Buenos Aires. The Precordillera displays an extensive calcareous platform developing from late-lower Cambrian to the beginning of the middle Ordovician. An active margin, located west of the platform sloped to a possible marginal rift depocenter area on the Eastern Andes region. The Ventana basin has been interpreted as an aulacogen (HARRINGTON, 1970, 1980), in the Tandilia area, Vendian (700-600 Ma.) limestones grade transitionally to Lower Cambrian carbonate deposits.

The described development associated to other Gondwana fragments is schematized in figure 2. Through this assemblage, the South American-African stratigraphic relationships lead to the correlation with their continuity in the Paleo-Tethys. The African platform has been given special consideration since in the North American sector of the Iapetus there is a general assumption that the non-Laurentian Lower Paleozoic sequences correspond to Laurassic allochthonous terranes (KEPPIE, 1989; DALLMEYER, 1989). The Eopaleozoic Gondwana fragments of Florida, with the Silurian Swanee basin are located in an intermediate position between South America and Africa (DALLMEYER, 1989; WRIGHT & al., 1989).

The African platform reconstruction is based on the data from LECORCHE & al., 1989; JABOUR & NAKAMAYA, 1988; DESTOMBES, 1971; LEGRAND, 1988; BENNIRAM, & al., 1988, and KLITZSCH, 1970.

The Adrar-Mauritanides, Hoggar-Tassili, and Southern Libya basins present lateral facies changes of continental sediments with African Shield source, to shallow marine strata for Cambrian and Lower Ordovician times. Also, from NE Egypt to the Sudan-Libya border zone a precise development history of "pull-apart" intracratonic basins has been defined (SCHANDELMEIER, 1988). These basins have thick, molasse-type sequences which correspond to Cambrian ages.

In the Mauritanides, Morocco, and the Sahara, the middle Cambrian-Ordovician depocenters are subparallel to the platform margins, also, some rift-type structure are present, in correspondance with controlling, continental border massifs, such as the Reguibat Uplift (DESTOMBES, 1971; PIQUE, 1989, and LECORCHE & al., 1989).

At this point, at the boundary with the Paleotethys, we establish a consistent integration scheme. It is characterized by the presence of a continuous Cambrian-Ordovician peri-Gondwana platform from Southern Argentina to Northern Africa. This platform shows sectors where fluvio-deltaic sequences and shallow platform strata are interfingered in a complex that also includes peri-Gondwana cratonic fragments that partially controlled the internal platforms and frequent rift marginal basins.

In figure 3, we present a series of eight type-sequences for a peri-Gondwana comparison. Their locations and authors references are:

1- Argentine Precordillera (BALDIS & PÖTHE de BALDIS, 1988; BALDIS & al., in press).

2- NW Argentina-Bolivia (ACENOLAZA & BALDIS, 1987).

3- Los Llanos-Colombia (ULLOA & al., 1982; Bogota, 1982, and BALDIS, 1988).

4-5- Baul-Orinoco-Venezuela Andes (MARTIN, 1977).

6- Adrar-Mauritanides (LECORCHE & al., 1989).

7- Morocco (Meseta Area) (JABOUR & NAKAYAMA, 1988; DESTOMBES, 1971).

8- SE Algeria-Central-West Libya (LEGRAND, 1988; BENNIRAM & al., 1988, and KLITZSCH, 1970).
9- Tabuk (S. Arabia)-Jordan (AL-LABOUN, 1988 and Mc CLOURE, 1988).

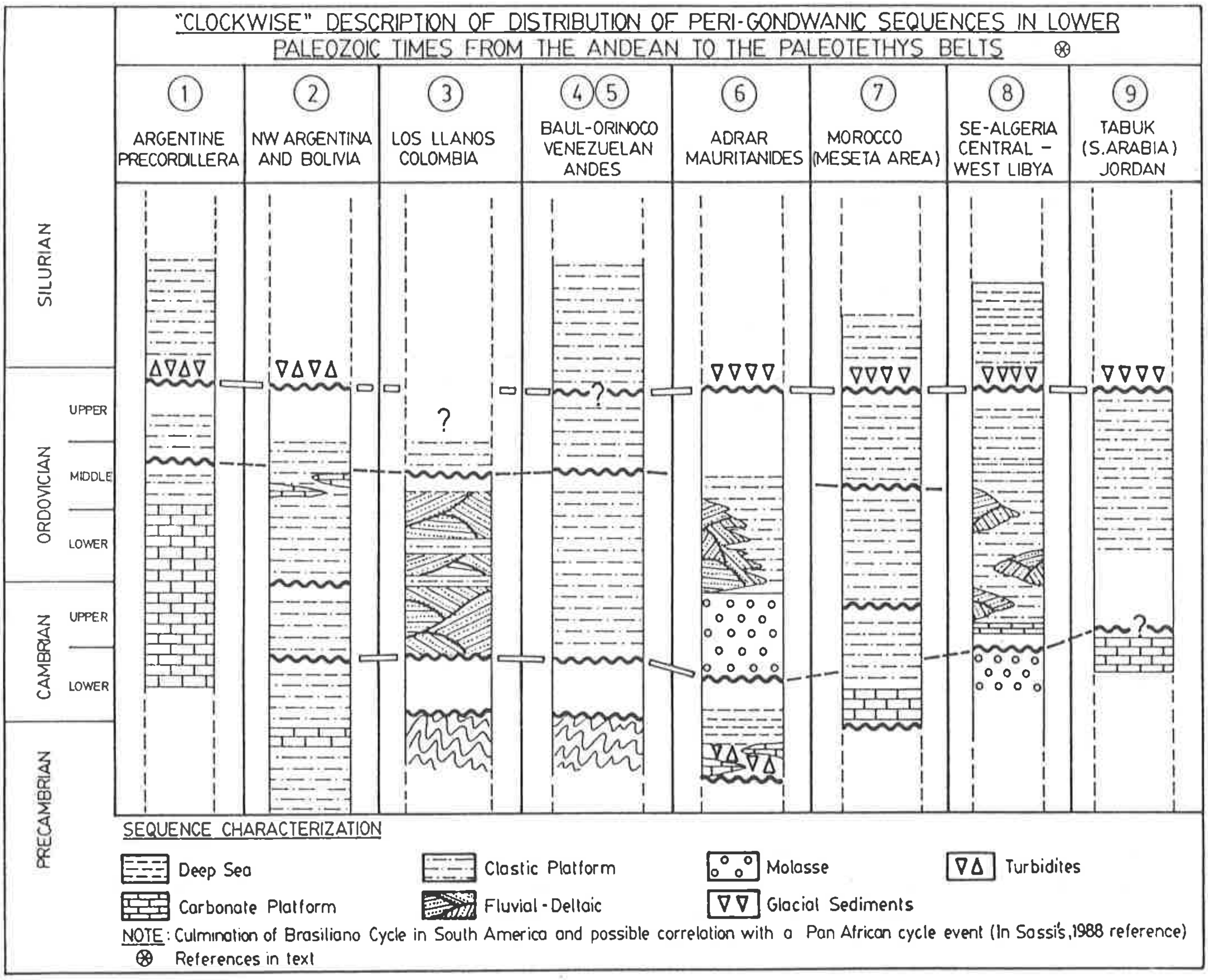

Fig.3: "Clockwise" description of distribution of Peri-Gondwanic sequences in lower Paleozoic times from the andean to the Paleotethys belts.

There exist logical lateral lithofacies variations but there is also a relatively precise correspondence of the upper Precambrian (Upper Vendian) to Upper Ordovician events, as listed:
1- A difened correlation of the Lower Cambrian unconformities from the Argentine precordillera to the Mauritanides, and that possibly reflected in the "pull-apart" basin of Chad and Sudan.
2- Some intracratonic movements that resulted in Upper Cambrian unconformities, which are well developed in NW Argentina, Bolivia, and Morocco.
3- Slight local unconformities at the Llanvirnian-Llandeilian limit.
4- The Gondwana intra-Ashgillian uncomformity, which corresponds to glacio-eustatic sea-level reactivation during Upper Ordovician times (JABOUR & NAKAYAMA, 1988; BEUF & al., 1971; VAIL & al., 1977).
In the figure 2 reconstruction, the Southern Europe contour represented has not been subject to palinspastic reconstruction. A possible Gondwana continental border has also been drawn for the Intra-Iapetus boundary, NE Caribbean region and Western South America.

3. CONCLUSIONS

This peri-Gondwana Lower Paleozoic reconstruction from the Tethyan to Andean areas indicates:
1- That continuous similar characteristics exist for the Cambrian platforms from South America and Africa to the Paleotethys boundaries. These uniform features are referred to the tectosedimentary conditions of the basins, as well as the geometry and the presence of controlling marginal cratonic fragments. All these considerations give a reasonable outline to continue the basin correlation analyses.
2- It is necessary to establish the characteristics of the platform borders, intra-platforms, and Gondwana cratonic structural elements that control the location of platform and intracratonic continental basins.
3- The Lower Paleozoic associated magmatic activity, crustal mobility, intracratonic fold belts reactivation, and any geodynamic complementary behavior have not been completely analyzed yet.
4- After the integration of the model and the literature analysis, a significant correspondence of Lower Paleozoic metamorphism and deformation events has resulted. The continuity of these phenomena from South America to Africa may refer them to the last Brasiliano cycle events of 570-450 M.a. of age (SCHOBBENHAUS et al., 1984) and to the Pan African II orogenesis (550 M.a.) (LECORCHE et al., 1989). The Pan African II orogenesis extent into the Paleotethys has been suggested as an intra-Ordovician event (SASSI, 1988).

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PALEOZOIC EVENTS WITHIN THE CAUCASUS SECTOR OF THE TETHYAN REALM

A.A. Belov

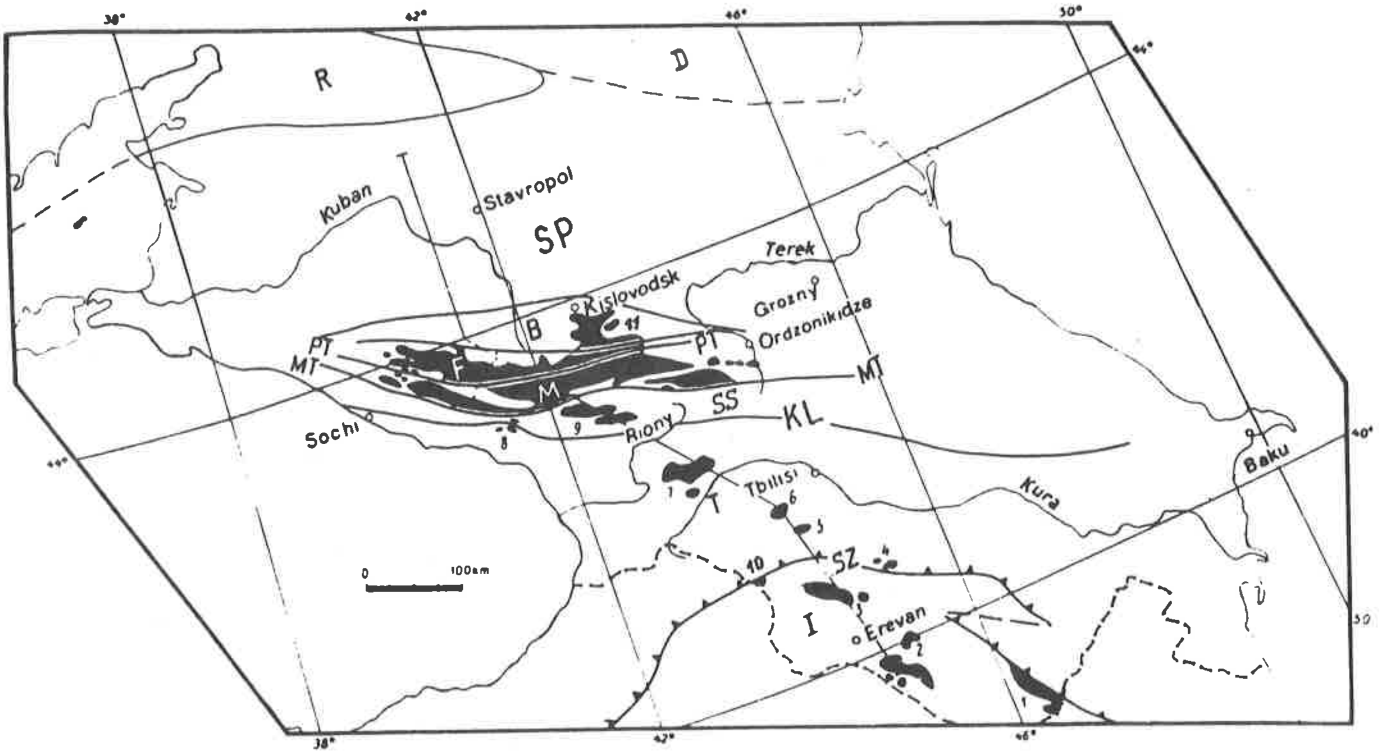

Fig. 1 - Scheme of distribution of the Caucasian pre-Alpine complexes. R -Rostov salient of the Ukrainian shield, D - continuation of the Donetz basin, SP - Scythian platform.
Pre-Alpine zones within the Alpine folded area: B - Bechasin, F - Forerange, M - Main Range, SS - Southern slope, T - Transcaucasian massif, I - nothern part of the Iranian plate.
Regions of distribution of pre-Alpine complexes over the surface: 1 - Zangezur, 2 - Daralagez, 3 - Tsakhunyats, 4 - Murguz, 5 - Loku, 6 - Khrami, 7 - Dzirula, 8 - Central Abkhazia, 9 - Svanetia, 10 - Amasya, 11 - Malka river upper course (Bechasin).
Fault zones: PT - Pshekish Tyrnyaus, MT - Main thrust, KL - Kakhetino-Lechkhum, SZ - Erzincan-Sevan.

1. INTRODUCTION

The witnesses of pre-Jurassic oceanic basins - pre-Jurassic ophiolite associations - have been known in several regions of the Caucasus (Fig. 1): in Hercynian zones of the Greater Caucasus - Bechasin (?), the Forerange and the Main Range and the Dzirula salient in Transcaucasia.
All attempts to prove the occurence of pre-Alpine ophiolites in the southern Sevan-Akera and Vedi zones and in the neighbouring regions cannot be considered successful. The available data are debatable and unconvincing. The direct data are missing. Evidences could possibly be obtained from the research done by KARYAKIN and ARISTOV. Their results are published in Doklady Akademii Nauk SSSR. Among the nappes of the Sevan-Akera zone, KARYAKIN and ARISTOV have discovered one composed of effusives with boudin layers of limestones. Petrochemical data classify the effusives as basalts similar to intraoceanic island types. The limestones are pelagic without any admixture of clastic material. Several conodonts have been found there (1-2 fragments), their age ranging from the late Paleozoic to Triassic, most probably, upper Permian. Additional paleontological research is required to confirm these data. The carbonate-volcanic pile established by KARYAKIN might appear even Triassic in age. In this case, it may be interpreted as a part of the Neotethyan oceanic crust.
In general, the Triassic system has not been studied properly enough in the entire investigated region. A lot of new data on the Triassic has been presented in the papers by SENGÖR and co-authors (1980, 1981, 1985). However, their conclusions give rise to certain doubts (for instance, BERGOUGNAN & FOURQUIN, 1982), the most important concerns the pre-Malm ophiolites of the Central and East Pontian as a Paleotethys suture. In view of their likely Mesozoic age, it would be more logical to regard them as the evidence of the early Neotethys rifting.
Triassic limestones and radiolarites have recently been found in the Sevan-Akera zone. The material is currently analyzed.

2. PALEOZOIC OF THE CAUCASUS

In spite of the fact that the future studies will surely confirm the presence of Paleozoic ophiolites in the Erzincan-Sevan-Akera zone, the corresponding oceanic basin can be attributed to Paleotethys II, if considering the time of its origin (possibly Early Carboniferous), or to Mesotethys (BELOV & al., 1986), proceeding from the time of its closing. It is noteworthy, however, that SENGÖR & al. (1985) support the ideas of BERGOUGNAN & PARROT (1981) and TEKELI (1981) and present the corresponding data as a possible opening of the Karakaya oceanic basin since the Carboniferous. However, reliable data do indicate a Jurassic-Neocomian age for the ophiolites in Transcaucasia (Fig. 2, X, XII).
Very interesting and significant Rb-Sr isochrones have been obtained by AGAMALYAN et al., (1982) for rocks of the crystalline basement of the Tsakhunyats (Miskhan) massif and Somkheto-Karabakh zone yielding ages of 620 and 300 m.y., respectively. They do not allow, in my opinion, to attribute the block located to the north of the Sevan-Akera suture to the East-European plate, and to demarcate the latter from the Gondwana-Land along this suture. Several manifestations of magmatism and weak dislocations of this time have been known to the south of Erzincan-Sevan line (for instance, alkaline monzonitic granites of the Bitlis massif, Rb-Sr 325-351 Ma. YILMAZ, 1971). Together with red formations of the verrucano type (for example, in the east of Central Taurus, BAYKAL, 1966), they indicate the influence of Hercynian orogenic processes over the entire territory to the south of the considered boundary.
Using paleobiogeographical methods to control the mobilistic reconstructions in the Caucasus sector of the Mediterranean belt presents a number of difficulties. The primary one is the abscence of faunistically confirmed deposits of the climatically contrasting Ordovician. Presence of the Ordovician in all Caucasian regions has not been confirmed by paleontological evidence. The second difficulty is the sporadic distribution and deficiency of fossils in Silurian and early Devonian rocks. During the Late Paleozoic and Early Mesozoic, the Paleotethys, broad in the East, grew thinner

there, or closed as a bay, and did not prevent the migration of plant and animal assemblages as was shown in many paleogeographical schemes compiled for that time on the basis of paleomagnetic data. The northern margins of Gondwana-Land, West Europe and the East European Platform were situated in the tropical and sub-tropical areas; hence, it is theoretically impossible to identify them by the paleobiogeographical method in first approximation. This is confirmed by similar Carboniferous floras of North Africa and Middle Europe, by finds of the European-type Namür-Bashkirian flora in the Khrami salient of the Transcaucasian massif, by the Carboniferous-Permian flora reported from the Zonguldak coal basin of Anatolia, and by the Carboniferous fauna (corals, foraminifers) of the Khrami, Svanetia and the Donetz basin regions (BELOV, 1981; BELOV & REITLINGER, 1966) as well as by the appearance of Early Permian fusulinids of the North-Tethys province in the southern margin of Paleotethys in the Elburs (LEVEN & SHCHERBOVICH, 1978). The same conclusions concerning the paleobiogeographical methods have been reached by ROBERTSON & DIXON (1985). They wrote: "The faunal evidence is thus consistent with a Mesozoic oceanic basin lying between Gondwana and Eurasia, but has little bearing on whether that ocean had persisted from the Paleozoic (Paleotethys), or whether it was newly created in the Mesozoic and if so, when. One must note that especially in the western Tethys, a wide area of carbonate platform (i.e., like the modern Bahamas) divided by deep rifted marine basins could have been a more effective barrier to faunal dispersal than any open oceanic separation" (p. 6).

The paleomagnetic studies of Paleozoic rocks carried out in the Caucasus and adjacent regions (ADAMIA & al., 1982, 1987), generally confirmed the present mobilistic reconstructions. However, these data are not sufficient yet, and the conclusions drawn on the basis of paleomagnetism are not convincing and require further checking.

Such kind of checking has been done for the Carboniferous of the Khrami salient in the Transcaucasian massif by ARAKELYANTS & al.(1989).

They have shown that all the rocks from the Carboniferous volcanic sedimentary sequence cannot be used for obtaining the ancient, i.e., Carboniferous component of remanent magnetization due to their paleomagnetic features. The only rock which has yielded good paleomagnetic results were diabases from the sills occurring in the Carboniferous beds. It is from the latter that our predecessors have received good paleomagnetic characteristics (ADAMIA et al., 1982).

However, the sills dated by K-Ar method appeared to be much younger than the Carboniferous; they were Cretaceous in age. Therefore, it is certain that in the Cretaceous the Transcaucasian massif belonged to the southern margin of Eurasia, whereas in the Carboniferous this remains questionable.

The phenomena of Paleozoic magmatism and metamorphism studied in the light of mobilistic reconstructions (ABEZADZE & al., 1982), give a coherent picture for the types of magmatic rocks and metamorphism associated with the continental margins and Paleotethys. It is especially the case for the Late Paleozoic orogenic magmatism of the Paleotethys northern margin (of Andean type) with its extended sublatitudinal volcano-plutonic belt (MOSSAKOVSKY, 1970). However, the Late Paleozoic acid magmatism of the Transcaucasian massif (both in its intrusive and extrusive form) does not allow to judge of the subduction zone inclination and its position to the south or north of the massif, because the dating of several manifestations of this magmatism is chiefly based on the K-Ar ratios.

We shall further discuss pre-Jurassic ophiolites and other Paleozoic complexes of the Greater Caucasus and Dzirula salient of the Transcaucasus which, as many authors believe, are the main witnesses of the Paleotethys former existence.

All mobilists asume that from the beginning of the Paleozoic to the end of the Devonian, a relatively broad Paleotethys ocean existed on the Caucasian traverse of the Mediterranean belt. It may be assumed that, primarily it was structurally more simple, i.e., it was a large trough-shaped deepening in the Earth's surface. The sections of ophiolite nappes in the Forerange characterize the oceanic bottom of this basin. Starting from the Silurian, the Paleotethys grew more complicated due to the changes in geodynamic conditions: the appearance of volcanic island arcs and detached marginal seas located in the active margin of the East-European plate.

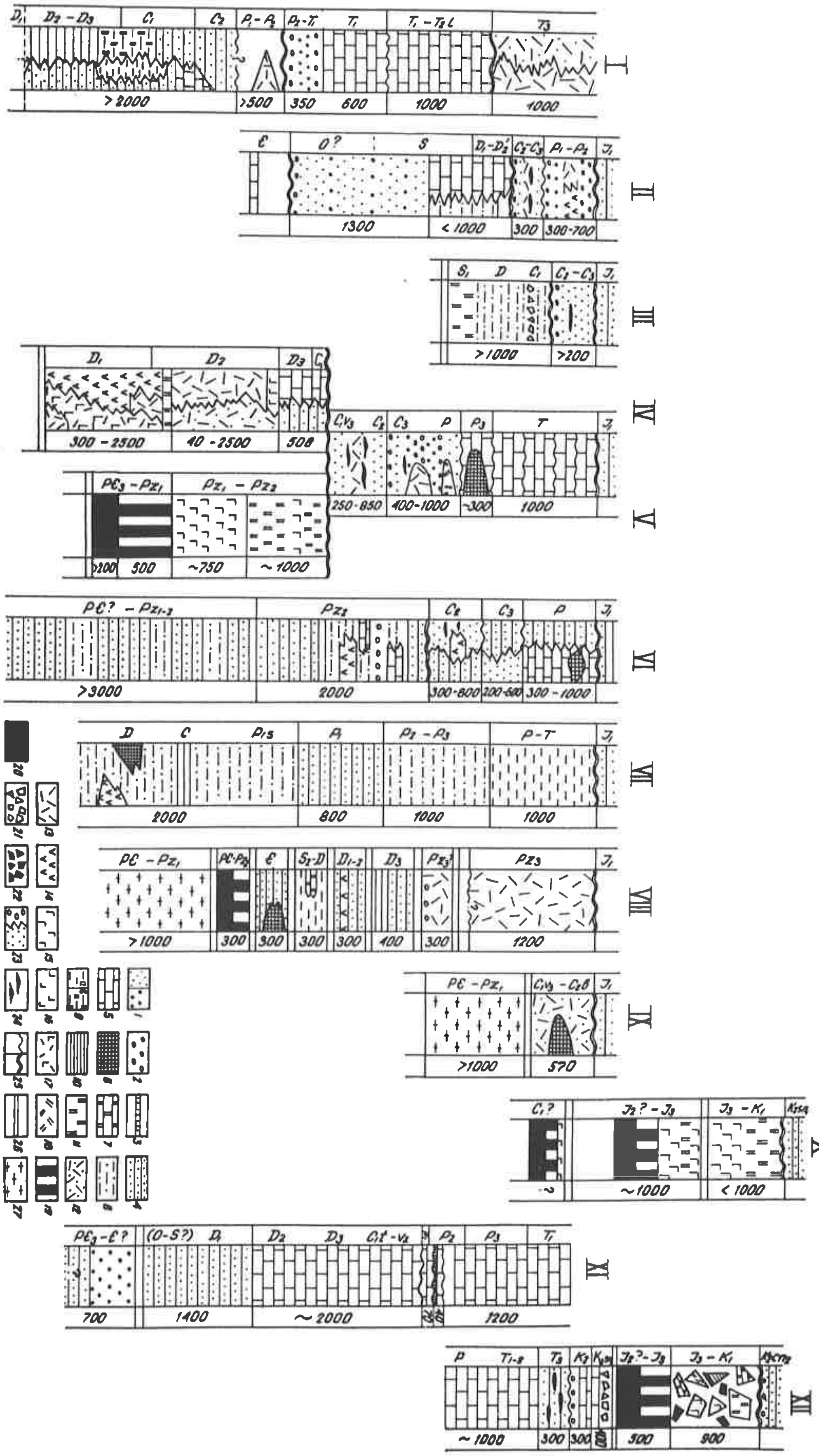

Fig. 2 - Leg. see next page.

Fig. 2 - Correlation of material formations of ophiolite belts and major pre-Alpine zones of the Caucasus. I - Precaucasus, II - Bechasin, III - Andryuk-Tokhana, IV-V - Forerange, IV - allochthon of the island arc complex, V - ophiolite allochthone, VI - Main Range, VII - Svanetia, VIII-IX - Transcaucasian massif, VIII - Dzirula salient, IX - Khrami salient, X - Sevano-Akera ophiolite belt, XI - South Transcaucasia, XII - Vedi; 1 - groups of mostly terrestrial terrigenous formations with predominance of sandy-clayey rocks: a) grey, b) red and multi-colored; 2 - terrestrial and coastal-marine terrigenous with predominance of coarse-clastic rocks; 3 - terrestrial bauxite-bearing crust of weathering; 4 - group of shallow-water marine terrigenous and carbonate-terrigenous formations (< 50% carbonates); 5 - group of shallow-water terrigenous-carbonate and carbonate formations (> 50% carbonates); 6 - reef limestones; 7 - pelagic limestone of moderate and considerable depths; 8 - flyschoid terrigenous and carbonate-terrigenous of various depths; 9 - deep-water flyschoid; a) terrigenous, b) siliceous-terrigenous; 10 - mostly deep-water black shale (aspidian); 11 - deep-water clay siliceous; 12 - marine of various depths, volcanic-sedimentary, mostly with acid tuffs; 13 - calc-alkaline volcanics of mostly acid composition (rhyolites, dacites); 14 - calc-alkaline volcanics of mostly mediate composition (andesites, andesite-basalts in island-arc complexes, sometimes together with tholeiitic basalt of higher potassium content); 15 - volcanics of higher alkali contents of acid and mediate composition (orthophyres, trachytes); 16 - low-potassic basalts of oceanic type; 17 - alkaline basaltoids of the potassium row (shoshonitic); 18 - group of volcanic formations of the contrasting (bimodal) association; 19 - gabbro-amphibolitic (IIIrd layer of the oceanic crust); 20 - Alpine-type ultramafites (fragments of the upper mantle); 21 - olistostromes, wild flysch; 22 - tectonic melange; 23 - lateral transitions between synchronous formations; 24 - coal content in sedimentary formations; 25 - unconformities: a) local angular, b) regional structural; 26 - tectonic contact of formations; 27 - metamorphic schists.

The Devonian-Carboniferous strata penetrated by holes in the pre-Caucasus (Fig. 2, I) correspond to the marginal-sea sediments, while the Devonian volcanism accounts for the first dislocations and complicated structure of the oceanic bottom as well as for appearance of island volcanic rises. The Silurian-Devonian sequences of the platform type, known in the Malka river upper course, show several rises with shallow-water carbonate sedimentation, possibly, of the Bahamas bank type; the facies transitions to terrigenous shaly sequences in the same region and to the Tokhana zone section can indicate the conditions of continental slope and basin within such rises (Fig. 2, II, III).
From the Devonian, the mobilistic reconstructions of the Caucasus began to alter. The early Devonian volcanism of the Forerange in the Greater Caucasus (Fig. 2, IV) is assigned by ADAMIA & SHAVISHVILI (1979) to the intraarc type, and the zone of the Forerange of that period is regarded as autochthone as well as rift related, thus complicating the Greater Caucasus island arc system. These concepts have been critisized by OMELCHENKO & al. (1984) as well as by BARANOV (1982) and KHAIN (1979). It is true that the composition of the allochthonous volcanic and volcano-sedimentary rocks of the Forerange, forming the Zaraus and Kartdzhyurt plates of the Kizilkov nappe, corresponds fairly well to an island-arc calco-alkaline volcanism. In the paleotectonic recontruction they occupy a zone which is three or four times wider than the present-day outcrops, and form a volcanic ensimatic arc. The adjacent zones of the Main Range and the Bechasin zone also comprise a number of nappes, partly including the island arc complexes, which is not the autochthonous margin of the intra-arc rift.
The main discussion, however, concerns the location of the Transcaucasian massif. ADAMIA & SHAVI-SHVILI (1979) attribute it to the northern continent from which it is separated by a small oceanic basin. The ophiolite allochthones of the Foreranges are considered to have originated from this basin (Fig. 2, V). True, they differ significantly in their effusive-sedimentary part from the first layer of the recent oceanic crust. According to their reconstruction, the main trunk of Paleotethys stretching southward from the Transcaucasus massif and throughout the Paleozoic and Mesozoic.

As the author had failed to find Paleozoic ophiolites to the south of the Greater Caucasus (the Dzirula massif will be discussed later) as well as in the continuous Paleozoic-Mesozoic sections in oceanic facies, he placed the Transcaucasian massif in the northern margin of the Gondwana-Land and in the southern margin of the Paleotethys. The absence of convincing data in the Paleozoic-Triassic-Jurassic sections of oceanic or near-oceanic facies is very important. The Paleotethys suture should be

looked for in the Greater Caucasus (ADAMIA & SHAVISHVILI, 1982; BELOV, 1981). In our opinion, the peculiarities of the ophiolite sections in the Forerange should be related to differences existing between paleo- and recent oceans. Georgian geologists consider that such a viewpoint may be confirmed by the interpretation of the history of the Iranian transverse through the Mediterranean belt (ADAMIA & al., 1982; SHAVISHVILI, 1979). They believe the Iranian part of Neotethys to have formed from the Zagros rift in the process of its expension during the Late Paleozoic, Early and Middle Triassic, whereas the South Caspian branch of Paleotethys has grown narrow. It follows from this assumption that the inherited Turkey-Transcaucasian part has continued since the Triassic as a newly formed Zagros part (see reconstructions by ADAMIA & al., 1982).
An intermediate version has been suggested by GRAMKRELIDZE (1982) who presented the Paleotethys as two branches separated by the Transcaucasian microcontinent (or island arc). Then the Transcaucasian microcontinent would become an accreted terrane within the active European margin.
The position of the Svanetia zone with a continuous, mostly terrigenous Devonian-Triassic marine section interpreted by most researchers as relatively deep-water deposits of the continental slope and rise, depends on the place occupied by the Transcaucasian massif in the reconstruction. The position of the Svanetia zone in the northern margin of the Transcaucasian massif is presently accepted as it has been confirmed by establishing a southern source area for the clastic material, including the sialic one (SOMIN, 1971; KUTELIYA, 1983). However, the reconstruction compiled by BELOV (1981, 1982) places the Svanetia zone at the foot of Gondwana-Land, separated from the Greater Caucasus by the major space of the Paleotethys. On the reconstructions performed by ADAMIA & SHAVISHVILI (1979), ABESADZE & al. (1982), this zone is located in the Southern margin of little oceanic basin between the Greater Caucasus and Pontian-Transcaucasian immature island arcs. The reconstruction by GAMKRELIDZE attaches no particular importance to the Svanetian section of the Dizi series. Late Paleozoic metamorphism and tectono-magmatic phenomema in the Transcaucasian massif, south of the tentative suture of Paleotethys can be explained by the Middle Carboniferous that they affected various deep levels in the mantle and crust and could result in shearing in the upper crustal horizons along the southern relatively passive margin (Fig. 3) of Paleotethys I, as well as by a possible two-sided subduction.

Over the last years, our knowledge of pre-Jurassic complexes of the Dzirula salient has considerably increased (ABESADZE & al., 1980; GAMKRELIDZE & al., 1981), however, it is still insufficient. The main information is as follows: the greater part of the salient pre-Jurassic basement is composed by plagiogneisses, crystalline schists (PE ?) and highly tectonized Late Paleozoic granites. Also we find there the Chiature sequence of quartz porphyry conventionally attributed to the Upper Paleozoic, and the narrow (2 km) Chorkhana-Utslevi band which shows the subvertical tectonic lenses and plates of

Fig. 3 - Paleogeodynamic profiles via the Caucasus sector of the Mediterranean belt.
1 - Upper mantle; 2 - hypothetical salient of the mantle in the mid-oceanic ridge; 3 - "basaltic" layer and layer 3 of the oceans; 4 - granite-metamorphic layer; 5 - island arc complexes; 6 - complexes of deposits of epicontinental seas and sediments of microcontinental cover; 7 - deposits of the continental slope and island arc slope; 8 - deposits of marginal and inner seas; 9 - layers 2 and 1 of the oceans; 10 - volcanogenic and sedimentary complexes of orogenic depressions; 11 - granites; 12 - plagiogranites; 13 - ophiolite complexes deformed in tectonic nappes and melanges; 14 - faults, surface of nappes, mantle and intracrustal shears; 15 - symbols of folds; 16 - volcanoes; 17 - water table; 18 - relative directions of movement of the upper parts of lithosphere.
The suture zone of Paleotethys could be drawn through the Dzirula salient characterized by pre-Jurassic ophiolites and other Paleozoic rocks, as it was done by H. BERGOUGNAN, C. FOURQUIN (1980).

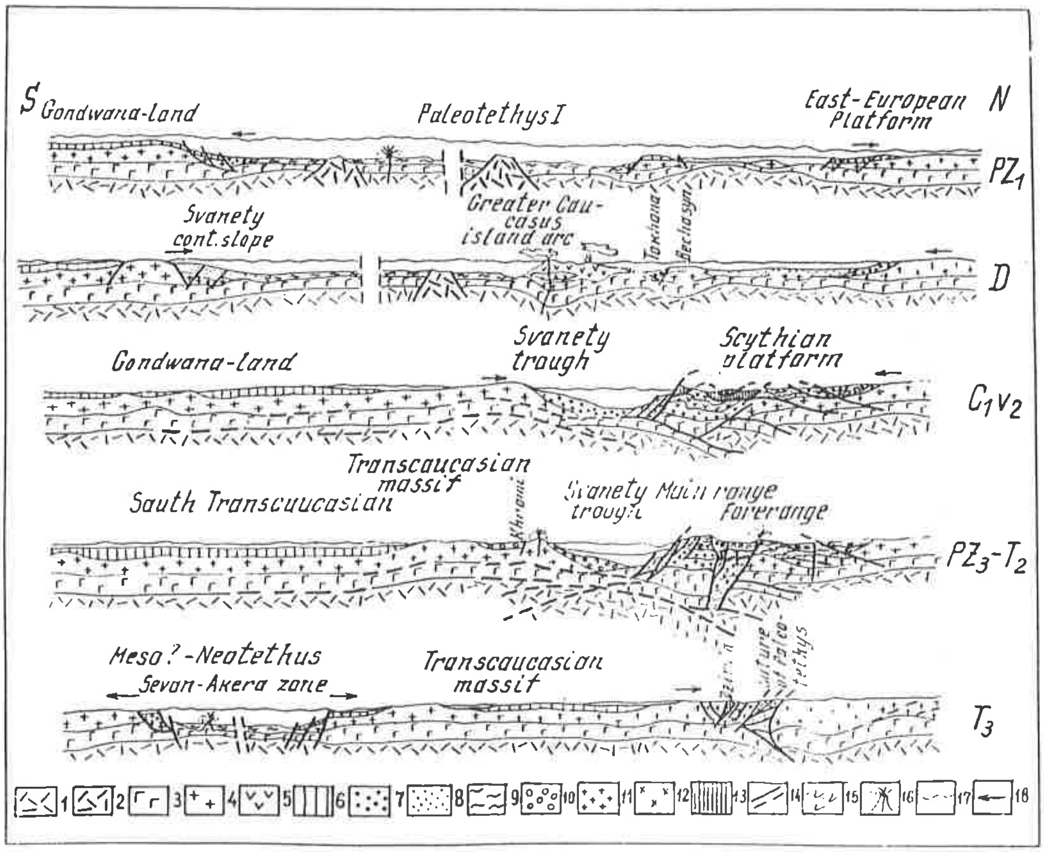

Fig 3 - **Leg.see preceeding page.**

serpentinites, milonites, amphibolites, gabbro-diabases of unknown age, crystalline sophists (PE ?) and Paleozoic rocks (Fig. 2, VII). The latter comprise phyllites, clay metasilt-stones and metasandstones with marble and quartzite lenses dated as Lower Cambrian; actinolitic slates, metabasites and metaporphyrites, basic metatuffs and phyllites assigned to the Upper Silurian-Devonian; metatuffs of quartz porphyries, metaconglomerates, metagritstones and metasandstones whose Upper Paleozoic age is not confirmed by faunistic evidence. The conglomerates incorporate pebbles of all the above rocks except serpentinites. Noteworthy is the abscence of normal contacts between all the above sequences and their insignificant thickness; the total visible thickness does not exceed 700 m (This testifies to their decrease in the process of tectonic displacement), different history of metamorphic transformations, attributing gabbro and diabases mostly to differentiates of the calco-alkaline orientation (GAMKRELIDZE & al., 1981), and the absence of a standard section of ophiolite association.

It follows from the above that various complexes of pre-Jurassic rocks of the Dzirula salient most probably belonged to different paleotectonic zones and were joined to each other tectonically. Their paleogeodynamic origin is not clear and ages are sometimes controversial. Based on the materials of the Dzirula salient, we can suggest several variants for the paleotectonic interpretation. In my opinion, there are grounds to date the structural formation of the Dzirula salient basement as Early Cimmerian (Indosinian). The same is true for the folding of the Svanetia zone and closure of Paleotethys. At the same time, there are non data on the direction of overthrusting of separate ophiolitic elements and their age as well as their interrelation and relation with faunistically characterized Paleozoic deposits. Their displacement to the present-day position is, most likely due to large strike-slip faults. In this respect, the ideas suggested by SENGÖR (1984) for the Caucasian-Anatolian sector of the Alpine zone are rather promising. KAZMIN & SBORSHCHIKOV have recently arrived at similar conclusion on the significant role of strike-slip faults in the formation of the Dzirula salient structure (1989).

3. CONCLUSIONS

The available data, including those obtained during the last 3-4 years, show that the history of tectonic evolution of the Caucasus is most convincingly and fully revealed through the analysis conducted in the light of mobilistic conceptions. The asymmetric structure of the Mediterranean belt and the existence of the Paleotethys ocean are almost generally accepted. The northern part of the Caucasus represented its northern active continental margin of the West Pacific type in the Middle Paleozoic and of the Andean type in the Late Paleozoic and of Andean type in the Late Paleozoic.
The evidence is that the North Caucasus Upper Paleozoic deposits were mostly formed in the continental conditions. They accumulated after the interval which followed the main Hercynian phase of folding (Sudet) in intermontane depressions, foothills of alluvial-proluvial plains, in small lakes inhabited by fresh-water fish and in bogs (Middle Carboniferous coal deposits in the Forerange).
The molasse sedimentation, both coal-bearing and red-colored, was accompanied by the calc-alkaline volcanism. The Late Paleozoic was characterized by further increase of the granite-metamorphic layer of the Earth's crust and by the formation and cooling of potassium granites (330-310, 280, 240-215 m.y., by K-Ar method).
The analysis of facies and thickness has shown the block character of the structures with considerable displacements along the now subvertical faults associated with overthrusting and formation of small nappes during the compression phases. In view of the above, the structure of the Upper Paleozoic complex is locally very complicated including its Middle Paleozoic elements (BELOV & OMELCHENKO, 1986). During the Paleozoic, the southern part of the Caucasus, Iran and almost entire Anatolia were a relatively passive margin of Gondwana.
The Hercynian events were manifested in the Armenian part of the Transcaucasus by a hiatus between the Visean and Lower Permian. They have resulted in the washing out of certain pre-Permian deposits and formation of lateritic crust of weathering with small bauxite deposits. Incidentally, the absence of certain horzons from the section increases in the northward direction. This suggests the existence of an upland further northward that could only be the Transcaucasian massif (Fig. 3).
At the end of the Paleozoic- beginning of the Mesozoic, part of the Gondwana margin including the Southern Transcaucasia area was transformed into a microcontinent as a result of rifting and formation of the new oceanic basin of Neotethys.
There are numerous controverses regarding the history of the Caucasian region, namely: the position of the Paleotethys suture and the time of its closure, the width of oceannic basins, the inherited or newly-formed nature of Neotethys. In my opinion, these are due to the lack of evidence as well as to the differences existing between the Earth's image with its large oceans and continents and the same homological structures of the past geological ages. This primarily refers to the Mediterranean belt of the Tethys area which was recognized by KARAMATA (1983) as an area of specific plate tectonics manifestation characterized by a relatively rapid reconstruction of the dynamic regime as well as by a successive change from extensions to compressions and vice versa within the limited oceanic space with numerous microcontinents. In this connection, the elementary estimates also disagree with the concept of the inherited Neotethys that assumes an ocean, 12 000 Km wide, with a spreading rate of 2 cm per year during 300 m.y. Even a considerable shortening of the northern half along the passive Gondwana-Land margin which should subsequently be consumed in the Late Mesozoic at an extremely high speed. Meanwhile, the acknowledgement of a limited subduction along the southern Paleotethys margin (BELOV, 1981; SENGÖR, 1981), i.e. under the Transcaucasian and South Anatolian massives in the Late Paleozoic and Triassic provides an easier solution to the problem of the Mediterranean paleooceans space.

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MISE EN EVIDENCE D'UNE DISTENSION TARDI-METAMORPHIQUE PRECOCE DANS LES SEBTIDES INFERIEURES (RIF INTERNE, MAROC).

*

EVIDENCE OF A PRECOCIOUS LATE METAMORPHIC EXTENSION IN THE LOWER SEBTIDES (INTERNAL RIF, MAROC).

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Abstract

The tectonic study of the gneisses and microschists of the lower Sebtides units, located in "Cabo Negro" and Torerha" areas allowed to evidence several late metamorphic extensional structures materealized by flexures with axes oriented N.140, which after become normal faults trending N.110 to N.150. These flexures and fractures, filled with metamorphic material, show that the distension occurred by the end of the metamorphism when the rocks were still hot and ductile. The plat of the planes on a wuffret, and the use of the right dihedra method shows a NE-SW to N-S extension. This extension stared by the end of the alpine metamorphism that affected these series of the internal Rif. i.e. since the Oligo-Aquitanien. This age of obtained through radiometric datations of the metamorphism of the Sebtides units (22 to 25 Ma). These Late-metamorphic faults run along the mediterranean coast from Sebta to Jebha, and can be related to the precocious down falling of Alboran sea.

Resumé

L'étude tectonique des gneiss et micaschistes des unités des Sebtides inférieures que affleurent dans les régions de "Cabo Négro" et de "Tarerha" a permis de mettre en évidence des structures distensives tardi-métamorphiques matérialisées par des flexures à axes orientés N.140, qui évoluent souvent vers des failles normales orientées N.110 à N.150. Ces flexures et cassures, remplies par du matériel métamorphique, montrent que la distension s'est produite vers la fin du métamorphisme pendant que la roche était encore chaude et plastique. Le repport de ces plans de failles normales tardi-métamorphiques sur Canevas de Wulff et l'application de la méthode des dièdres droits montre une extension NE-SW à N-S. Cette distension a débuté vers la fin du métamorphisme alpin ayant affecté ces séries, du Rif interne, c'est à dire, dès l'Oligo-Aquitanien, âge attribué par datation radiométrique au métamorphisme des unités Sebtides (âge absolu de 22 à 25 Ma). Ces failles normales longeant la côte méditerranéenne des Sebtides à Jebha, peuvent être liés à l'effondrement précoce de la mer d'Alboran.

1. INTRODUCTION

L'effondrement de la partie centrale de la plaque d'Alboran est considéré par certains auteurs comme ponto-plio-quaternaire (GLANGEAUD, 1961 et 1971, EL GHARBAOUI, 1981, ESTEVES et SANZ DE GALDEANO, 1983, MOREL, 1987), par d'autres comme Miocène moyen (OLIVET et al., 1973, RENAULT, 1984). Cependant, il s'avère qu'il a débuté beaucoup plus tôt dès la fin des charriages des domaines internes rifains (CHALOUAN 1986). D'après cet auteur, la distention responsable de cet effondrement aurait fonctionné en saccades à l'Oligo-Aquitanien, après les retrocharriages des nappes du Rif interne. Celle-ci est caractérisée par des failles normales parallèles et perpendiculaires à la côte entre Sebta et Jebha, dont l'âge est post-burdigalien-anté-pliocène (CHALOUAN et al.,1989). Le but de notre étude est de caractériser d'avantage cette distension précoce oligo-aquitanienne. Elle est bien décelée cette fois-ci, dans les unités métamorphiques de Filali des régions de "Cabo Negro" et de Tarerha" (Fig. 1).

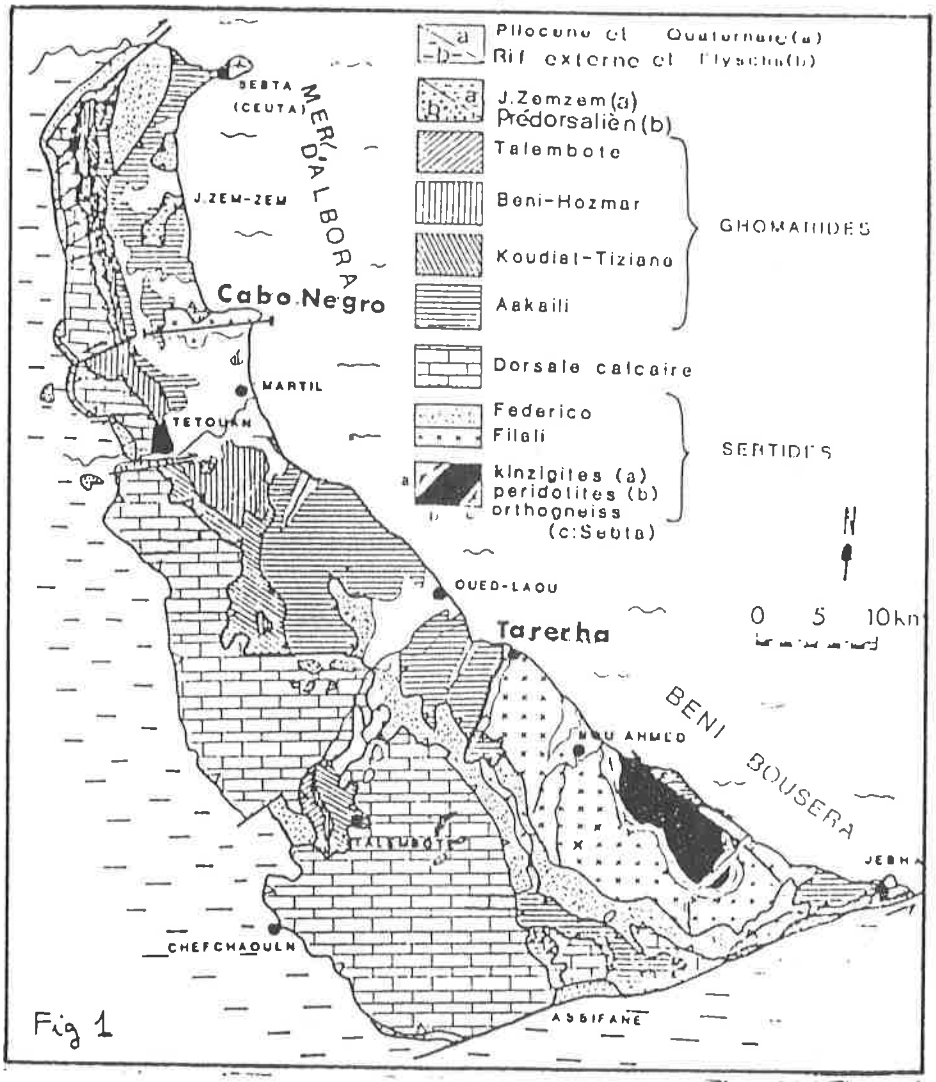

Fig. 1: Carte structurale du Rif interne, in SUTER 1980 modifiée.

2. UNITE DE SIDI AHMED-EL-FILALI:

Cette unité métamorphique de Filali couvre une importante superficie du Rif interne. En plus du massif de "Cabo Négro" où elle est cernée par la mer et les terrains paléozoïques des Ghomarides, elle affleure au Sud de Tetouan, où elle ceinture l'ensemble kinzigite-peridotite de l'unité de "Beni Bousera". Au Nord de Tetouan, dans la région de Sebta, on retrouve les traces de cette unité intercalées sous les assises métamorphiques des "Unités de Federico" qui sont totalement absentes dans la région de Tétouan. Cette unité assez monotone, est constituée de gneiss armès de niveaux leptynitiques à la base et des micaschistes et des quartzites au sommet (Fig. 2).

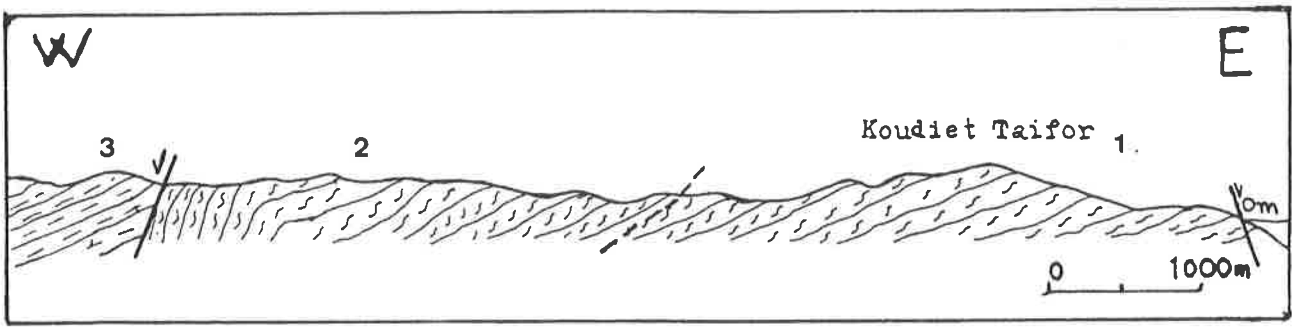

Fig. 2: Coupe dans l'unité de Filali dans le Cap Noir. 1: gneiss, 2: micaschistes, 3: schistes (Ghomarides).

2.1 Métamorphisme :

L'unité de Filali est affectée par un métamorphisme polyphasé qui affecte l'ensemble des terrains Sebtides.

2.1.1 - Structures synmétamorphiques :

Que ce soit dans les gneiss ou dans les micaschistes, on observe des structures synmétamorphiques telle que les foliations, les plis et les linéations. La foliation principale "S_2" est marquée dans les micaschistes par l'alternance de niveaux micacés et quartzeux, elle est parallèle ou rubannement quartzo-feldopathique dans les gneiss. Dans les deux formations elle est associée à des plis isoclinaux "P_2" dont elle constitue le plan axial. Une linéation "L_2" a été mesuré dans toute l'unité, elle résulte du plissement "P_2" accompagné de la foliation "S_2" (KORNPROBST, 1974). La reprise d'une foliation primaire "S_1" par "S_2" a été observée aussi bien dans le terrain qu'en lames minces (SADDIQUI, 1988). Elle est associé probablement à un plissement "P'_2" ou "P_3" qui reprennent la linéation d'étirement "L_2" (SADDIQUI, 1988).

2.1.2 - Composition minéralogique :

a) - Gneiss : Ces gneiss sont constistués d'une alternance de lits clairs quartzo-feldspathiques et des niveaux sombres riches en biotite et en sillimanite. La composition minéralogique de ces roches est constituée de quartz, plagioclase, feldspath potassique, biotite, sillimanite, muscovite, grenat et tourmaline. Tous ces niveaux n'appartiennent pas à la même paragenèse ils sont le résultat d'une succession de paragenèse associées à l'histoire tectono-métamorphique complexe subie par les gneiss (KORNPROBST, 1974; SADDIQUI, 1988). Ces gneiss renferment à leur base des niveaux migmatitiques dont la composition minéralogique n'est pas très différente de celle des gneiss sauf la présence de la cordiérite et des barres de leptynites à grenat, silicates d'alumine et tourmaline.

b) - Micaschistes : Ces roches sont en continuité avec les gneiss, le passage progressif d'une formation à l'autre ne permet pas le tracé d'une limite cartographique nette entre les vrais gneiss et les micaschistes proprement dite (Fig. 2). La composition

PHASES	PLIS	FOLIATIONS	LINEATIONS	ROCHES	PARAGENESES
Phase1	P_1	S_1		Micaschistes	Grenat Almandin.
				Gneiss	Plagioclase,Disthéne,Grenat,Feldspath potassique,Sillimanite.
Phase2	P_2	S_2	L_e	Micaschistes	Sillimanite,Biotite,Almandin,Plagiclase,Quartz. Sillimanite,Biotite,Muscovite,Almandin,Plagioclase,Quartz. Biotite,Muscovite,Almandin,Plagioclase ,Quartz.
			L_m	Gneiss	Biotite,Sillimanite,Feldspath Potassique,Plagiclase,Quartz. Biotite,Sillimanite,Muscovite,Plagiolase,Quartz.
Phase3	P_3	S_3	L_o	Micaschistes	Biotite,Muscovite,Disthène,Staurotide,Andalousite en cristaux.
				Gneiss	Biotite,Muscovite en paillétes.

Fig. 3: Tableau récapitulatif des structures syn-métamorphiques dans les Sebtides inférieures d'après KORNPROBST (1974) et SADDIQUI (1988).

minéralogique est constituée de minéraux dérivant de paragenèses légèrement différentes (grenat, biotite, muscovite, sillimanite, plagioclase, quartz, disthène, feldspath, potassique, staurotide et andalousite). Le tableau de la figure 3 récapitule toutes les structures synmétamorphiques affectant l'unité de Filali ainsi les paragenèse correspondants.

2.2 *Geochronologie*

L'âge du métamorphisme que affecte l'ulnité de Filali a été considéré comme précambrien ou paléozoïque ancien (KORNPROBST, 1971, 1974, 1976), puis varisque (MICHARD & CHALOUAN, 1978) ou alpin (REUBER & al., 1982; MICHARD & al., 1983), deux types des données ont été utilisées pour dater le métamorphisme alpin des Sebtides : d'une part des données stratigraphiques et d'autres part des données de géochronologie isotopique.

a) - Données stratigraphiques : L'âge alpin a été déterminé au début sur des bases stratigraphiques par MILLARD dès 1960 (in DURAND DELGA & al., 1960-62). Cet auteur a pu daté par des algues les dolomies métamorphiques qui couronnent les métasédiment des nappes Federico. Leur âge triasique (Trias supérieur) lui a permis de conclure à un âge alpin de ce métamorphisme, sans plus de précisions. Par ailleurs, la couverture "post-nappe" détritique discordante sur ces divers unités métamorphiques est datée du Miocène inférieur (Formation burdigalienne de Venuela transgressive sur les Alpyarides homologues espagnoles des ghomarides (Formation aquitanienne d'Alozaina de BOURGOIS & al., 1972, OLIVER, 1984).

b) - Données isotopiques : Les datations isotopiques qui ont été réalisées dans les Sebtides donnent des résultats dispersées qui indiquent la présence d'un métamorphisme alpin de haut degré: 20 Ma par la méthode K/Ar sur une muscovite (veine) et une biotite (gneiss) (LODMIS, 1975); 20 à 22 Ma par la méthode Rb/Si et Sm/Nd sur grenat dans les péridolites, Kinzigites et sur une aplite (POLVE et ALLEGRE, 1980, POLVE et al., 1988); et enfin 20 à 25 Ma par la méthode K/Ar depuis les Kinzigites de Beni Bousera jusqu'au Permien de Beni Nezala (MICHARD et al., 1983). Ces âges correspondraient au dernier évènement thermique ayant touché ces unités

2.3 Distension tardi-métamorphique

Cette distension précoce a été mise en évidence dans les Sebtides inférieures représentée par les gneiss de "Cabo-Négro" et les micaschistes de "Tarerha" de l'unité de Sidi-Ahmed-El-Filali.

2.3.1 - Gneiss de "Cabo Négro" :

Situé à 12 Km au Nord de Tétouan, le Cap culmine à 332m laissant affleurer les gneiss et micaschistes métamorphiques de l'unité de Filali. La coupe le long de la côte par deux types de déformations :
- des flexures à axes orientés généralement N.140 (fig. 4), elles affectent les lits clairs quartzo-feldspathiques et les lits sombres riches en micas. Le matériel étant encore ductile;
- Ces flexures évoluent souvent vers des failles normales tardi-métamorphiques de direction général N.120 à N.160, inclinées entre 50 et 90^{o} tantôt au NE, tantôt au SW (fig. 5). Les plans de ces failles portent des stries de très fort pith (70 à 90^{o}) qui montrent que ce sont des failles nomales dans la région de "Cabo Négro", leur report sur Cavenas de Wulf et l'application de la méthode des driedres droits (ANGELIER et MECHLER, 1977), après correction du dernier plissement, permettent de donner à cette distension précoce une direction générale NE-SW à N-S (fig. 5).

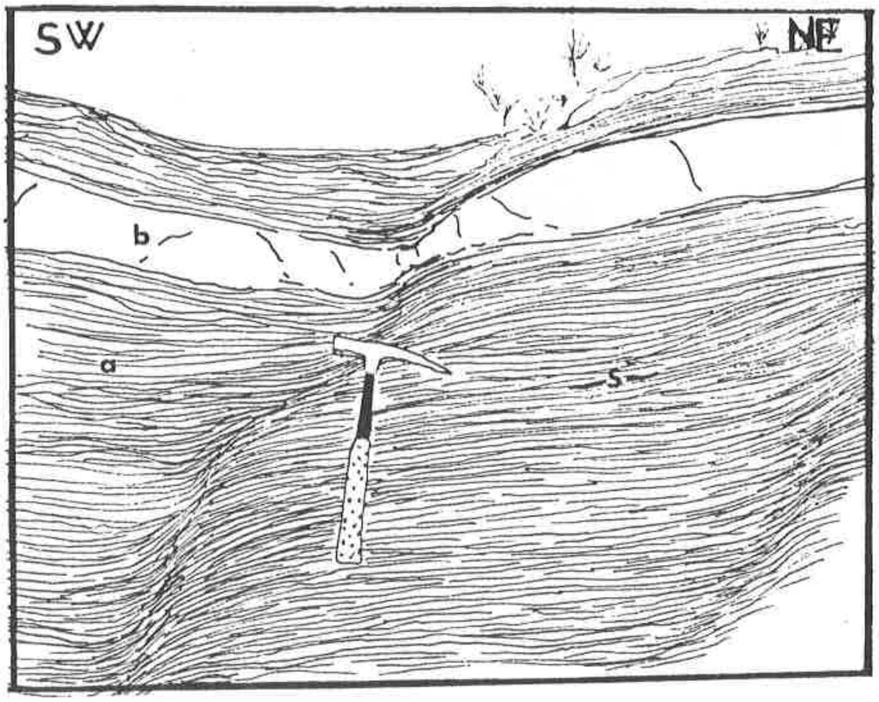

Fig. 4: Flexures distensives dans les gneiss de Cabo Negro. a: lits sombres riches en micas, b: lits clairs quartzo-feldspathiques, S: foliation principale.

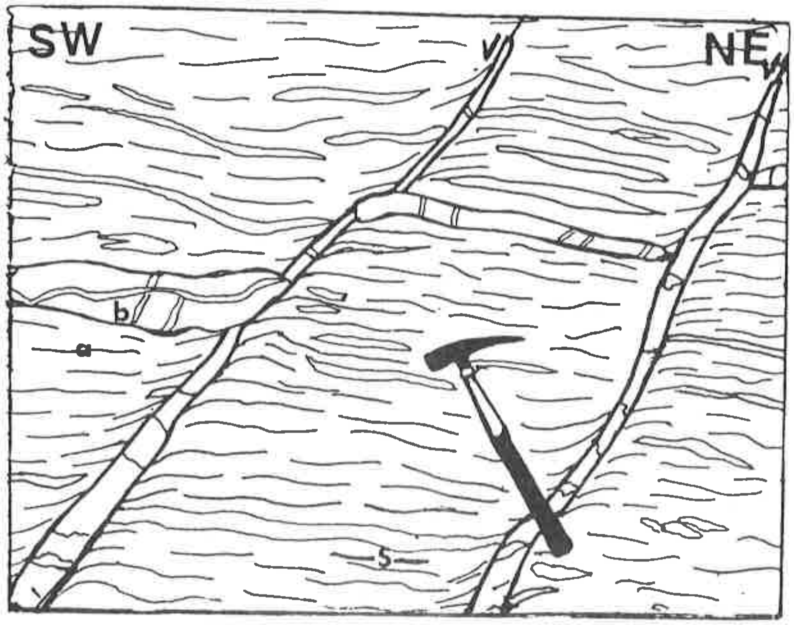

Fig. 5: Failles normales tardi-métamorphiques dans les gneiss de Cabo Negro. Légendes voir figure n°4.

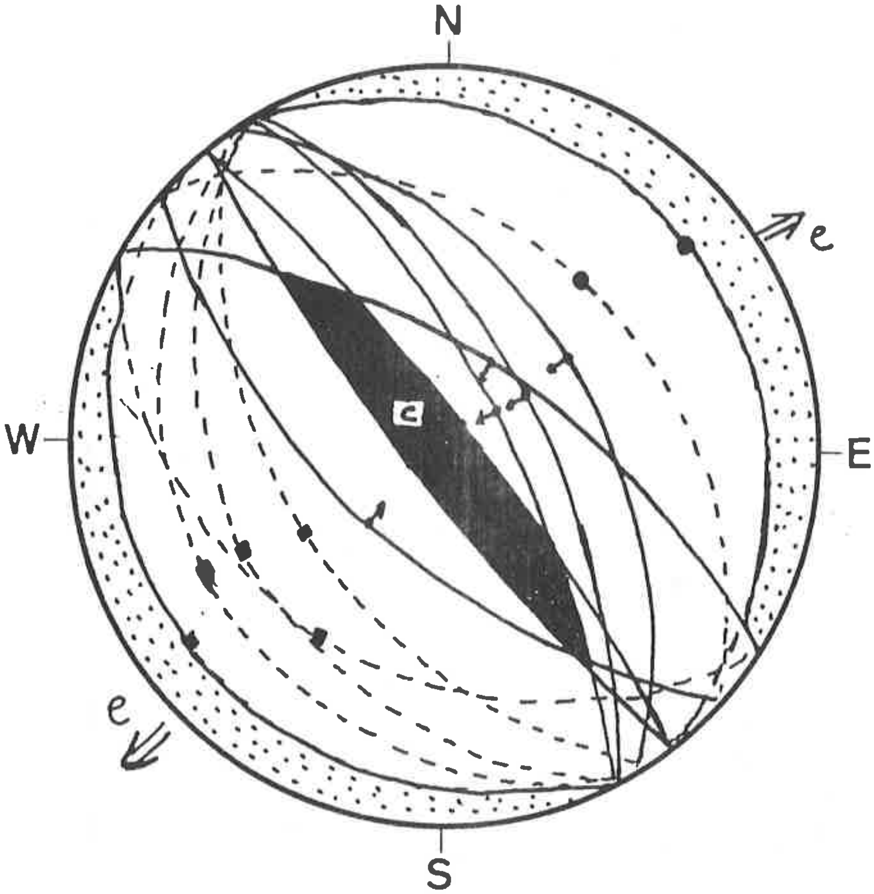

Fig. 6: Détermination par méthode des dièdres droits des aires de compression et d'extension pour les failles tardi-métamorphiques dans les gneiss de Cabo Negro. Hémisphère inférieur, (e) Extension, (c) compression, () Polaire de la faille normale, () Stries de friction, () Trace cyclographique du plan perpendiculaire à la faille et à la strie.

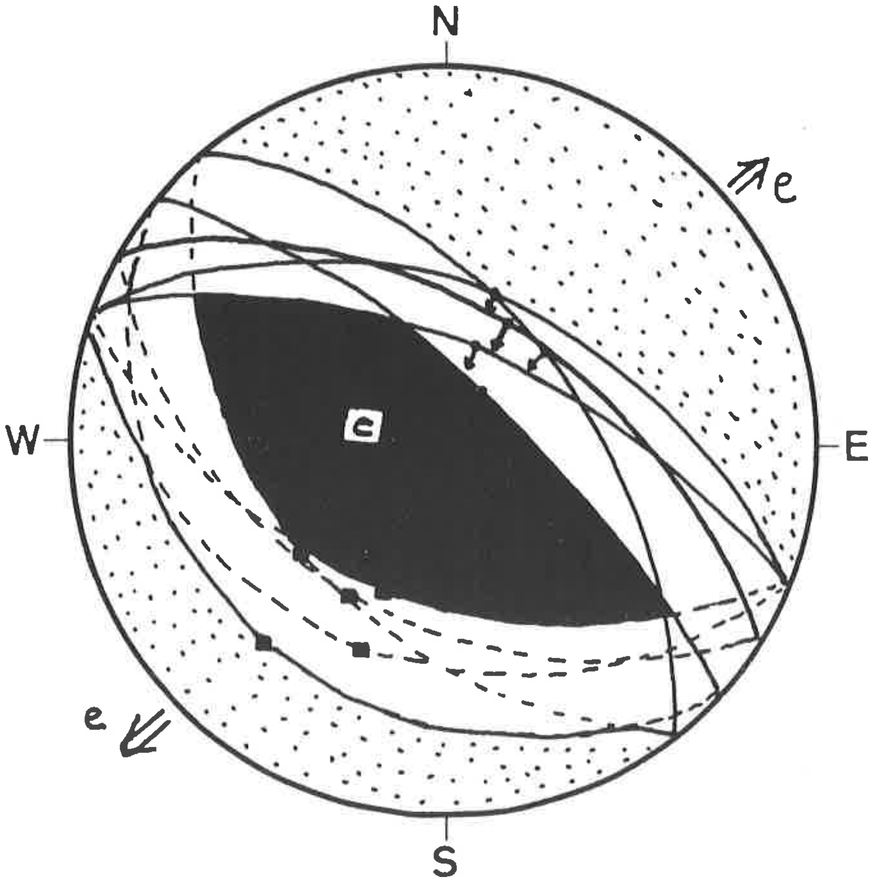

Fig. 7: Détermination par la méthode des diedres droits des aires de compression et d'extension pour les failles tardi-métamorphiques de Tarerha. Même légendes que la figure 6.

2.3.2 - Micaschistes de "Tarerha"

Les mêmes gneiss et micaschistes de Filali affleurent dans la région de Tarerha à 60 km au sud de "Cabo Négro". Ils présentent les mêmes structures que dans la région de "Cabo Négro". En effet, le long de la côte entre le village de Stihate et de Tarerha affleurent les gneiss et micaschistes de l'unité de Filali, avec une foliation principale sub-horizontale ou faiblement incliné vers le Nord. Les terrains sont affectés par les mêmes types de failles normales tardi-métamorphiques dont les cassures sont remplies par du materiel acide et métamorphiques. La direction de ces failles varie généralement entre N.100 à N. 140, ce qui permet de conclure à un épisode distensif de même orientation que celui observé dans la région de "Cabo Négro".

2.3.3 - Conclusion

Les cassures de ces failles normales sont colmatées par du matériel de remplissage clair dont l'étude au microscope montre que ce n'est pas seulement un remplissage quartzo-feldspathique mais on y trouve également des paragenèses de métamorphisme, à savoir, biotite, sillimanite, feldspathe potassique, grenat et cordierite (fig. 8). La présence de ces failles normales, tardi-métamorphiques dans l'unité de Filali montre que la distension a débuté avant le refroidissement totale de la roche, le matériel étant encore plastique et dans des conditions thermo-barométrique telles qu'il peut y avoir fusion anatectique et phénomène de migmatisation. De tels phénomènes ont été observé dans le plateau d'Arguronde dans le massif central français (M. FAURE et al., 1990). Ici la déformation extensive responsable se produit également en climat métamorphique rétrograde à une époque (300 à 320 Ma.) considéré comme le stade terminal de l'orogenèse varisque.

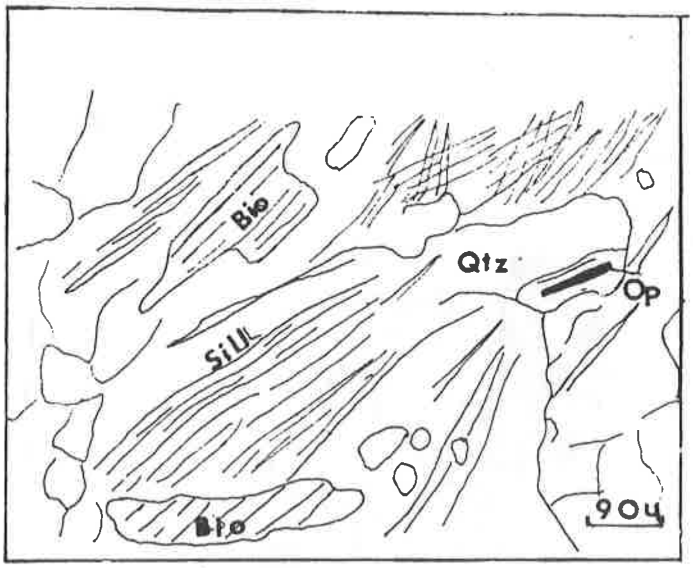

Fig. 8: Lame mince dans les cassures des failles normales tardi-métamorphiques montrant que le remplissage est constitué par un matériel migmatitique. On y trouve de la sillimanite (sill.), biotite (Bio.), feldspath (Fds), quartz (Qtz) et les minéraux opaques (Op.).

3. DISTENSION TARDI-METAMORPHIQUE A L'ECHELLE DU RIF INTERNE

Une distension fort contemporaine a été observé et datée, plus au Nord, dans l'extrémité septentrionale du Rif interne, dans la région de Sebta, à la sortie du tunel de Fnidek (CHALOUAN, 1986). Elle se manifeste par des failles normales décimétriques à plurimétriques conjuguées (N.167,66 N et N.170, 60 E, après correction de pendage en dépliant le synclinal oligo-miocène) affectant les écailles de carbonifères et de Trias de la nappe d'Aakaïli. Ces failles décalent en jeu normale les bancs de grès et de conglomérat carbonifères et les lentilles tectoniques de grès rouges et dolomies du Trias, que l'on peut associés au chevauchement initiaux ghomarides d'âge post-éocène supérieur - anté-oligocène supérieur (CHALOUAN, 1976). L'orientation de quelques miroires de failles normales conjugués observés dans cet affleurement permet de conclure à un épisode distensif de même orientation (NE-SW) que la distension tardi-métamorphique rencontré dans l'unité de Filai. Ces failles normales sont scellées par la transgression oligo-aquitanienne (fig. 9).

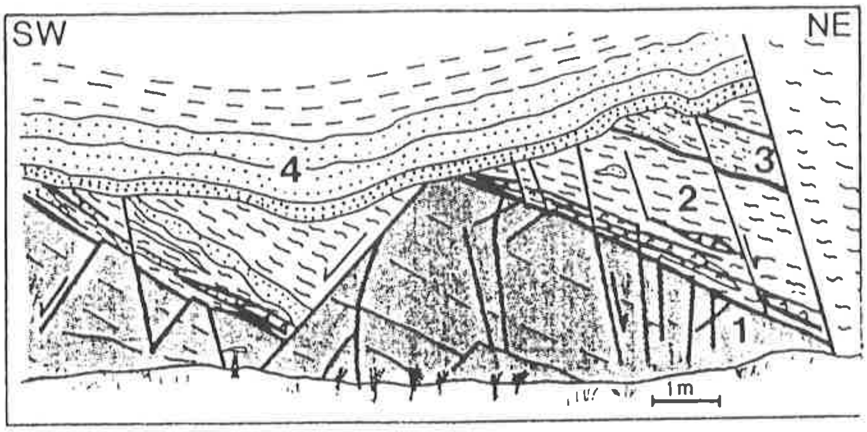

Fig. 9: Lég. voir page suivante.

Fig. 9: Coupe du tunnel de Fnidek montrant la transgression oligo-miocène. 1: flysch carbonifère (Akaili), 2: zone broyée (pélites paléozoiques et lentilles boudinées de dolomies triasiques), 3: grès rouges triasiques, 4: grès, marnes et conglomérats oligo-aquitaniens. CHALOUAN 1986.

3.1 Age de la distension

L'âge de cette distension pose un grand problème du fait de la confrontation des données radiométriques et micropaléontologiques. En effet, les failles normales recoupent les structures synmétamorphiques, la distension tardi-métamorphique peut être considéré comme post-oligocène supérieur, âge équivalent de la période située entre -22 et -25 Ma (âges isotopiques donnés par les différentes datations, mentionnées ci-dessus). Dans la région de Fnidek, ces failles normales sont scellées par la trangression oligo-micène inférieure qui correspond à un âge antérieur à -22,5 Ma. qui représente le sommet de NP 25 (HAQ & al., 1987). La vitesse du remonté d'un matériel chaud à la surface est en moyenne de 1 à 2Km/Ma (NIKONOV, 1989). Dans les Sebtides par exemple, il faut au moins, 5 Ma pour qu'elles remontent à la surface pour être érodées et transgressées par le matériel détritique aquitanien. Une des deux propositions suivantes peut être choisie pour résoudre le problème de l'âge de la distension : soit le métamorphisme alpin est d'âge plus ancien que 22 Ma, autour de 25 à 28 Ma., soit la base de la trangression est plus jeune.

4. CONCLUSION

L'étude structurale des terrains métamorphiques des Sebtides inférieures a permis de mettre en évidence des structures extensives tardi-métamorphique matérialisés par de flexures qui évoluent souvent vers des failles normales. Les cassures remplies par du matériel métamorphique et qui coupent la foliation principale "S_2", montrent que la distension s'est produite vers la fin du métamorphisme. La direction générale de cette distension est NE-SW à N-S. La présence de ses failles normales parallèles à la côte actuelle du Rif interne permet de confirmer que l'effondrement de bassin d'Alboran est pour l'essentiel d'âge Miocène moyen supérieur (CHALOUAN & al., 1989) et qui s'est poursuivi au ponto-plio-quaternaire (GLANGEAUD, 1961 et 1971, EL GHARBAOUI, 1981, ESTEVES & DEGALDEAN, 1983, MOREL, 1987). Cet effondrement a cependant débuté par une première distension antérieure à la transgression oligo-aquitanienne. Cette distension précoce intervient juste à la fin des paroxysme tectono-métamorphiques alpins ayant affectés les unités internes des chaînes bético-rifaines. Elle marque le début d'une extension régionale importante qui va être responsable de l'amincissement crustal au centre de l'arc de Gibraltar et de l'effondrement du bassin d'Alboran (AUZENDE et al., 1973, DURAND DELGA, 1980).

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THE PALEOZOIC GRANITIC PLUTONISM IN THE PRESENT ALPINE FOLD BELT: BASIC CONCEPTS AND A GENERAL SCHEME FOR SPACE AND TIME EVOLUTION.

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Abstract

A review of some basic concepts on granite geology is presented and discussed. Classifications based on modal and major-element geochemical compositions provide simple and useful tools for unravelling the nature of magmatic associations and geodynamic settings. Data collected in the Hercynian basement of the Alpine fold belt substantiate significant variations of rock types.
An irrevertible evolution in time from mantle-derived tholeiitic oceanic granites to hybrid orogenic calc-alkaline and crustal anatectic granitoids, then to mantle-derived subalkaline and alkaline granites is documented during Paleozoic times from the Cambrian-Ordovician up to the Permian-Triassic. Thus, granitic massifs illustrate dramatic changes in the geodynamic environments from initial oceanic spreading regime through subduction - obduction - collision processes to uplift and intra-continental distensive regimes, prelude to the Mesotethys oceanic basins opened during the Liassic.
The building of the Palaezoic orogens has been accompanied by the emplacement of large plutonic and/or volcanic massifs in which granitoids predominate. The magmatic activity appears to be continuous; however, the abundance and the compositional characteristics of plutonic massifs are strongly differing in space and time from the region to another.
These last years, several attempts have been made to define an environmental classification and nomenclature of granitoids. Such a classification and nomenclature would reflect the different types of belts, different mechanisms of crustal deformation, varying width and nature of the lower crust, different mechanisms for magma ascent, various sources of magmas and maybe even the possible multicycle genesis of granitoid magmas.
After ZWART (1967), PITCHER has first defined two main orogenic environments of granites in terms of Hercynotype and Andinotype, while a third type, the Alpinotype, is believed to rarely develop magmas. Subsequent refinements have lead to amore complete classification, based mainly on geological data (PITCHER, 1982, 1987): namely Pacific-type, Andinotype, Hercynotype and Caledonian-type, which all display granitoid batholiths, and Alpinotype, devoid of granitic rocks.
The bases for mapping and interpreting magmatic variations in terms of geodynamic setting are numerous: they may be petrographic ones (LAMEYRE and BOWDEN, 1982), chemical ones (STUSSI and DE LA ROCHE, 1984), mineralogical ones (e.g. zircon morphology, PUPIN, 1985) and isotopic ones (DUTHOU et al., 1984).
The covered area of the I.G.C.P. No276 is of special interest as it provides a unique example of alleged "Hercynotype" orogenic segments, later involved in an "Alpinotype" orogeny (even if the Alpine fold belt is not so devoid of granites as in its original definition...). Therefore, the Palaeozoic plutonites in the present Alpine fold belt can provide a good test for such a classification.
The aim of this paper is not to present a definitive scheme of granite repartition in the Palaeozoic basement of the Alpine fold belt. Actually, it is, first, to present some basic concepts on granite classifications and geoddynamical relationships and, second, to provide a general frame for space and time magmatic evolution in the Palaeozoic aera of basement zones that belong presently to the Alpine fold belt.

Thus, pertinent examples will not be fully discussed but we hope that this preliminary address will provoke and promote further works under auspices of the I.G.C.P. No276 Working Group on Granites, not only in the Alpine fold belt but also in all the area covered by the Project.

1. GRANITE CLASSIFICATION: A BRIEF REVIEW

Since 1967, the nomenclature of igneous rocks has been reexamined in detail by STRECKEISEN (1976) and his co-workers using the modal propositions of quartz, alkali feldspars, plagioclase (An content higher than 5) and feldspathoids. Although STRECKEISEN and his I.U.G.S. Committee never intended the modal Q A P F diagram to be used for any other purpose than systematic classification of *individual* rock types, LAMEYRE and BOWDEN (1982) have shown that this diagram is valuable for identifying various granitic series. These authors proposed the recognition of the following series (figure 1):
1/ anatectic mobilizates and related granitoids,
2/ calc-alkaline series and its variants,
3/ alkaline series,
4/ tholeiitic series.

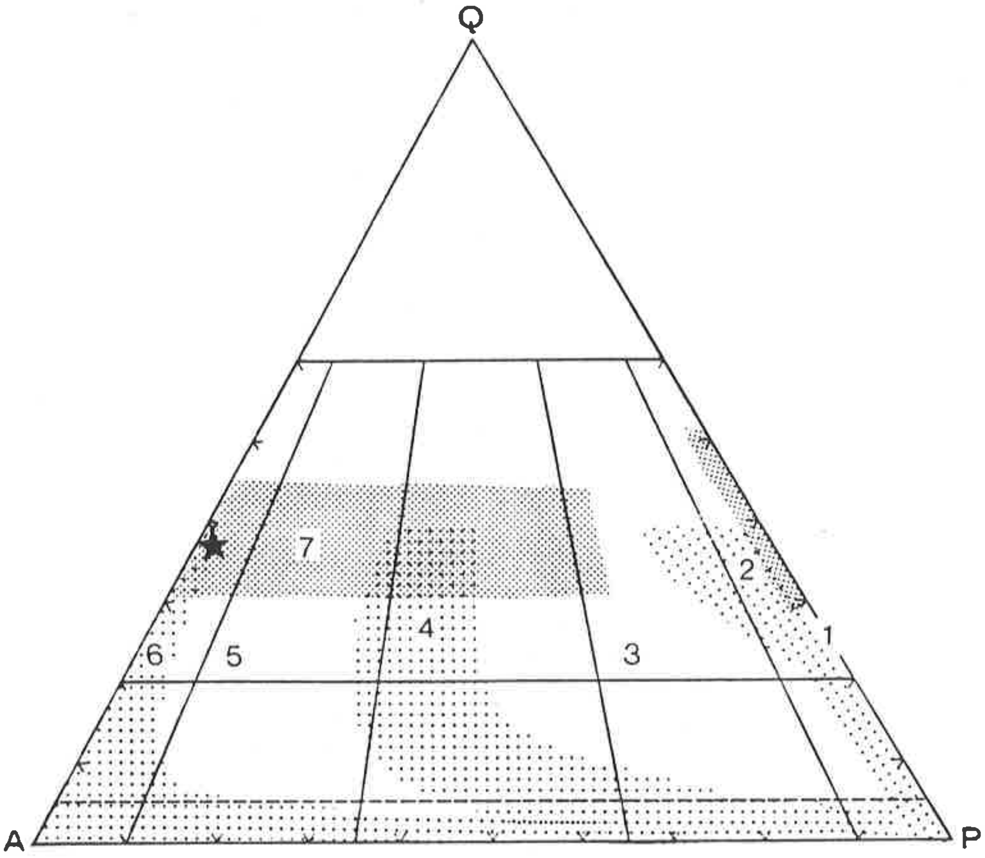

Fig. 1. Summary QAP diagram showing important fields of various granitoid series (greater than 20% vol. Q), discussed in the text: 1=tholeiitic; 2=calc-alkaline-trondhjemitic (low K); 3=calc-alkaline-granodioritic (medium K); 4=calc-alkaline-monzonitic (high K); 5=aluminous granitoids found in alkaline provinces; 6=alkaline and peralkaline; 7=overlapping field of granitoids formed by crustal fusion; black star=median minimum composition.

These series are represented by plutonic rock types as well as by their volcanic equivalents. The use of Q A P diagram by plotting modal data as well as normative data is very easily understable for any field geologist. This scheme has been proved useful to distinguish tectonic settings (anorogenic versus orogenic) (LAMEYRE, 1988; LAMEYRE et al., 1982) and possible source rocks (mantle versus crustal and hybrid) (DIDIER et al., 1982; BOWDEN et al., 1984).
More recently, several attempts have been made to couple the Q A P diagram, based on modal data, with other diagrams, based on major element analyses (STUSSI and DE LA ROCHE, 1984; BATCHELOR and BOWDEN, 1985). When plotted on multicationic diagrams (for a review, see DE LA ROCHE, 1986), granitoid rock compositions show a systematic change and result in a simple classification of granitoid series and their geodynamic setting in one given orogenic cycle (figure 2):
1/ pre-plate collision with calc-alkaline series,
2/ plate collision with calc-alkaline series,
3/ post-collision anduplift with high-K calc-alkaline series,
4/ late-orogenic stage with subalkaline monzonitic series.

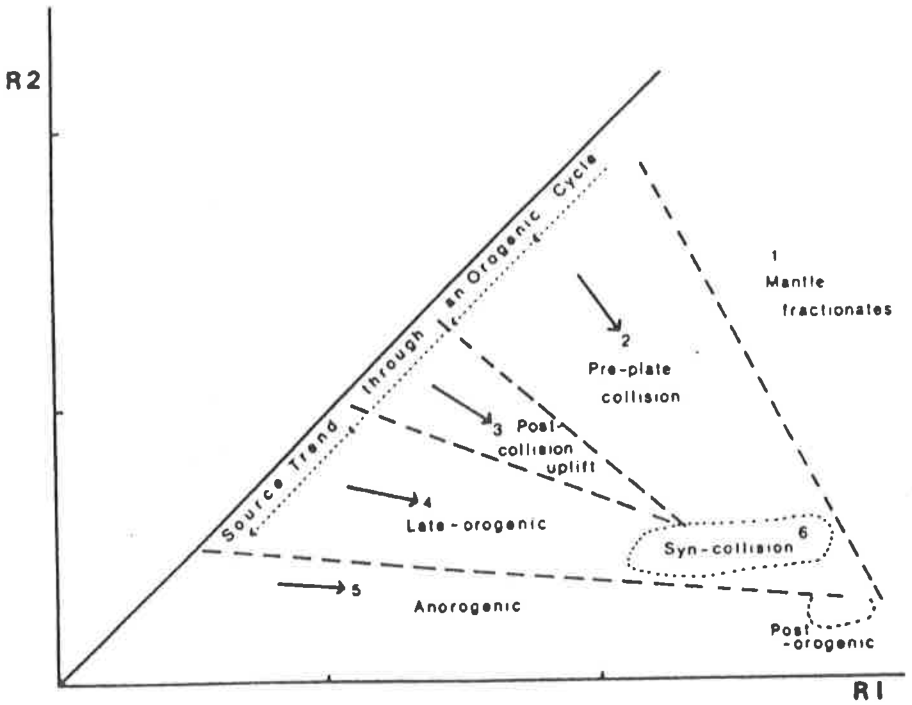

Fig. 2. Summary diagram of the major granitoid associations, based upon Pitcher (1979,1982) and Harris et al. (1983). Petrological equivalents (after Lameyre and bowden, 1982): group 1 - tholeiitic; group 2 - calc-alkaline and trondhjemitic; group 3 - high potassic calc-alkaline; group 4 - sub-alkaline monzonitic; group 5 - alkaline and peralkaline; group 6 - anatectic 2-mica leucogranites.

$R1 = 4Si - 11(Na + K) - 2(Fe + Ti)$ $R2 = 6Ca + 2Mg + Al$

Anorogenic alkaline granites come after, as a result of a relaxation effect after plate collision which would generate tensional rifts and/or pull-apart structures along large shear zones (BONIN and LAMEYRE, 1978; BLACK et al., 1985D).
In the Alpine fold belt, Palaeozoic plutonic rocks obviously pertain to all the granitoid series and therefore could indicate all the types of geodynamic settings that could appear successively in an orogenic cycle.

2. ZIRCON MORPHOLOGY AND MAGMATIC ZONING

The typlogical method, which defines zircon crystals on the basis of the relative development of prism faces (index T) and pyramid faces (index A), has been described previously (e.g. PUPIN, 1980, 1985). The use of the zircon method as a tool for the classification and interpretation of plutonites and volcanites have been improved during the last ten years. Together with chemical, geological and isotopic data, the following classification of granitoids was proposed (figure 3):
A. granites derived by crustal melting due to regional anatexis and/or induced melting by rising granitic bodies; this group comprises three subgroups (peraluminous leucogranites, monzogranites and granodiorites), all types bearing Al-rich silicates.
B. hybrid granites, which embrace a twofold subdivision: calc-alkaline granites (low-K, normal-K and high-K types) and K-subalkaline granites.
C. mainly mantle-derived granites, which also comprise two subgroup: alkaline granites and transitional-tholeiitic granites.

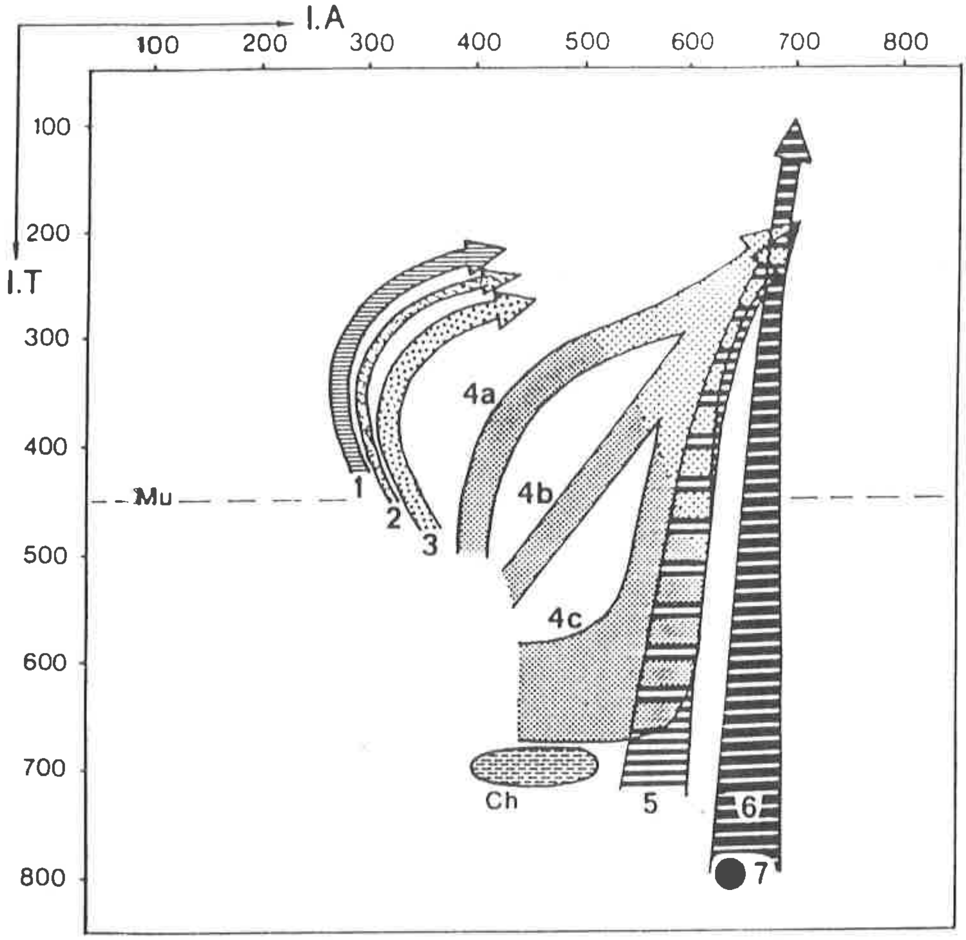

Fig. 3 Distribution of mean points and mean TET of zircon populations from: Granite of crustal or mainly crustal origin: (1) aluminous leucogranites; (2) (sub)autochtonous monzogranites and granodiorites; (3) intrusive aluminous monzogranites and granodiorites.- Granites of crustal+mantle origin, hybrid granites: (4 a, b, c) calc-alkaline series granites (*dark doted area*=granodiorites+monzogranites; *clear doted area*=monzogranites+alkaline granites); (5) sub-alkaline series granites. - Granites of mantle or mainly mantle origin: (6) alkaline series granites; (7) tholeiitic series granites. - *Mu*. limit of the muscovite granites (IT<450); *Ch*. magmatic charnockites area. If stock 4 exactly represent the repartition known up to now of the granites of the calc-alkaline series, stock 1, 2, 3, 5, 6 are the areas of the mean points greater frequencies of the corresponding zircon populations, graphically drawn. For stock 1 to 3, which overlap broadly each other, the *three curved bands 1-3* mark the relative positions of the areas of repartition of the mean points and of the global typological evolutionary trend of populations: the great majority of the mean points of the zircon populations of a stock lie on the corresponding curved line and inside its concavity.

PUPIN (1985) has presented an outline of magmatic zoning of orogenic granitoids and the following conclusions are arising from his comparative study:
1. two main groups of granitoids occur: anatectic crustal granites and hybrid granites.
2. the different subgroups of the anatectic crustal granites display approximately the same T values of their zircon populations (i.e. they have formed at nearly the same temperatures). Therefore, the differences between the subgroups must result from differences in their source rocks as well as the type of anatexis process.
3. hybrid granites also show a regular spatial distribution. From the suture zones, they begin by calc-alkaline (low- and normal-K) granites derived from water-rich magmas. These "hydrous" granites overlap or are followed by calc-alkaline (normal- and high-K) granites and by K-subalkaline granites derived from "hot" and "dry" magmas, high temperatures are indicated by the fairly high T values of their zircon populations, These last granites are associated with extremely active volcanism.
4. an overlap exists between the different types of crustal anatectic granites and hybrid granites, but in a general way the dry and high temperature calc-alkaline granites occur spatially behind the main zone of anatectic granites. The existence of anatectic granites in the area of hybrid granites may be explained by the induced melting due to rising magmas related to calc-alkaline granites.
5. the spatial zonation is characterized by the following succession of discrete granitic belts from the suture zone to the foreground:
- calc-alkaline granites derived from water-rich magmas,
- parautochthonous or rarely intrusive aluminous anatectic granites,
- intrusive peralumious anatectic granites,
- calc-alkaline granites derived from hot and dry magmas,
- K-subalkaline granites.
6. the intrusion of granites, notably calc-alkaline, is perhaps the result of a long process because geochronological data gathered in different orogens, from Precambrian to present ones, indicate that their emplacement occurs only after the end of subduction process in the zones analysed.
The late high-level intrusion of subduction-related magmas in clearly post-subduction periods has already been pointed out (SMITH, 1977). Magmatic zonings, that are interpreted as the result of subduction and collision processes, are still visible in a post-collision context, and indicates that the disappearance of the relict structures of geodynamic polarity requires a very long period of time.

3. GEOCHRONOLOGICAL INTRICACIES

If spatial distribution of different granitic series can be outlined by modal, mineralogical and major-element chemical data, the chronological distribution must also be considered.
The reconstitution of the magmatic events having affected the Alpine segment of the Palaeozoic orogens necessitates the use of a geological time scale. Two ways for establishing this time scale are commonly used and connected together: relative geological methods, based on field observations, and absolute isotopic methods, based on the closure concept of minerals and/or whole rock at determined temperatures. The two kinds of methods require strong constraints to give significant ages, interpretable in terms of geological events.
In the Alpine orogen, not all the plutonic bodies assigned to Palaeozoic magmatic events have been fully dated up to now. The spatial distribution and the accuracy of avalaible isotopic ages are highly heterogeneous and, unfortunately, isotopic ages are completely absent in some regions: in many places, an "Hercynian" age is inferred only from the Mesozoic or even Cainozoic sedimentary cover lying unconformably on the considered massifs.
Moreover, the following reasons make it difficult to interpret univocally the isotopic data:
1. many plutonic bodies have suffered at different degrees the effects of Alpine folding and associated matamorphic phases: some massifs located near Alpine suture zones are completely converted into orthogneissic lenses where magmatic features are only

recognizable at places, while others are affected by shearing zones with local retromorphic effects. In both cases, isotopic clocks have been reset more or less completely to zero. In this respect, many data concerning minerals must be used carefully, as their systems can be easily overprinted by different successive thermal events, some of which are related to late- to post-Hercynian magmatic events and the others to the Alpine history itself.
2. some pretectonic or synkinematic plutonic bodies could have been reset a more or less long time after their emplacement, due to subsequent tectonometamorphic events, and therefore indicate a younger age. The homogeneous ages obtained on a whole region may be interpreted as a result of subsequent overprinting during the culmination of thermal metamorphism and slow cooling homogeneization effects: this model has been substantiated in the Adrar des Iforas batholith for the Pan-African orogen (LIEGEOIS and BLACK, 1984). The same case may be repeated during the Phanerozoic orogenies, the late-orogenic magmatic events and their thermal aureole causing the complete resetting of isotopic clocks located in earlier plutonic bodies.
3. a geological control on the avalaible isotopic ages is often lacking, because of the scarcity of relative chronological criteria in the host rocks if plutonic bodies are emplaced at dephs in polymetamorphic belts. Many tectonometamorphic phases are not yet fully documented and dated. Moreover, in some cases, isotopic data are in conflict with geological evidences: they can display too much younger or, more rarely, too much older ages which cannot be easily interpreted.
4. in many multi-injection plutons, field evidences indicate a relative chronology of the different intrusive phases but the observed succession is not always substantiated by isotopic data, as the time intervals of emplacement are notably less than the analytical errors. In these cases, even if the isotopic ages seem similar, the field evidences should be preferred. Analytical errors as well as the previously described factors should be born in mind when comparing geochronological data on various granitoids series in contrasting regions.

3.1. Age distribution of the granitoid series.

When characterizing a magmatic suite, a description of the rock types making up the single massifs is not self-sufficient, and many aspects must be considered at the scale of the whole belt. The plutonic rocks are generally compositionally expanded, though, on the scale of one single pluton or in special domains, a single rock type may prevail or be exclusive.
For a more complete review and list of regional references, see I.G.C.P. n^{o} 5 final report to be published (BONIN et al., in press). The major conclusion is that the following granitoid series can be observed in the basement of the Alpine-Mediterranean fold belt:
- calc-alkaline series with its low-K and high-K variants,
- anatectic aluminous granites,
- alkaline series.

The major massifs belonging to thoeliitic series are clearly pre-Hercynian (Cambro-Ordovician ophiolitic units). However, some basic-acid pre-tectonic bodies have been related to tholeiitic series, interpreted as magmatic response to marginal basin opening and stratigraphically dated at Upper Visean.
Calc-alkaline massifs may be subdivided, according to their rock types and K-contents.
A. Low-K calc-alkaline suites, displaying diorites, tonalites and trondhjemites, are issued from water-rich magmas and their ages range at the Devonian-Dinantian boundary, i.e. very close to the plate collision event and just after the culmination of Acadian tectonometamorphic events.
B. Normal-K calc-alkaline suites are widespread in space and time and their time range covers the whole Carboniferous and the Lower Permian, i.e. from Upper Visean to the end of the Autunian. For this suite, the magmatic activity reaches a maximum at the Carboniferous-Permian boundary with the development of huge batholiths.
C. Lastly, the high-K calc-alkaline suites are nearly synchronous to and sometimes younger than normal-K suites. But they appear to be spatially restricted.

The anatectic series is compositionally very expanded, as it embraces autochthonous mobilizates as well as parautochthonous and intrusive aluminous granitic bodies. For this reason, the time related to regional anatexis, ranging from 360 to 330 Ma., and with allochthonous suite yielding ages from 330 to 270 Ma. Note that, in this last suite, aluminous monzogranites and granidiorites are as a rule significantly older than peraluminous leucogranites.
Unrelated to Hercynian orogenesis, the alkaline series predates and points to the subsequent Tethysian oceanic basin opening, prelude to the Alpine orogeny. Accordingly, alkaline complexes are ranging from Middle Permian to Upper Triassic.

3.2. Timing of magmatic activity in different domains.

The Palaeozoic orogens are so large and so wide that it can be supposed that magmatic activity has not taken place at the same time in the different domains. The following table indicates grossly the time span of magmatic (plutonic) activity for discrete zones of the Hercynian orogen subsequently affected by the Alpine orogeny.
This table has been compiled from data collected by numerous collaborators working for I.G.C.P. n^{o}5 and slightly corrected from more recent publications, respectively for the different covered areas: ROSE for Morocco, ENRIQUE for Pyrénées, BONIN for other French areas with the collaboration of VIVIER and OLIVER, AYRTON for Switzerland, VISONA for Italy and Austria with the collaboration of a team of 13 collaborators and KEKELIA for Caucasus. A complete list of references is to be found in BONIN et al., (in press).
From the examination of the table, it appears that magmatic activity was nearly continuous since the end of Devonian to Permian in most regions. The chronological gaps now existing for some areas are merely due to either the small number of datable plutonic occurences or the insufficiently represently representative set of data.
Table: averages of age distribution of plutonic activity (alkaline plutonic complexes are indicated in brackets).

Morocco:	from 350-360 Ma. to Ma. (270-250 Ma.)
Pyrénées:	325-335 Ma. and 300-270 Ma.
Montagne Noire:	350-330 Ma., 320 Ma. and 295-270 Ma.
Provence:	350 Ma. and 320-300 Ma.
Corsica:	from 340 Ma. to 290 Ma. (270-200 Ma.)
French Alps:	from 350 Ma. to 280 Ma. (280-260 Ma.)
Switzerland:	from 320 Ma. to 250 Ma. (270-250 Ma.)
Italy-Austria:	from 310 Ma. to 250 Ma. (270-240 Ma.)
Caucasus:	from 380 Ma. to 250 Ma.

Note however that in the Alpine mountain belt, especially in Austria, Italy and Switzerland, no age older than 320 Ma. is recorded and that the maximum of availaible data, within the analytical errors, is centred around 290-270 Ma., suggesting:
1/ that the culmination of magmatic activity occured at this time, as for many plutons, these isotopic ages agree with geological evidences,
2/ that other plutonic bodies, possibly older, have been completely rejuvenated by the associated thermal event.
A second maximum at ca. 250 Ma. is related to the emplacement of alkaline ring-complexes and associated volcanic formations, but this age range is also recorded by many mineral systems in previous calc-alkaline granitoids: this fact was interpreted by FERRARA and INNOCENTI (1974) as due to and important Triassic thermal event.
Thus, at least two thermal imprints are recorded by the isotopic clocks in the regions which were later strongly involved in whole Alpine orogenic events, whereas these imprints are not so evident in the other zones.

4. A COMPARISON WITH PITCHER'S OROGENIC TYPES.

In order to outline granite typology with reference to tectonic environment during the Hercynian orogeny, the revised classification proposed by PITCHER (1987) must be considered as a scientific tool. Subsequent refinements of this orogenic classification scheme along with the granite letter subdivision (I-, S-, A- and M-types, CHAPPELL and WHITE, 1974; LOISELLE and WONES, 1979; PITCHER, 1982) has resulted in a typology (PITCHER, 1987) which defines the characteristic rock types and their geodynamic settings:
- W. Pacific type: gabbro and subordinate M-type granitoids (plagiogranite), emplaced in volcanic and volcanoclastic aprons as small zoned plutons during an ocean-ocean subduction. The magmatic episode is short-lived and originates in partial melting of mantle-derived, metamorphosed underplate giving a hot and "dry" quartz-diorite magma.
- Andinotype: I-type tonalite, granodiorite, with gabbro, associated with andesites in great volume and amplaced as disharmonious, linear cauldron batholiths feeding volcanoes (BUSSELL et al., 1976) during an ocean-continental subduction. Magmatic episodes are long-lived and have their origin in partial melting of mantle-derived underplate, with some crustal contribution within continental lip, giving hot, "dry" tonalitic magma rising high into the crust.
- Hercynotype: migmatites, reworked as S-type granites, emplaced as harmonious diapir batholiths in early phase during an oblique continental collision (continent-continent "subduction"), with episodic recycling and partial melting of crustal materials by anatexis and reworking as batch-melts composed of relatively warm, "wet" granitic mush freezing at depths, with autometamorphic recrystallization processes.
- Caledonian-type: high-K calc-alkaline plutonic rocks with biotite granite, appinitic diorite and gabbro, emplaced as discordant plutons and distension diapirs during a rapid, post-closure uplift giving way to and adiabatic decompression and a relatively short-lived magmatic episode. Moderately hot and "dry", evolved, crystal-bearing magmas rising to various levels are issued from the partial melting of an old, tonalitic lower crust with mantle contribution.
- anorogenic Nigeria-type: biotite granite, alkali granite and syenite, emplaced as resurgent subsidence cauldrons during encratonic or post-orogenic rifting. This relatively short-lived magmatic episode is characterized by a relatively cool, fluidal magma, rising to near surface with sub-solidus crystallization, and originating in partial melting of old mantle or exhausted lower crust under anhydrous but F-rich conditions.

This classification scheme is now very popular and very often used to typify (meta)granitic formations (e.g. granite geochemical classification by PEARCE et al., 1984). Thorough examinations of magmatic features and of space-time distribution of Palaeozoic plutonic formations in the Alpine fold belt compared with these orogenic types lead to the following observations:

1. the earliest orogenic plutonic formations, such as Rioupéroux-Livet magmatic unit in the French External Crystalline Massifs, characterized by small dioritic-tonalitic bodies display apparently similar features as Andinotype and even W. Pacific-type magmatism. However, PUPIN (1985) has stressed relatively cool and "wet" conditions for magmas producing this low-K calc-alkaline suite, in disagreement with hot and "dry" conditions postulated for the formation of Andinotype plutonism.
2. the major part of granitic massifs is compositionally calc-alkaline, with a typical quartz diorite - tonalite - granodiorite - granite asociation (relatively low- to normal-potassium, magnetite-bearing plutonic suite which can be designated M- or I-types, according to PITCHER, 1982). Numerous examples can be correlated with similar massifs in the European foreland as well as in "African" (Southern Alps) promontory.

This magma series, therefore, could again indicate the Andinotype of orogeny. However, PITCHER et al., (1985), in their epilogue to the study of Andean batholith, have claimed that "a firm distinction should be made between Andean I-types and some others so designated, but which are significantly poorer in calcium and richer in potassium at the same level of SiO_2, and which were derived from a mixed source in a post-subduction regime" (page 289). The same authors have "urged caution in

interpreting this type of post-closure I-type plutonism in terms of an Andean plate-margin context" (op. cit.).
3. the anatectic plutonic formations, emplaced as autochthonous migmatic bodies and as well as later allochthonous diapirs, typically define the Hercynotype of orogeny. In Alpine regions, for example in Belledone and Vallorcine areas in the External Crystalline.
Massifs, or in Calabria - Sicily, they are usually accompanied by huge volumes of calc-alkaline products. Moreover, in some areas, anatectic massifs are missing or present in very subordinate amounts. Thus, characteristic Hercynotype plutonism may be lacking in the Hercynian basement.
4. with the exception of the earliest calc-alkaline plutonic massifs, which can be synchronous with ending ocean-continent subduction rapidly followed by continent-continent collision processes, calc-alkaline granite complexes, and especially the late- to post-orogenic plutonic-volcanic high-K massifs, have been emplaced during post-closure tectonic events and are likely to indicate the orogenic Caledionian type.
5. the alkaline Nigeria-type plutonic-volcanic complexes followed very closely the end of orogenic high-K calc-alkaline (Caledonian-type) and anatectic (Hercynotype) plutonisms. But they are distinctly post-orogenic and associated with extensional regimes accompanying large shear fault-zones. They seem to be more related to a pre-Tethys oceanic basin opening than to a waning orogenic stage.
6. propably, the most stricking feature of granitic plutonism is that is no significant difference between what happened in the western part of the Alpine belt, classically related to oceanic closure and to continent-continent collisional processes, and in the eastern segment, supposed to be a part of a continuously active continental margin during the Palaeozoic and Early Mesozoic times (DERCOURT et al., 1985).
As PITCHER (1982) has already pointed out, anatectic granitoids are likely to be accompanied by calc-alkaline massifs in Hercynotype context, as real thickening of the continental crust ensures that a variety of mantle-crustal components are available as source rocks. For the same author, the Caledonian-type is to be regarded as typical of the late stage of the orogenic process.
Therefore, PITCHER's orogenic types are likely to represent different stages of the same orogenic cycle. Magmatic products, related to the different types, are present in the Palaeozoic basement of the Alpine fold belt but in varying abundance according to their location and emplacement time.

5. CONCLUDING REMARKS.

In summary, the Palaeozoic granitic massifs emplaced in the Alpine orogen provide fine examples of evolving magma types. Eo-Alpine plate convergence has resulted in the juxtaposition of Palaeozoic segments which were not previously necessarily contiguous.
Probably the major difficulty in connecting granitoid belts with tectonometamorphic zones lies in the fact that the major folding episodes clearly predate the culmination of granitic magmatism. This fact can be encountered everywhere in different orogenic belts and especially in the Palaeozoic basement of the Alpine fold belt. In addition, it is not sure that the site of granitoid emplacement, mainly controlled by large lithospheric fault zones, has been previously the exact site of tectonometamorphic events. The additional fact that the magmatic activity is generally not synchronous with its geodynamic cause (subduction-collision processes) contributes to obscure the problem. Only when tectonometamorphic events will be entirely known, the genetic link between granitoid formation and global geotectonics will be fully understood.
Presently, it seems that there is a considerable time interval (about 30 Ma. or more) between the beginning and the end of certain geodynamic conditions and the beginning and the end of their magmatic responses (e.g. SMITH, 1977). The complexity of the plutonic history may be ascribed either to contrasting geodynamic environments, or differing continental microplates situated between large continental plates before their subsequent aggregation into the Permian Pangaea, or both.

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The list of references comprises only papers cited in the text. It must ont be considered as a compendium of what has been published, even recently, on the area of Palaeozoic granites in the Alpine fold belt. It is only an introduction to the world of references on granite classifications and assciated geodynamical environments.

N.B.: The following figures are taken from:

Figure 1: LAMEYRE and BOWDEN (1982) / Figure 2: BATCHELOR and BOWDEN (1985) / Figure 3: PUPIN (1980)

FLYSCH SEDIMENTATION RELATED TO THE VARISCAN OROGENY WITHIN THE CIRCUMMEDITERRANEAN MOUNTAIN BELTS

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Abstract

Devonian - Late Paleozoic clastic/volcanoclastic sequences of 52 Circummediterranean areas have been evaluated as potential Variscan flysch environments. As synorogenic sediments they reflect the Variscan orogeny, which varies in space, time and intensity. Flysch environments related to orogenic movements were recognized as followed:
1) Two flysch stages in the Alboran-Balearic-Trough: a prebretonic Devonian stage followed by an Upper Visean - ? Westfalian (preasturian) one.
2) Prebretonic metamorphosed flysch in the Western Carpathians, N-Dobrogea (Tulcea unit), Caucasus (Main Range Zone).
3) Presudetic flysch environments (d = Devonian, cu = Lower Carboniferous, m = metamorphosed): Eastern Alps (d, cu), Balkanids (d, cu), Eastern and Southern Carpathians (m: cu), Dobrogea-Macin unit (m: cu), Sardinia and Calabria/Sicily (cu-) Caucasian Fore Range Zone (d-cu).
4) Preasturian flysch environments:
Catalonian Coastal Ranges, Massif of Mouthoumet, Pyrenees, Mt. Noire, Southern Alps, ? Outer Dinarids.
5) Upper Carboniferous (Westfalian) - Permian: Hellenids.
Flysch of the "filling up" type (= without unconformable superposition): Bükk-unit, ? Jadar Paleozoic, Pontids.

1. INTRODUCTION

Devonian to Late Paleozoic siliciclastic sedimentary units interpreted as flysch are widely distributed within the alpine Circummediterranean mountain belts. As synorogenetic sediments they reflect the Variscan orogenetic movements. The change in tecto- and sedimentary environment from flysch to molasse sedimentation which is often marked by a clear unconformity, points to the culmination of Variscan orogenetic activities within the sedimentary units. Detailed stratigraphic and sedimentological analysis shows that these orogenetic activities were different in space, time and intensity. The presented paper is based on systematic data collections perfomed within IGCP No. 5 (published IGCP No. 5 Newsletters, Vol. 1-7) will give a rough overview on lithostratigraphic units which are well established or possible Variscan flysch units. An identification of these units as "flysch" was made by collecting and evaluating of critical flysch-criteria listed in tab. 1-5. An identification as flysch only on the basis of published data is often subjective, difficult or even impossible on account of the different level of sedimentological investigations, scarce biostratigraphic data and the widely scattered literature.
Nevertheless this data collection should provide a little mosaic by tracing out paleogeographic restorations within IGCP Project No. 276.
The area considered covers the total Circummediterranean area from Spain and Morocco in the West to the Caucasus-, Pontide- and Tauride-area in the East. Detailed information on the actual tectonic positions of these prealpine units can be taken from IGCP No. 5 geotraverse documentations (also published in IGCP No. 5 Newsletters).

2. DEFINITION OF FLYSCH

SEILACHER (1967) demonstrated in clear concept that typical flysch is characterized by tectonical, sedimentological and biological features. If one or two features are lacking, the silicicalstic rock sequence can only be classified as aberrant flysch or as flysch like sediment.
One of the most important features is the close relationship to an orogenetic event in space and time. Thus, flysch is a synorogenetic sediment of deeper water origin (indicated by typical solemarks and other sedimentary structures) and is composed of prevailing turbidites and debris intercalated in hemipelagic materials (see also SEILACHER, 1967, HSÜ, 1970, FÜCHTBAUER, 1988: 947).
Flysch occurs in two significant plate tectonic settings:
1. In a subduction regime, oceanic crust and sediments are transported to the trench, where they are buried by thick turbiditic flysch sediments. This sedimentation terminates after the incorporation of the sequences in an accretionary prism.
2. In a foredeep/foreland setting flysch is deposited on top of continental crust above shelf sediments in front of the prograding upper plate and an isostatically uplifting area. In intraplate settings a similar situation is represented in front of nappes. In a further step the axis of the trough is moving more externaly. Now in this "exogeosyncline" the marine regime is gradually replaced by a fluviatile one filled up by molasse sediments (FRISCH & LOESCHKE 1986).
It is worth noting that turbiditic sequences bound to passive continental margins (e.g. the recent Atlantic margin) are not flysch in its tectonic sense, because these sediments will not be incorporated in an orogen immediately after sedimentation (FÜCHTBAUER 1988:957).
Important criteria for recognizing flysch:

I. Tectonic features
- relationship to an orogenetic event
- great thickness (x 100 m)
- main source area of the clastics: uplifted orogenic systems.

II. Sedimentological features
- alternation of siliciclastic and hemipelagic materials
- gravitational mass transported sediments (olistolithes, olistostromes, turbidites, slumps etc.)
- sedimentologically well established turbidites
- specific solemarks
- low carbonatic content of autochthonous sediments
- conglomeratic levels with extra- and intrabasinal clasts (KELLING & HOLROYD 1978).

III. Biological features
- lack of autochtonous fossils
- typical ichnofossils
- transported allochthonous plants and reworked fossils.

Further important tools for recognizing the geotectonic setting are heavy mineral evolution paths (STATTEGGER 1986) and the geochemical character of volcanic associated with the flysch sediments.

3. DATA COLLECTION

A compact data collection of potential flysch units from 52 sedimentary to low grade metamorphic sequences which represent proven or possible Variscan flysch environments within the Circummediterranean is given in tab. 1-5.

LEGEND TO TAB. 1 - 5

column 1: Location

column 2: Name of formations and stratigraphic age

column 3: Lithology and thickness (in m)

p pelites	v volcanics
s sand-/siltstones	L autochthonous limestones
g graywackes	m marls
c conglomerates	l lydites
b breccias	

( ) subordinate + more than....m
META metamorphic sequence

column 4: Volcanics

a acid	A alkaline	F lava flows
b basic	CA calkalcaline	P pyroclastites
i intermediate	Th tholeitic	D dykes
		s sills
		Pi pillows

column 5: Gravitational mass transported sediments

t turbidites in general
T turbidites classified to BOUMA sequences
Tp, i, d proximal, intermediate, distal turbidites
O olistolithes
Os olistostromes
S slumpings
M mudflows
D debris flows

FU fining upwards sequences
CU coarsening upward sequences

column 6: Age and composition of allochthonous materials

0 Precambrian	La allodapic limestones
1 Cambrian	L limestones
2 Ordovician	D dolomites
3 Silurian	l lydites
4 Devonian	p pelites
5 Lower Carboniferous	c sandstones/siltstones
6 Upper Carboniferous	g graywackes
7 Permian	v volcanics
8 Triassic	S sediments in general
	G granitic rocks
. materials of conglomerates, sandstones	M metamorphic rocks
* olistolithic blocks	Q quartz

column 7: Direction of sedimentary transport

N: from north to south

column 8: Heavy minerals

Heavy mineral evolutionary path (STATTEGGER 1986):

C Chios type B Betic Cordillera type P Pyrenean type
* other data available

column 9: Fossils (partely reworked)

E macrofossils
M microfossils
P plants
I ichnofossils

column 10: Position within the Variscan cycle

a) base:
no. = stratigraphic age (see also column 6)
l Lower m Middle u Upper
P sediments of preflysch stage
L lydites
C carbonates
x clastics in general, p pelites, s sandstones
V volcanics
f.e.: 4uPC = Upper Devonian carbonatic preflysch sediments

b) superposition
M continental sediments
S marine shallow water sediments

continuous sequences —x— tectonic boundary
→ angular unconformity

column 11: Classification of clastic sequences and depositional interpretation

F well established flysch according to SEILACHER 1968	f flyschlike sediments
F(s) sedimentologically atypic flysch	c clastic sequence in general
F(t) tectonically aptypic flysch	V volcanoclastic sequence in general

column 12: References

TAB. 1

Region	1 LOCATION	No.	2 NAME of FORMATION STRATIGRAPHIC AGE	3 LITHOLOGY THICKNESS	4 VOLCA-NICS	5 GRAVIT. TRANSP. SED.	6 ALLOCHTHONOUS MATERIALS	7 DIR. SED. TRANSP.	8 HEAVY M.	9 FOSSILS	10 SUPER-POSITION BASE	11 CLASSIFICATION DEPOSITIONAL INTERPRETATION	12 REFERENCES
INTERNAL RIF (GHOMARID NAPPES)	Aakaili-nappe	1c	"Culm-flysch": fine flysch *U. Visean-Namurian*	sgp(cl) 500-600		Td MOs FU	.Q1LGM *L56	S	• KORNPROBST 1976	M	8mM	F Distal turbiditic flysch.	MILLIARD 1959, BOURROUILH 1977, CHALOUAN 1987
		1b	"Culm-flysch": culm-flysch-molassic *Visean*	sg(Lm) 200-250		T						F(s)? melange. Post orogenetic sediments with molasse affinities and fining upwards turbiditic sequences	CHALOUAN 1987
		1a	Calizas alabeadas "Devonian flysch" *Devonian*	gmLp(l) 500-600		Tid	La			M	41PCLV	Distal to intermediate flysch facies in a subsiding oceanic basin or trench.	
	Koudiet Tiziane-nappe	2d	"Culm-flysch" *Visean-Namurian*: sandy-pelitic flysch with conglomerates	sgcp 300		t		S			8mM	F	BOURROUILH 1977 CHALOUAN 1987
		2c	marly-limy flysch	sgm(v) +50	bS	t	La					F	
		2b	sandy pelitic flysch	sgp(Llv) +100		t				M		F	
		2a	"Devonian flysch" Fm. of sandy-pelitic turbidites and gravel bodies *Devonian*	gsp(cL) 600-700		Tpi	La .SMV			EM I	3Px	F Proximal to intermediate flysch facies in a subsiding oceanic basin or trench.	
	Beni Hozmar-nappe	3c	"Culm-flysch" *Visean-Namurian*: fine flysch	spc(L) +150		Td FU		S			8mM	F Positive sequential unit from a proximal to a distal flysch.	BOURROUILH 1977 CHALOUAN 1987
		3b	conglomerate	c(spv) 100	bD	t	.S345MG					F	
		3a	micro-conglomerate with olistostroms	sgp(cl) 200		Tpd OsS FU	La *L5			M	4uPC	F Cyclic sequential unit positive - negative - positive	
	Talembote-nappe	4	"Culm-flysch" Wildflysch and olistostroms *Upper Visean*	gpc 100-200		OM	*L45	S			T —x—x— —x—x— T	F Proximal "wildflysch" channel in turbiditic fan.	CHALOUAN 1987
INTERNAL KABYLEAN MASSIF	Massif of Chenoua	5b	"Culm flysch" *?Visean-Namurian*	spc		M		S			78M 5PL	F Lyditic "preflysch".	BOURROUILH 1982 LEPRIER & BOURROUILH 1986
		5a	"Devonian flysch" *Devonian*	sp(v)	b	T M	La			E	5PL —x—x— T	F Oceanic flysch environment.	
	Grand Kabylie	6b	"Culm flysch" *?Visean-Namurian*	spc							78M 5PL	F	BOURROUILH 1982 BOULLIN & BOURROUILH 1986
		6a	"Devonian flysch" *Devonian*			M	*VL			EP	5PL 4uFp	F Oceanic flysch environment.	
	Little Kabylie	7	"Culm flysch" *Upper Tournaisian-Namurian*	psc(L)			.SQMG			M	78M 41Px	F	BOURROUILH 1982 BOULLIN & PERRET 1982 BOULLIN & BOURROUILH 1986

TAB. 2 LOCATION [1]		No.	NAME of FORMATION STRATIGRAPHIC AGE [2]	LITHOLOGY THICKNESS [3]	VOLCA-NICS [4]	GRAVIT. TRANSP. SED. [5]	ALLOCHTHONOUS MATERIALS [6]	DIR.SED TRANSP. [7]	HEAVY M. [8]	FOSSILS [9]	SUPER-POSITION BASE [10]	CLASSIFICATION DEPOSITIONAL INTERPRETATION [11]	REFERENCES [12]
BALEARIC ISLANDS	Menorca	8b	"Culm flysch" *Namurian*	sgc x.100		t MS	.MGVL456	S	P	M	M 5PL	F Deep sea fan with channel fillings.	BOURROUILH 1973 BOURROUILH & LYS 1976
		8a	"Devonian flysch" *?Middle-U. Devonian*	psc(vL) x.100	bS	t S	.L4p4c4V Q	S	•	EM	?4u5PL 41-mP	F Deep sea fan with channel fillings.	STATTEGGER 1979 BUCHROITHNER et al. 1980a
SPAIN	Betic Cordillera	9c	Marabella-Fm. *Namurian-?L.Westfalian*	csbg 100		DO?	.L56CQGMV *L?			EM	8M	Feeder channels or main upper fan channels. (F related to a collisional divergent stage.)	BOURROUILH 1976 BUCHROITHNER et al. 1980b HERBIG 1984 STATTEGGER 1989 STATTEGGER & HERBIG 1989
		9b	Almogia Fm.: Olive shales *Visean-Namurian*	p(g) 150				S	B	M		Sediments on a continental slope.	
		9a	Almogia Fm.: Retamares Member *Visean*	gc 40/60- +280		OD	.L45CVQGM *L34				5PL	Braided suprafan at the toe of a continental slope.	
		9	Santi Petri-Fm. *M. Silurian-* *U. Devonian*	Lg 200-500		T O	*L34			EM	5P1 2/3Pp	F Calcareous flysch sequence of a passiv continental margin in a divergent setting.	
	Catalonian Coastal Ranges	10	"culm flysch" *Upper Visean-* *? L. Westfalian*	sgpc(v) 1500-2000	b	t SO	.1GM		P	PI	5PCL	F	JULIVERT & MARTINEZ 1980 STATTEGGER 1980 JULIVERT & DURAN 1983
PYRENEES	Western part	11	Carbone detritique culm-flysch *Namurian C-Westfalian*	sgpc 300-500		O	.1MQ			MP	67M 61PCL	F	BOURROUILH 1983 ENGEL 1984
	Eastern part	12	Carbone detritique culm-flysch *Upper Visean-Westfalian*	sgpc 1000		T OSMD	.S1-5QVM		P	IP	67M 5PCL	F Base of slope - inner-fan deposits, locally with channel fillings.	BUCHROITHNER & MILAN 1977 MILAN & BUCHROITHNER 1978 STATTEGGER 1978, MUNOZ et al. 1983, BARROUQUERE et al. 1983, RAYMOND 1983 ENGEL 1984
S-FRANCE	Massif of Mouthoumet	13	Carbone detritique culm flysch *Upper Visean-Westfalian*	sgpc x.100		t O				EMPI	6M 5PCL	F Synorogenetic flysch deposits.	BARROUQUERE et al. 1983 ENGEL 1984
	Mt. Noire	14	Carbone detritique culm flysch *Upper Visean-Namurian*	sgpcL +4000 m		T $OOsSMD_{Cu}$	La5 .C12L+D14515 *L451C56V2	N		EMPI	6M 5PCL	F Submarine turbiditic fan.	ENGEL et al. 1978 ENGEL 1984 FEIST & FLAJS 1987
EASTERN ALPS (Grauwackenzone)	W Zell/See	15	Schattberg-Fm. *M. Devonian-* *?L. Carboniferous*	META ps(cb) 450		tpd O	MG				M6 4mPCVp	F (t) Proximal flysch environm. in a channelized fan. Transformation of a passive margin to a synorogenetic flysch basin.	HEINISCH et al. 1987 HEINISCH 1988 SCHLAEGEL-BLAUT & HEINISCH 1989
	Noric nappe Eisenerz area	16	Eisenerz-Fm. *Visean-? Namurian*	META ps(1v) 100-150							7M 4/5PC	VC	NIEVOLL 1983, 1987
	Noric nappe/E	17	Stocker-Unit *? Lower Carboniferous*	META ps(v1) 600	aFP	S			•		M7 4/5PC	VC	SCHÖNLAUB et al. 1980 SCHÖNLAUB 1982
	Gurktal thrust system/Stolzalpe nappe	18	Flysch *Visean*	sgp(1) 40-50		O	.1VM *L4			M	M6 5PLC	F Possible part of a flysch sequence.	NEUBAUER & PISTOTNIK 1984 NEUBAUER & HERZOG 1985

TAB. 3 LOCATION [1]	No.	NAME of FORMATION STRATIGRAPHIC AGE [2]	LITHOLOGY THICKNESS [3]	VOLCA-NICS [4]	GRAVIT. TRANSP. SED. [5]	ALLOCHTHONOUS MATERIALS [6]	DIR.SED TRANSP. [7]	HEAVY M. [8]	FOSSILS [9]	SUPER-POSITION BASE [10]	CLASSIFICATION AND DEPOSITIONAL INTERPRETATION [11]	REFERENCES [12]
SOUTHERN ALPS	19	Dimon-Fm. Floritz-Fm. Hochwipfel-Fm. *M. Visean-Westfalian*	sgc(bl) 600	abiFPPi ACA	Tp-d FU OOsSM	.L34515MVQG •L45	N	B	EMPI	6S 5PLC	F in a channelized fan in context with an oblique slip setting.	TESSENSOHN 1971, EBNER et al. 1980, 1990, SPALLETTA et al. 1980, FLORA et al. 1983, VAI & ROSSI 1986, SPALLETTA & VENTURINI 1988
TUSCANY/ELBA	20	-quartzites and upper phyllites of Apuane Alps -Buti-phyllites -Ortano-Fm./Elba *? Lower-Middle Carboniferous* sensu VAI	META sgpV x.100	ab	?O	•?L3c			?M	6S,6/7M P	- Hypothetic Variscan flysch sensu VAI. - Silurian-Devonian Preflysch environment sensu BAGNOLI et al.	VAI in COCOZZA et al. 1974 BAGNOLI et al. 1978 VAI 1978 COCOZZA et al. 1987
SARDINIA	21	parts of "Postgod-landiano" *Visean*	META sgc x.10			.L41C			M	6M 5PC	Possibly base of a flysch sequence.	SPALLETTA 1982 BARCA et al. 1986
CALABRIA	22	Pazzano Phyllites and graywackes *L.- ? U.Carboniferous*	META gp(Vb) x.100	b	T	.1			P	5PL	F	PICCARETTA et al. 1983 MAJESTE-MENJOULAS et al.1984 SPALLETTA & VAI (in press)
SICILY Peloritani Mts.	23	slates, metaarenites, conglomerates *L.- ? U.Carboniferous*	META gpV +500	abCAA	?T					8M 5PL	F	PICCARETTA et al. 1983 SPALLETTA & VAI (in press)
WESTERN CARPATHIANS: South Gemeric Zone	24	Gelnica Group (upper part) *Silurian-M. Devonian*	META psv(1L) +1000	abiPCA	Tp-d MS	.1pv		•	M	7M	F - Vc (oceanic environment).	GRECULA et al. 1981
WESTERN CARPATHIANS: North Gemeric Zone	25	Racovec-Group (upper part) *Silurian-? Tournaisian*	META psV +1500	baFPiP Th					M	56S	Vc oceanic environment.	GRECULA et al. 1981 GRECULA & HOVORKA 1987
WESTERN CARPATHIANS: Brusnik-anticline	26	Borehole BRU 1 1-600 m *Namurian B*	META ps(bv) x.100	aP	T O	.V1pcM •L6			M	T T	F	EBNER et al. 1990
BÜKK-UNIT: Szendrö Mts.	27	Szendrö-phyllit-Fm. *Namurian B - Westfalian*	META psg 500-600		t O	La •L456			M	61PC	F (t)	KOVACS & PERO 1983a, b KOVACS et al. 1983 ARKAI 1983 EBNER et al. 1990 KOVACS (in press)
BÜKK-UNIT: Uppony Mts. Talpolcsany-Unit	28	Eleskö-Fm.	META ps 50-100		O	•L4			M		Part of F (t).	
BÜKK-UNIT: Bükk Mts.	29	Silvasvarad-Fm.	META ps(c) 1100		O	•1				6S	F (t) "Filling up type".	

Tab. 4

LOCATION	No.	NAME OF FORMATION STRATIGRAPHIC AGE [2]	LITHOLOGY THICKNESS [3]	VOLCA-NICS [4]	GRAVIT. TRANSP. SED. [5]	ALLOCHTHONOUS MATERIALS [6]	DIR.SED TRANSP. [7]	HEAVY M. [8]	FOSSILS [9]	SUPER-POSITION BASE [10]	CLASSIFICATION DEPOSITIONAL INTERPRETATION [11]	REFERENCES [12]
EASTERN CARPATHIANS: Buccovina nappe	30	Kuznizkaya-Fm. *Lower Carboniferous*	sm 150–200	ab b						6M → 1	C (? F)	RUDAKOV 1980
Infrabuccovina nappe	31	Buhaescu-Fm. Fata Mutelui-Fm. *Lower Carboniferous*	META psV(L) 100–150 psV 550	ab b						6M → 4uPC	V (? F)	KRÄUTNER 1983, 1987
SOUTHERN CARPATHIANS: Supragethic-nappe (Poiana Rusca)	32	Upper Pades-Grp. *Lower Carboniferous*	META sgV	ab						6M 51PC	V (? F)	
Supragethic-nappe (Locva)	33	Lescovita-Grp. *Devonian-Lower Carboniferous*	META psV(Lc) x.100	baP					M	6M → 3xV	V (? F)	NASTASEANU & KRÄUTNER 1983 KRÄUTNER 1987
Danubian Unit	34	Sevastru-Fm. *Visean*	META sgcv	b						6M → 51PC	V (? F)	
APUSENI MTS. a) Biharia-/ b) Codru-nappe	35	a)Upper Paiuseni-Group b)Arieseniphyllit-Fm.	META sgcv	b						6M →	V (C)	KRÄUTNER 1987
BALKANIDS: Stara Planina	36b	Diabase-Phyllitoid-Fm.: Dulgi Djal-Group *Ordovician-Devonian*	cspv 600–900	bi						6M →	VC sensu VAI: Hypothetic Carboniferous flysch !	HAIDUTOV et al. 1967 VAI (in press)
	36a	Bercovica-Group *Preordov.-Ordovician* sensu VAI: *Middle Carboniferous!*	psmv 1340	bFP								
	36c	Stara Planina flyschoid-Fm. (Pescocnica-Rajanovci-, Stacevci-Fm.) *Devonian-Carboniferous*	psc			.G5				6M →	F	HAIDUTOV et al. 1967 TENCHOV & JANEV (m.s.)
	36d	Katina-Fm. *?Middle Devonian-U. Devonian*	ps(Ll) 900						MP	6M 23/4uP	F	SPASSOV et al. 1978 SPASSOV 1983
Kraistiden b) E a) W	37	b) Katina-Fm. *M. Devonian-L. Carboniferous* a) Tranovdol-Fm. *M.Devonian-U.Devonian*	ps(cL) b) 900 a) 350					b) B	EMP	6M → a)4u5cL —X— 41-mPCx	F	SPASSOV et al. 1978 SPASSOV 1983 STATTEGGER 1986
Kucaj-Unit	38	Kucaj-Zvonce flysch *Middle Devonian-L. Carboniferous*	gcp(Ll) +550		Tp-d OOsSD	.cpMGlvQ .c2p/13p4 •L345C4plc	NNW NNE NE	•	EMPI	6M → 41PCp	F Related to a fan complex (slope-outer fan); mostly retrograde (in Rtanj and Suva Planina prograde).	MASLAREVIC & KRSTIC 1985a,b KRSTIC & MASLAREVIC 1987a,b

Tab. 5

Region	LOCATION [1]	No.	NAME OF FORMATION STRATIGRAPHIC AGE [2]	LITHOLOGY THICKNESS [3]	VOLCA-NICS [4]	GRAVIT. TRANSP. SED. [5]	ALLOCHTHONOUS MATERIALS [6]	DIR. SED. TRANSP. [7]	HEAVY M. [8]	FOSSILS [9]	SUPER-POSITION BASE [10]	CLASSIFICATION DEPOSITIONAL INTERPRETATION [11]	REFERENCES [12]
DINARIDS INNER	Vadar-Zone	39	Veles series *Devonian-Carboniferous*	META spVL x.100	bF					M		V (oceanic environment)	RAMOVS et al. 1984 KRSTIC et al. 1989
DINARIDS INNER	Drina Golija	40	sandstone-schist unit sensu RAMOVS et al. 1984 *Devonian- M. Carboniferous*	META spv(L) x.100	abiF					M	78M → 3PxC	V (? F)	RAMOVS et al. 1984
DINARIDS INNER	Jadar Paleozoic	41	Sandstone-schist Complex *Lower Carboniferous- Namurian*	sgcL x.100		OOs	•L45			EMI	6S → 51S 4uPc	? F "Filling up type"	FILIPOVIC 1974 STEVANOVIC & VESELINOVIC 1978 RAMOVS et al. 1984
DINARIDS INNER	SE-Bosnia Praca-area	42	Praca and Drina Beds *Lower-Middle Carboniferous*	sgpc 1100-1500		Td OOsD	.L5?61gcv •L345			EMPI	7S → 51PCL	F	DIMITRIJEVIC & DIMITRIJEVIC 1972, RAMOVS et al. 1984, KRSTIC et al. 1988
DINARIDS OUTER	Central Bosnia	43	Bojska-Fm. *?L.-U. Carboniferous*	spv 800	a						7S → 4uPC	V, C	KULENOVIĆ 1983 RAMOVS et al. 1984
DINARIDS OUTER	NW-Bosnia Sana-Una region	44	Sana Una unit *Lower Carboniferous- ? Westfalian*	sgp(VL) 1000	abP	?O	•?L56			EMPI	7S ? 4uPxC	(V, C) F (s)	KULENOVIĆ 1983 NAJDENOVSKY et al. 1983 RAMOVS et al. 1984
DOBROGEA	N-Dobrogea Macin-Unit	45	Carapelit-Fm. *Lower Carboniferous*	psgv 1700	b					P	6M —x—x— 4mPC	VC (? F)	MIRAUTA 1983 BELOV et al. (in press)
DOBROGEA	N-Dobrogea Tulcea unit	46	Bestepe-Fm. *Devonian*	META psv(L)	ab					EM	7M → 3PpC	flyschlike V	MIRAUTA 1983
HELLENIDS	Attika, Salamis, Chios autochth.	47	olistostrome unit terrigenous flysch-fm.	sgcp(vLb) +3000	bDP	t OS CU	La •L3456		C	EMP	8S ?	F	PAPANIKOLAOU & SIDERIS 1983a, b STATTEGGER 1984
HELLENIDS	Chios allochthonous	48	flysch type-fm. *Upper Carboniferous- Lower Permian*	spL x.100		?t	La		C	EM	7S —x—x— T	sF (V)	PAPANIKOLAOU & SIDERIS 1983a, b STATTEGGER 1984
PONTIDS	Istanbul-Zone	49	Trakya-Flysch Unit *Upper Tournaisian- Middle Visean*	psgc(Lb) 2000		Tp-d	.MG15	NW		EMPI	S56 51PL	F (t) "Filling up type"	KAYA 1969, 1978a, b KONAK et al. 1985
GREAT CAUCASUS	Main Range-Zone	50	Lashtraks kaya-Suite *Lower Carboniferous*	META ps(LV) 500	biF					M	6M → 4uPC	? F	BELOV et al. 1978 ADAMIA et al. 1980
GREAT CAUCASUS	Forerange-Zone	51	"Famennian"-Tournaisian" olistolithe sequence *U.Devonian-L.Carbonif.*	psc(L) +1100		O CU	•L45			EM	5/6M → 4m/uPxVp	F	BELOV et al. 1978 ADAMIA et al. 1980
SCYTHIAN PLATFORM	Precaucasian-Zone	52	*Middle Devonian- Westfalian*	META psc(lLv) +2000	abF					MP	7M → 4mPp	Vc	BELOV et al. 1978 ADAMIA et al. 1980

4. TOOLS FOR PALEOGEOGRAPHIC AND GEODYNAMIC RESTORATIONS

4.1 *Breccias, conglomerates, olistolithes:*

Following KELLING & HOLROYD (1978) extrabasinal clasts (magmatites, metamorphites, older sediments) derive from a source area outside the sedimentary basin. On the other hand intrabasinal reworked sediments/volcanics are +or- contemporaneous with the flysch. Together with olistolitic blocks they reflect the composition of the adjacent marginal shelf area and continental margin.
Excellent studies on conglomeratic/olistolithic horizons involving paleogeographic restorations were perfomed by BOURROUILH 1977, 1979, 1980, BOURROUILH & LYS 1976, BUCHROITHNER & al., 1980 a, b, HERBIG, 1984, ENGEL & al., 1978 in the Westmediterranean area, by TESSENSOHN 1971, SPALLETTA & al., 1980 and SPALLETTA & VENTURINI 1988, in the Southern Alps, by MASLAREVIC & KRSTIC 1985 a, b, in the Balcanid Kucaj-Zone, and by KRSTIC & al., 1988, in the SE Bosnian Praca area.
Further high amounts of feldspar in the graywackes could point to the erosion of a crystalline complex in the hinterland (ENGEL & al., 1978, FEIST & FLAJS 1987:24).
Lydites/lydite breccias, and conglomerates are usually found in the basinal parts or just below the Carboniferous flysch. SPALLETA 1984, recognized two types of mud-supported breccias and conglomerates:
- monomictic, angular, grey - black chert breccias,
- polymictic conglomerates with rounded cherts, granites and metamorphites.

The angular monomictic (intrabasinal) breccias suggest that brecciation took place in a submarine environment just before the material was involded into a debris flow.
On the other hand rounded extrabasinal materials of the second type point to a reworking in a litoral of fluviatile regime, implying uplift of the source area which exposed the hinterland followed by rapid drowning. This trend is proven by rounded granitic and metamorphic pebbles for the Alboran Balearic Trough, Catalonian Coastal Ranges, Pyrenees, Southern Alps, Stolzalpe Nappe of Gurktal thrust system (Eastern Alps), Balcanids and the Istanbul-Zone of Pontids.

4.2 *Heavy mineral spectra:*

The evolutionary paths of heavy mineral spectra demonstrate different plate tectonic settings of the Circummediterranean Variscan orogenetic clastics.
a) The Chios type (Chios allochthonous and autochthonous; STATTEGGER, 1984) indicate a convergent stage with anaccretionary prism including oceanic crust (source of chrome-spinel) which is followed by a collisional fold-thrust-belt and basement uplift (source of metamorphics and felsic plutonics).
b) The Beric Cordillera type (STATTEGGER, 1982, HERBIG & STATTEGGER, 1989) reflects a divergent cratonic stage with basement uplift (apatite, zircon, some garnet) and short periods of convergence (rare chrome-spinel; garnet point to a metamorphic core complexe in the hinterland). This evolutionary path was also recognized in the Carnic Alps (FENNINGER & STATTEGGER 1977, SCHNABL 1976) and the Kraistids (STATTEGGER 1986).
c) The Pyrenean type (STATTEGGER 1978, 1982) with apatite, idiomorphic zircon and ZTR minerals indicates a stabilized cratonic hinterland strongly uplifted due to granitic emplacements. Similar spectra were found in Menorca (STATTEGGER 1979) and the Catalonian Coastal Ranges (STATTEGGER 1980).

4.3 *Volcanic rocks:*

In non-metamorphosed areas beside volcanogenetic detritus and blocks (see tab. 1-5) effusive materials are sometimes also found in situ within the flysch sequences. The best information is available from the Carnic Alps. There the Floriz-Fm. starts within the flysch sequence above the Middle Visean "Flora der Marinelli-Hütte" (AMERON et al.,

1984). It includes volcanics and distal volcanogenetic influenced turbidites. The above following Dimon-Fm. is composed of volcanic breccias, massive lavas, pillow lavas (breccias), hyaloclastites, tuffs, agglomerates, and intercalations of reddish to greenish shales and volcanogenetic influenced shales (= proximal and distal volcanogenetic flysch is at least 200 m (GENTILI & PELLIZER 1964, SPALLETTA & al., 1980). First investigations on trace elements point to calcalkaline and alkaline character of the magmas (ROSSI & VAI 1986).
In areas (Tuscany, Carpathians, Inner Dinarids, Dobrogea, Great Caucasus/Main Range Zone, Scythian Platform) in which +or- metamorphosed clastic occur in a "flysch"-position these sequences bear volcanogenetic rocks of different geochemical character.
In general, more chemical and trace element data would be necessary to recognize the geotectonic settings in a more detailed way.

5. FLYSCH RELATED TO SUBMARINE FAN DEPOSITS

By means of sedimentological criteria the flysch environment of many localities (Alboran Balearic Trough, Pyrenees, Mt. Noire, Carnic Alps, Balcanids) can be recognized as clastic submarine fan deposits with different turbiditic facies sensu WALKER & MUTTI 1973. Often the intercalation of turbidites and olistolithic blocks points to a positive progressive or a coarsening upward trend by approaching the orogenetic phase in space and time. The most beautiful example is represented in the Mt. Noire (S-France) where the turbiditic facies leads to olistostromes and olistolithes finally culminating in the "Escailles de Cabriere" which are interpreted as chaotic gliding synsedimentary nappes (ENGEL et al., 1978).

6. REGIONAL ASPECTS

The stratigraphic age of the flysch sequences considered is very important for tracing out differences in space and time of the Variscan orogenesis. In this context, all sedimentologically well established non metamorphosed flysch sequences and metamorphosed clastic/volcanoclastic piles which take in a "flysch position" (= at the top of a preflysch environment and unconformabely below molasse sediments) are summarized as flysch or at least as "flyschlike". Generally the following environments can be recognized:
1. Devonian prebretonic flysch followed by Tournaisian lydites and by a second preasturian (Upper Visean - ? Lower Westfalien) flysch stage within the Alboran Balearic Trough.
2. Devonian (prebretonic) metamorphosed flysch in the Gemerids and N-Dobrogea Tulcea unit.
3. Presudetic flysch-environments:
a) Devonian - Lower Carboniferous: ? Western Grauwackenzone/Eastern Alps, Balcanids, Caucasian Fore Range Zone.
b) Lower Carboniferous metamorphosed flysch environments in the Eastern and Southern Carpathians (in Locva unit no. 33 the flysch begins possibly within the Devonian), Dobrogea/Macin unit, Caucasian Main Range Zone.
c) Upper Visean-? Lower Namurian flysch in the Eastalpine Gurktal Thrust system, Sardinia and ? Calabria/Sicily.
4. Preasturian flysch environments:
a) Intravisean - Westfalian: Catalonian Coastal Ranges, Pyrenees, Massif of Mouthoumet, Montagne Noire, Southern Alps, ?Outer Dinarids (possibly starting already within the Tournaisian).
b) Second flysch stage of Alboran Balearic Trough.
5. Upper Carboniferous (Westfalian) - Permian flysch environments in the Hellenids.
6. "Filling up type" without relation to an orogenetic phase:
a) In Bükk Mts. Upper Visean - Namurian sedimentologically well established flysch is superposed without unconformity by the marine shallow water Malyinka-Fm. It seems that there it is even possible that the flysch basin had been filled up by the end of the

Lower Moskovian (KOVACS & PERO 1983 a, b). That agrees with ARKAY's (1983) opinion that the Bükk-unit was only affected by alpidic orogenetic events. On the other hand the tectonically isolated flysch of the Brusnik anticline which is comparable which is comparable with those of the Bükk-unit has a metamorphic (? Variscan) overprint (EBNER & al., 1990).
b) In the Jadar-Paleozoic (no. 41) the flysch is overlain conformabely by shallow water (Namurian to Moskovian) sediments with algae, fusulinids, and brachiopods below a tectonic unconformity related to the asturian phase (STEFANOVIC & KULLMANN 1963; FILIPOVIC, 1974).
c) In the Pontids the Visean Thracian flysh is also overlain without unconformity by shallow marine sediments.
7. Long ranging (?) flysch environments:
Inner Dinarids (Drina - Golja): Devonian - Middle Carboniferous; Scythian Platform: Middle Devonian - Westfalian.
8. Hypothetic flysch environments:
Tuscany (presudetic) sensu VAI (1978), and the Balcanid Diabase-Phyllitoid-Fm. (up to the Middle Carboniferous) according to VAI (in press).
9. No Variscan flysch:
Flysch is only missing in those areas of Late Paleozoic marine sedimentation areas in which the Variscan orogeny was only weak or non-effective (Taurids, Caucasian Southern Slope Zone, Moesian Platform, Predobrogea Downwarp, Macedonia, Velebit).
In the Paleozoic of Graz the paleozoic sequences is terminated within the Namurian C/Westfalian A by non flysch shales without any younger Late Paleozoic cover.

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L'EVOLUTION PALEOZOIQUE DE LA PARTIE ORIENTALE DE LA REGION DES BALKANIDES (D'APRES LES DONNEES SEDIMENTOLOGIQUES)

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Abstract

The paleogeodynamic interpretations of the region are based on sedimentological, paleogeographic, paleoclimatic, in part petrological and paleomagnetic data.
The formation of the oceanic crust, the subduction processes and the development of the volcanic island arc took place far to the south of the present day geographic latitude of the region, before the Arenigian. During the Ordovician Gondwanian fragments were still transported from the south and accreted to the Russian plate forming a young mobile platform (transitional type of crust). The latter was covered by a marginal sea of variable bathymetry. The sedimentation was pelitic with stages of shallowing (sea current and Hummocky type arenites), marine iceberg sediments, sedimentation breaks, metamorphism and erosion after the Ordovician, deepening during the Silurian. The beginning of the Devonian was marked by moderately deep-water carbonates on the west and supply of clastics from Dobrudgea. During the Givetian the southern part developed into an active continental suture and the northern - into an active part of the marginal sea. The Upper Devonian sediments indicate a frontal arc (west of the area of Kraishte), intra-arc flysch basin (Kuchai - Cherna Gora), back-arc (the Western Balkanides) and a back-sea (the Moesian platform) with mobile floor (carbonate rocks with wash-out phenomena). The height differences, resulting from the contrasting movements in individual zones, lead to olistostrome phenomena from the arcs to the trough. In the Moesian area there was a carbonate platform during the Late Tournaisian. In the beginning of the Late Visean a dry land formed in the area of the back-arc and the active part of the marginal sea while the region to the north was loaded by silicate clastics. The movements in the middle of the Carboniferous played an important part in the continentalization of the crust. An Alpine-type orogenic structure formed on the place of the arc and an Jura-type folded area - on the place of the former back-sea. The final stage of the Variscan development (Late Carboniferous - Permian) is governed by collisional processes between the island arc and the young mobile platform.

L'évolution paléozoique des territoires de la Bulgarie contemporaine peut être reconstruite, principalement, à partir des critères paléoclimatiques et paléogéologiques (et surtout - sédimentologiques). En plus on dispose de données paléomagnetiques sporadiques sur le Stéphanien et sur le système permien (NOZHAROV et al., 1980).
Les renseignements recueillis récemment peuvent être utilisés pour une division tripartite du "Complexe diabaso-phyllitoïdes" (CDPh). La partie inférieure de ce complexe (Groupe de Tcherni Vrag) est composée par des ophiolites formées dans la zone de spreding; la partie moyenne - (Groupe de BERKOVITZA d'après HAYDUTOV et al., 1979) est interpretée comme originaire d'un arc insulaire ensimatique caractérisé par séries magmatiques tholéitiques et calco-alcalines; et, enfin, la partie supérieure (Groupe de Dalgui dial) est en général representée par une formation sédimentaire riche en phénomènes d'olistostromes. Ces processus, liés consécutivement à la croûte océanique, la subducion et la formation de l'arc insulaire volcanique, cessent, en effet avant le début de l'Arénig (YANEV, 1989). En plus ils s'effectuèrent assez loin au Sud de nos latitudes géographiques contemporaine.
La formation du socle des terrains bulgares possède le caractère de collage. Nous avons déjà souligné (YANEV, 1976) la structure hétérogène de la région moesienne. Ce

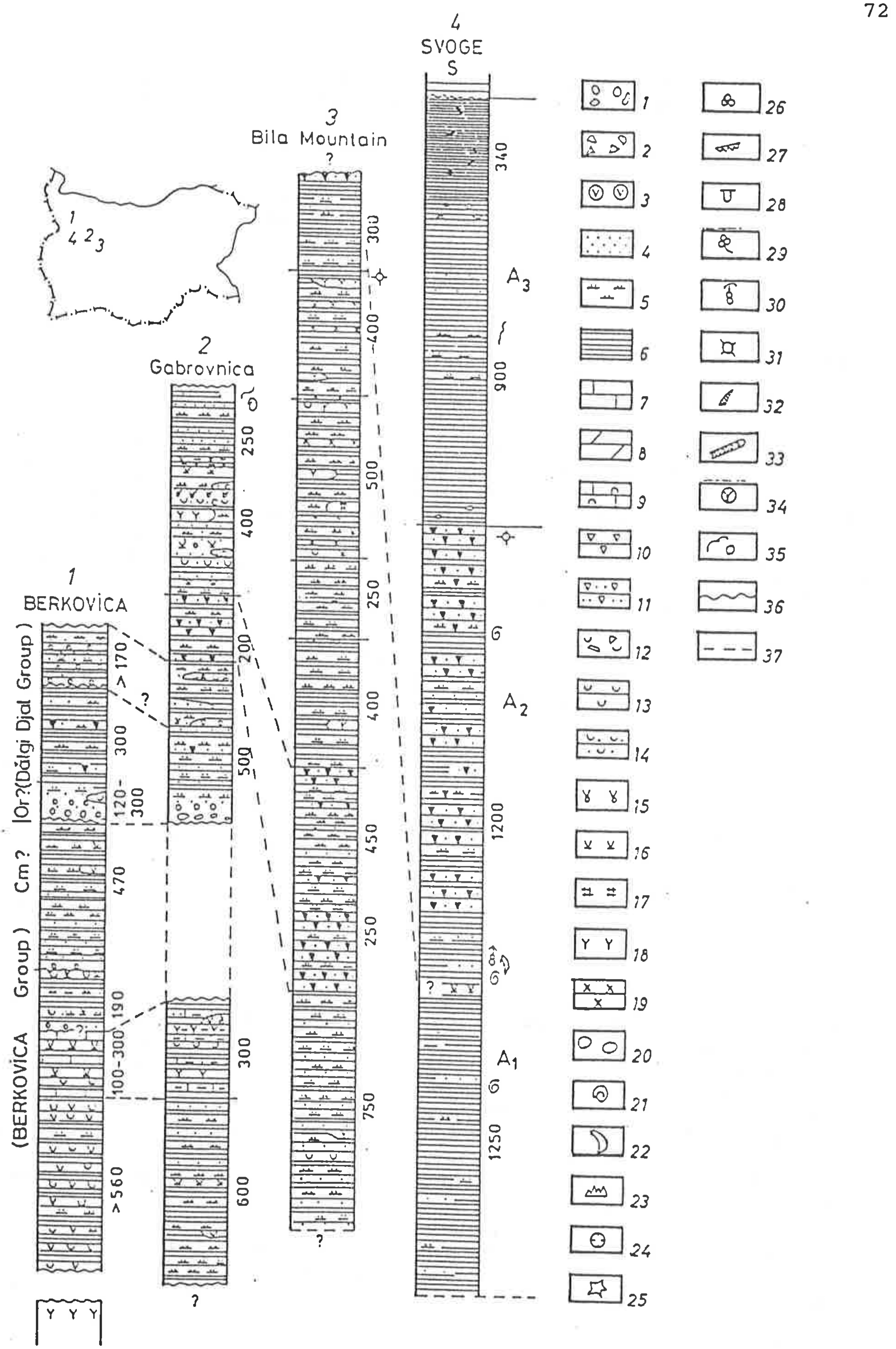

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Fig. 1. Colonnes lithologiques du complexe diabase-phyllitoïde (Groupe de Berkovitza - Ordovicien?); Groupe de Dalgui Dial - Cambrien-Ordovicien?) et de l'Ordovicien identifié dans la Bulgarie occidentale. 1 - conglomérats; 2 - brèche; 3 - conglomérats volcanoclastiques; 4 - grès; 5 - aleurolites; 6 - argilites; 7 - calcaires; 8 - dolomies; 9 - calcaires fauniques; 10 - lydites; 11 - quartzites; 12 - brèche volcaniques; 13 - tufs; 14 - tuffites; 15 - quartzporphyres; 16 - albitophyres; 17 - ultrabasites; 18 - gabbros; 19 - Gondites; 20 - concrétions; 21 - Goniatites; 22 - Brachiopodes; 23 - Conodontes; 24 - Coraux; 25 - Crinoïdes; 26 - Foraminifères; 27 - Graptolithes; 28 - Pélécypodes; 29 - flore; 30 - Trilobites; 31 - Acritarches; 32 - Tentaculites; 33 - Algues; 34 - Traces fossiles; 35 - limite d'érosion; 36 - corrélation supposée.

processus est dû à l'entrée successive et, pendant un intervalle de temps prolongé, de blocs qui venaient du Sud se joindre à la Plate-forme russe en formant, de cette manière, une plate-forme jeune et mobile. Les blocs s'unissaient déjà au cours du mouvement vers son emplacement actuel (le mécanisme et le temps du contact - entre la Plate-forme "moésienne" et le Massif "thrace" restent discutables). C'est avec ces "envois" consécutifs mis en mouvement par la subduction venant du Sud et son empilement sur le bord de la plate-forme mobile (prisme d'accrétion) que s'explique la formation d'olistostromes dans le Groupe de Dalgui Dial.
Une subduction des ophiolites précambriennes balcano-carpatiques sur une plate-forme moésienne épibaïkale s'effectua, d'après HAYDUTOV (1987), pendant le Cambrien approximativement.
Nous admettons (TENCHOV & YANEV, 1987) que le Groupe de Dalgui Dial (Fig. 1) pourrait être d'âge Ordovicien inférieur. Durant l'Arenig un fragment de croûte terrestre de type transitoire, acheminé vers nos latitudes géographiques, fut recouvert d'un bassin marin au caractère de mer marginale. L'activité volcanique dans cette mer fut pratiquement éteinte (on connait seulement des manifestations sporadiques, insuffisamment datées). En outre les composants clastiques apportées pour la plupart par des courants marins (YANEV, 1989), sont déjà mûres du point de vue minéralogique (Fig. 2) et partiellement structural. Le bassin de l'Ordovicien supérieur (Fig. 3) fut assez étendu. Ses parties périphériques sont marquées par des clastites plus grossières roches carbonnatées et par une minéralisation leptochloritique - dans les territoires actuels de la Serbie orientale et la Macédoine. En Bulgarie prédomine la sédimentation pélitique mais, tout de même, les niveaux et les lentilles de quartzites (au milieu de la coupe sans métamorphisme) possèdent des preuves structurales d'une sédimentation du type hummocky - c'est à dire - un bassin relativement peu profond, lors de cette étape de l'évolution tout au moins. L'absence d'interruption de la sédimentation entre le Groupe de Dalgui Dial et les roches plus oligomictiques de l'Arenig-Caradoc montre que le déplacement du fragment de la croûte en migration s'effectuait dans une zone alimentée à partir d'une source tout à fait différente (constituée de roches sédimentaires). Les hauts niveaux de la coupe de l'Ordovicien (formation de Tzeretzel) apportent une information très précieuse sur la situation géographique du bloc en question pendant cette étape. Il faut souligner que dernièrement on a rencontré dans les sédiments pélitiques exclusivement fins, qui rappelent beaucoup les "leederschifer" de l'Europe centrale, de rares fragments clastiques. Comme il est connu, ces sédiments s'interprètent comme des dépôts des eaux profondes avec des fragments clastiques tombants des icebergs en fonte. On peut, donc, affirmer que nos terres se situèrent, pendant cette époque, dans la zone du 40^{ième} parallèle de l'Hémisphère Sud.

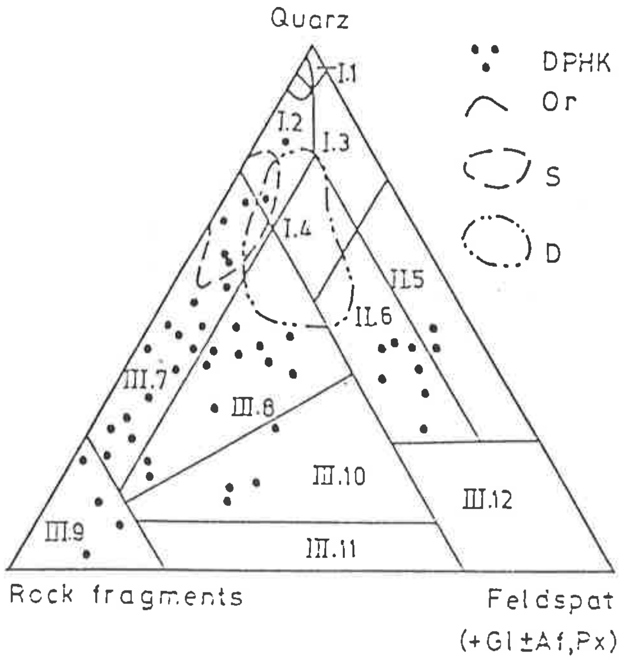

Fig. 2. Composition des roches psammitiques dans le complexe diabase-phyllitoïde, l'Ordivicien, le Silurien et le Dévonien (diagramme de classification d'après YANEV, 1970).

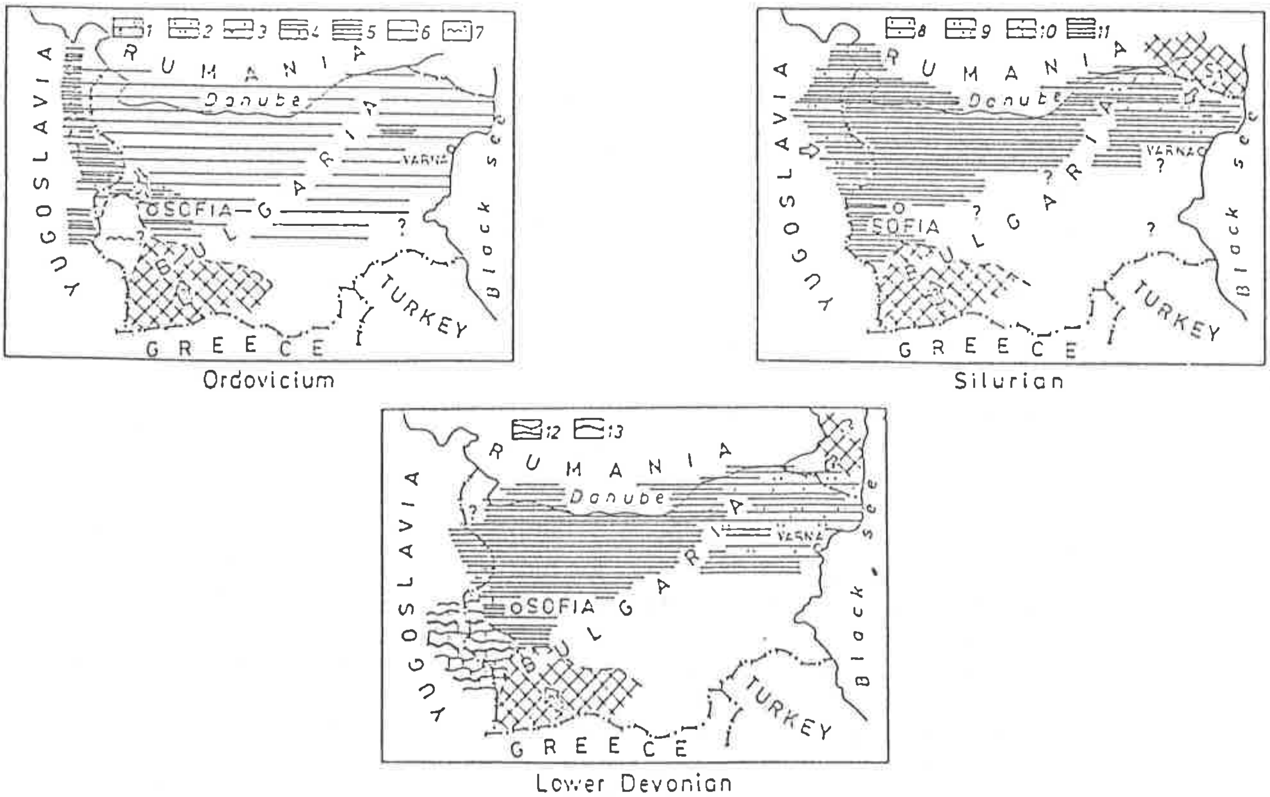

Fig. 3. Schéma litho-paléogéographique pour la Bulgarie pendant l'Ordovicien moyen et supérieur, le Silurien et le Dévonien inférieur. Ordovicien moyen-supérieur: 1-7. Domaine de sédimentation ordovicien argilo-terrigène. 1 - conglomérats; 2 - grès et aleurolites; 3 - chamosites; 4 - calcaires; 5 - argilites dominantes; 6 - distribution supposée de l'Ordovicien; 7 - zone d'érosion tardive supposée des sédiments ordoviciens. Silurien: 8-11. Domaine de sédimentation pélitique (silico-argileuse). 8 - grès; 9 - aleurolites; 10 - marnes et calcaires; 11 - argilites dominantes. Dévonien inférieur; 12-13. Domaine de sédimentation argilo-carbonatée et siliceuse; 12 - calcaires noduleux et laminés et marnes; 13 - paquets de lydites. Les autres signes - comme ceux de l'Ordovicien et le Silurien. Avec carrés - les terres fermes, avec flèches les directions du paléotransport.

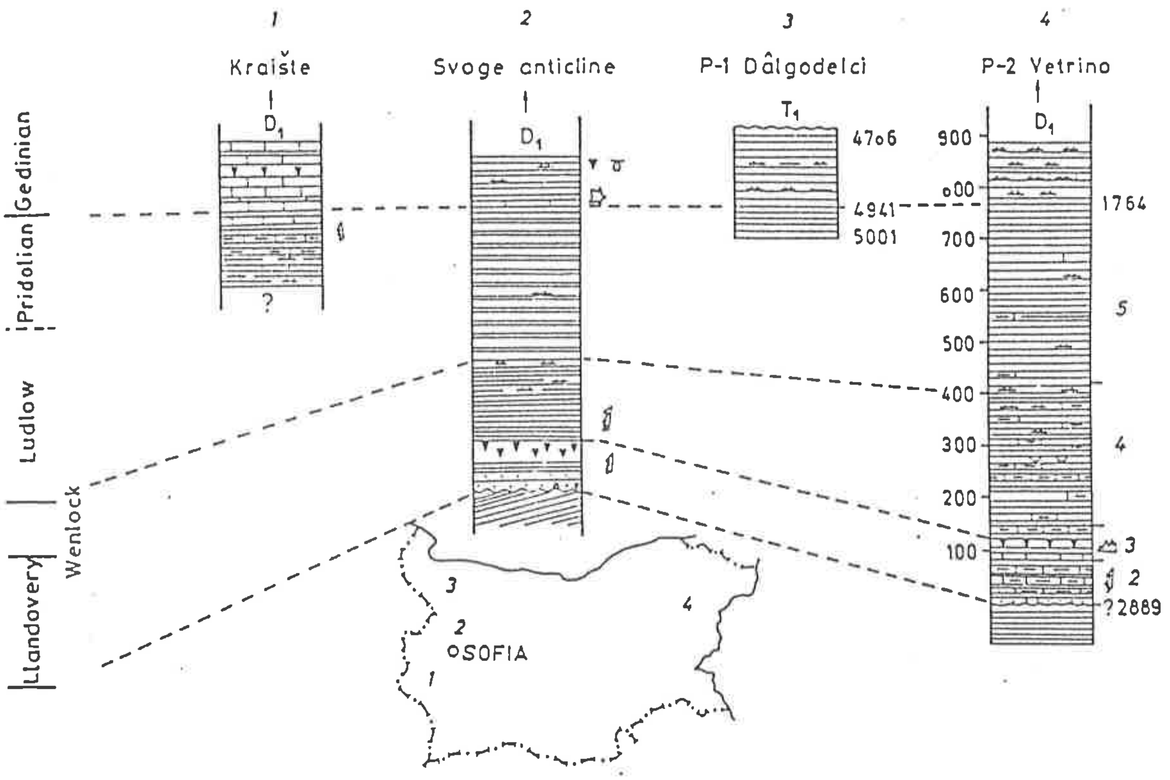

Fig. 4. Colonnes lithologiques des coupes siluriennes en Bulgarie (légende - voir fig. 1).

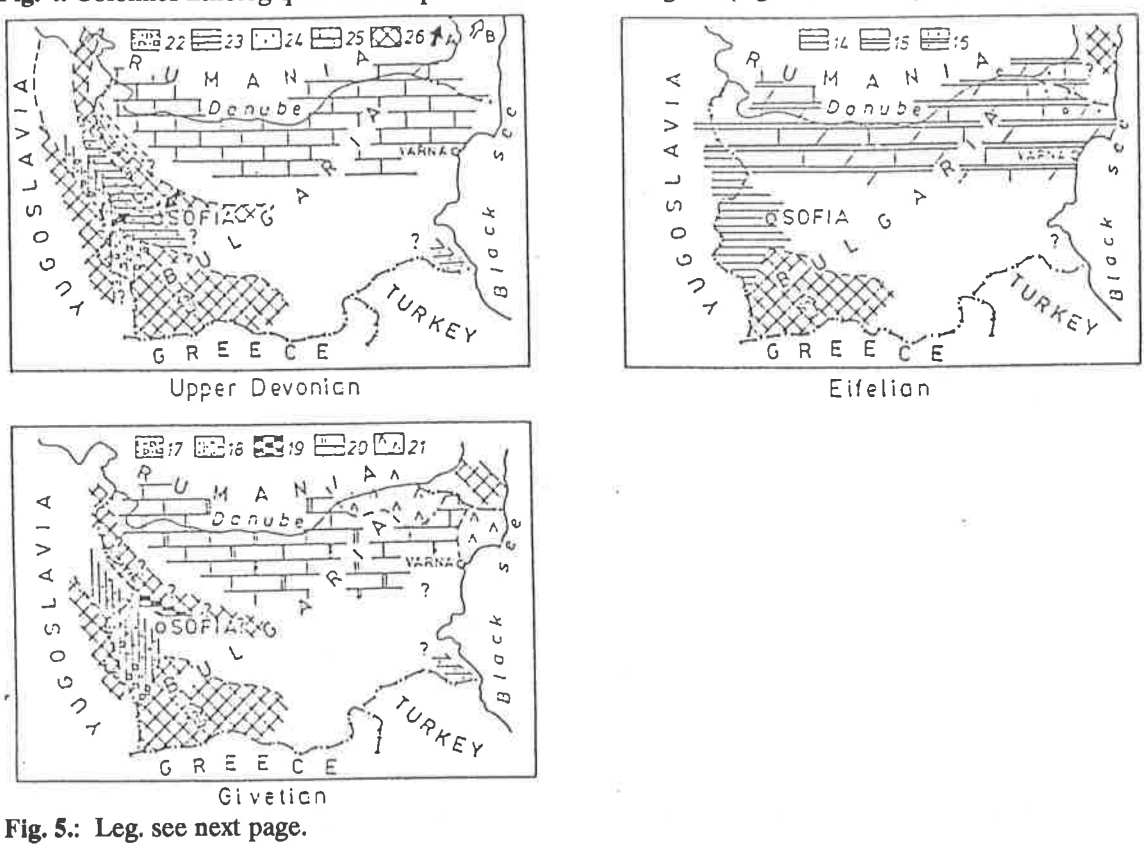

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Fig. 5. Schéma litho-paléogéographiques pour la Bulgarie pendant le Dévonien moyen et supérieur. Eifélien: 14 - Domaine de sédimentation pélitique dominante (argilites); 15 - domaine de sédimentation argilo-carbonatée (argilites, marnes, calcaires argileux, calcaires); 16 - secteur avec contaminations extraclastiques (alternation des roches indiquées ci-dessus avec des niveaux de conglomérats, gravites,calcaires grêseux et calcaires silteux). Givetien: 17-18. Domaine de sédimentation turbiditique: 17 - flysch de grauwacke avec flexoturbidites conglométatiques: 18 - flysch de grauwacke; 19 - zone de sédimentation silico-pélitique (argilites et silicites); 20-21. Domaine de sédimentation homogène: 20 - calcaréo-dolomitiques; 21 - sulfato-carbonatée (dolomies, anhydrite). Dévonien supérieur: 22-23. Domaine de sédimentation turbiditique; 24-25. Domaine de sédimentation carbonatée: 24 - calcaires littoraux clastiques et de bio - herme; 25 - calcaires, un peu de dolomies; 26 - A,B - directions du paléotransport: A - transversales au bassin turbiditique, B - longitudinales.

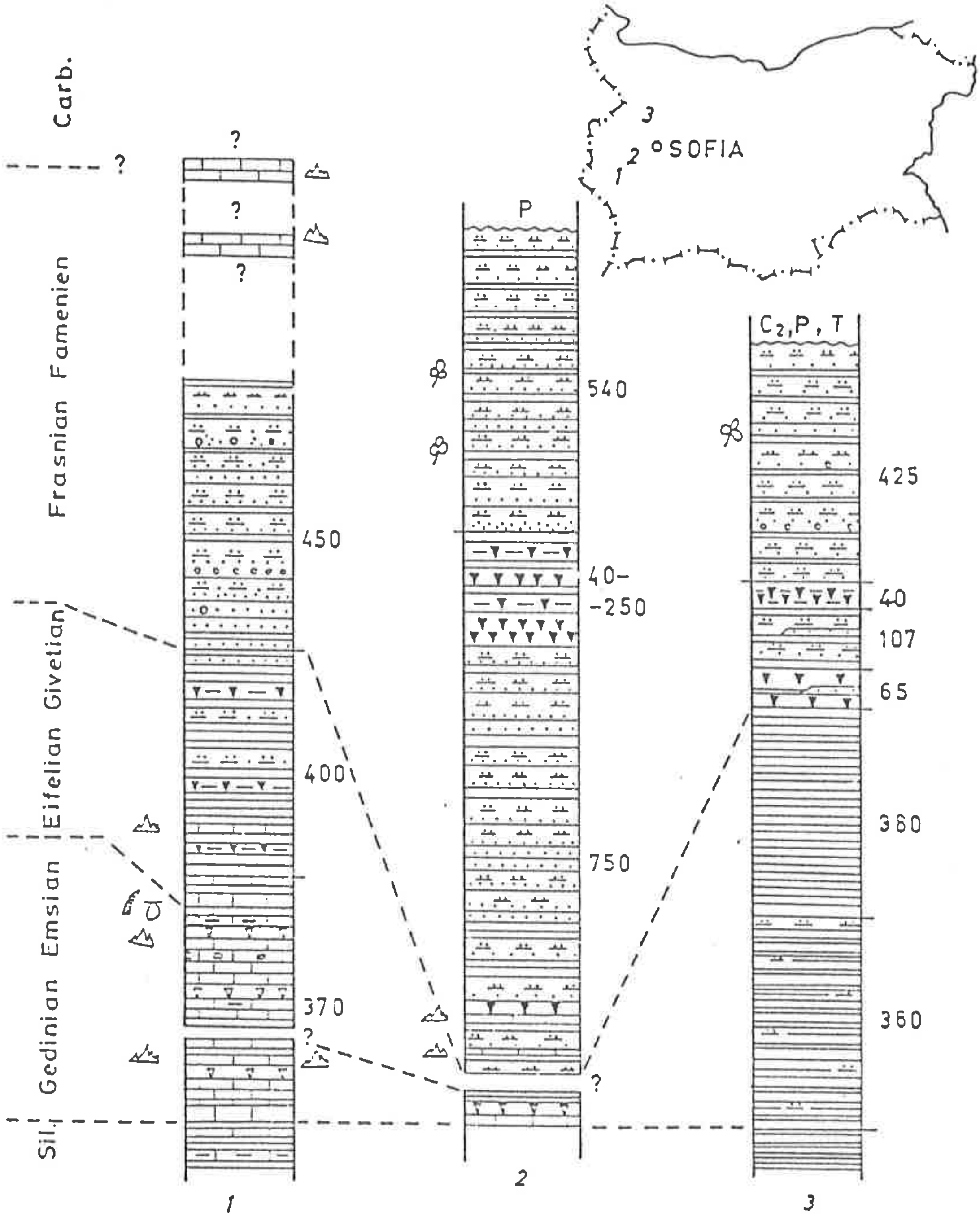


Fig. 6. Colonnes lithologiques schématiques du Dévonien dans la Bulgarie occidentale (légende - voir fig. 1).

Entre l'Ordovicien et le Silurien se manifestent des processus de plissement, de métamorphisme et d'érosion (discordance angulaire et métamorphique). Au début du Silurien la sédimentation s'alimente indiscutablement à partir des terrains ordoviciens (il n'existe pas d'observation d'érosion du CDPh) (Fig. 2). Plus tard la région apparaît déjà dans la partie profonde d'une mer marginale (avec de roches siliceuses et d'argilites à Graptolithes - Fig. 4). Les terres fermes de la zone d'alimentation étaient éloignées et basses. Des zones plus marginales du bassin se distinguent (Fig. 3) dans la partie septentrionale de la Serbie orientale et en s'approchant vers une paléoterre ferme de Dobroudja. Pendant le Devonien inférieur, à la zone occidentale et moins profonde s'incorpore déjà la région du Kraischté bulgare de l'ouest (avec des calcaires rubanés - type "welenkalk" et des Conodontes caractéristiques, pour des sédiments des eaux peu profondes (Fig. 3). En même temps la paléoterre de Dobroudja s'approche (on observe plus de matériel clastique et plus grenu, à part cela dans les sédiments de la Bulgarie du Nord-Est). C'est durant l'Eifélien (Fig. 5) qu'autour de la paléoterre ferme de Dobroudja se différenciaient des zones de sédimentations argileuses (au Sud-Sud-Ouest) et carbonatées aux calcaires clastiques (au Nord) (Fig. 6). Pendant l'âge givetien (Fig. 5, 9) la mer marginale commence à se transformer de cette manière que sa partie méridionale devient un bord continental actif tandis que sa partie septentrionale acquiert le caractère de partie passive d'une mer marginale. Les évaporites de la Dobroudja du Sud marquent le cadre lagunaire de la paléoterre de Dobroudja, ainsi que son adhésion à la zone climatique aride méridionale. Les données sur les évaporites givetiennes dans la Bulgarie Nord-Est (Fig. 5) et sur le minerai de fer oolitique et la végétation psyslophytique dans le Devonien supérieur (et Moyen ?) de la Bulgarie de l'Ouest et du Sud-Ouest, témoignent en faveur d'un mouvement continu des blocs de la croûte terrestre et d'effets d'une contraction plus tardive de la croûte. Alors dans la partie active du bassin se forma un système composé d'un arc insulaire (de direction Nord-Ouest-Sud-Est) et d'un bassin interarc de flysch. Les données matérielles et de paléotransport indiquent très bien la présence, pendant le Dévonien supérieur, de:
a) Un arc frontal - dans le Karichté occidental qui approvisionnait le flysch au matériel clastique d'âge silurien et autre;
b) Un bassin interarc de flysch (Fosse de Kutchai - Tcherna Gora);
c) Un arrière-arc (+ou- à l'emplacement des Balcanides oxidentales au moins) et,
d) Une mer arrière (sur la région moésienne contemporaine) (Fig. 5, 9).
Différentes sections de cette dernière montrent pendant l'époque du Devonien moyen et supérieur une mobilité variable en dépendance de la préhistoire des blocs collés. En conséquence de cela on a constaté plusieurs érosions internes. La surrection des arc insulaires fut intensive et par suite les roches du CDPh s'englobent, pour la première fois, dans la zone d'alimentation (Fig. 2). En plus il s'avère qu'en certains secteurs se developpèrent des processus de formation d'olisthostromes lors des quels dans le flysch du Dévonien supérieur (jusqu'à la base du Carbonifère inférieur) entrèrent des olistholithes de calcaire du Dévonien inférieur et moyen, des lydites siluriennes, des granitoïdes, des gabbros etc. (bien observables dans la Serbie orientale et partiellement dans le Kraischte bulgare).
Les mouvements les plus differenciés furent ceux de la limite méridionale de l'arrière-arc où, à travers des accidents tectoniques, pénétraient des solutions et des exhalations qui favorisaient le dépôt des silicites et des lydites (Fig. 6) et, d'après les dernières données - des roches volcaniques (ZAGORČEV & BONČEVA, 1988). Pendant le Tournaisien supérieur (Fig. 8, 9) probablement commencent aussi les mouvements aux alentours de l'arc frontal (les roches siliceuses associées aux argilites et peu de calcaires dans le karischté oxidentale - voir Fig. 7). Ce sont les messagers de l'activité tectonique qui, au début du Viséen supérieur amène à la transformation de la partie active de la mer marginale en une terre surélevée. Cette terre ferme fournissait le matériel silico-clastique au Nord où, pendant le Tournaisien supérieur - Viséen moyen - début de Viséen supérieur (Fig. 7) avait existé une Plate-forme carbonatée (sur un terrain dévonien profondément érodé). Les mouvements tectoniques vers le milieu du Carbonifère représentent l'étape la plus importante dans le processus de la continentalisation de l'écorce qui est lié aussi à l'intrusion des granitoïdes de Stara planina (et des granitoïdes de la Bulgarie du Sud). Cependant après les événements

sudètes, la région avait conservé une mobilité différenciée et par conséquence se forma la dorsale très surélevée des Variscides - comme un édifice orogénique alpinotype très compliqué (à la place de l'arc) et d'autre part, une zone de plissement germanotype moins elevée - sur l'aire marine de l'ancienne mer arrière, sur la périphérie Nord-Est de laquelle se forma une avant-fosse (Fig. 9).
L'étape finale de l'évolution varisque (Carbonifère supérieur - Permien) fut, en effet, prédestinée par les processus de collision entre l'arc insulaire et la jeune et mobile plate-forme qui fut, grâce au collage, adjointe à la Plaque russe (Fig. 8).
Les premières données matérielles et de paléotransport évidentes sur la présence des roches métamorphiques de haut degré de métamorphisme et des roches magmatiques pour la région du Massif de Thrace datent du Stéphanien.
Quelques fragments de roches métamorphiques apparaissent dans les sédiments du Dévonien supérieur du Kraishté mais ils sont sporadiques et d'origine non identifiée.
Les volcanites (et les roches intrusives) westphalo-permiennes sont intimement associées aux formations sédimentaires des molasses inférieures. Il faut les considerer, donc, comme les manifestations d'un magmatisme typiquement orogénique lié aux secteurs plus labiles de l'écorce terrestre. Du point de vue paléoclimatique, pendant la période carbonifère, les terres bulgares contemporaines passèrent par la zone climatique équatoriale et, durant la période permienne - atteignirent l'Hémisphère Nord ou plus exactement - tombèrent dans le cadre de la zone aride septentrionale.
En comparant les paléogéographies varisque et alpine on peut mettre en évidence certaines similitudes. Probablement ces similitudes sont dûes à l'influence du plan structural et géodynamique varisque, puisqu'elles apparaissent regénérées dans les évènements tectoniques alpins tardifs, et cela après une période continue d'interrelations et de zonalités tout à fait différentes durant la plupart du Secondaire.

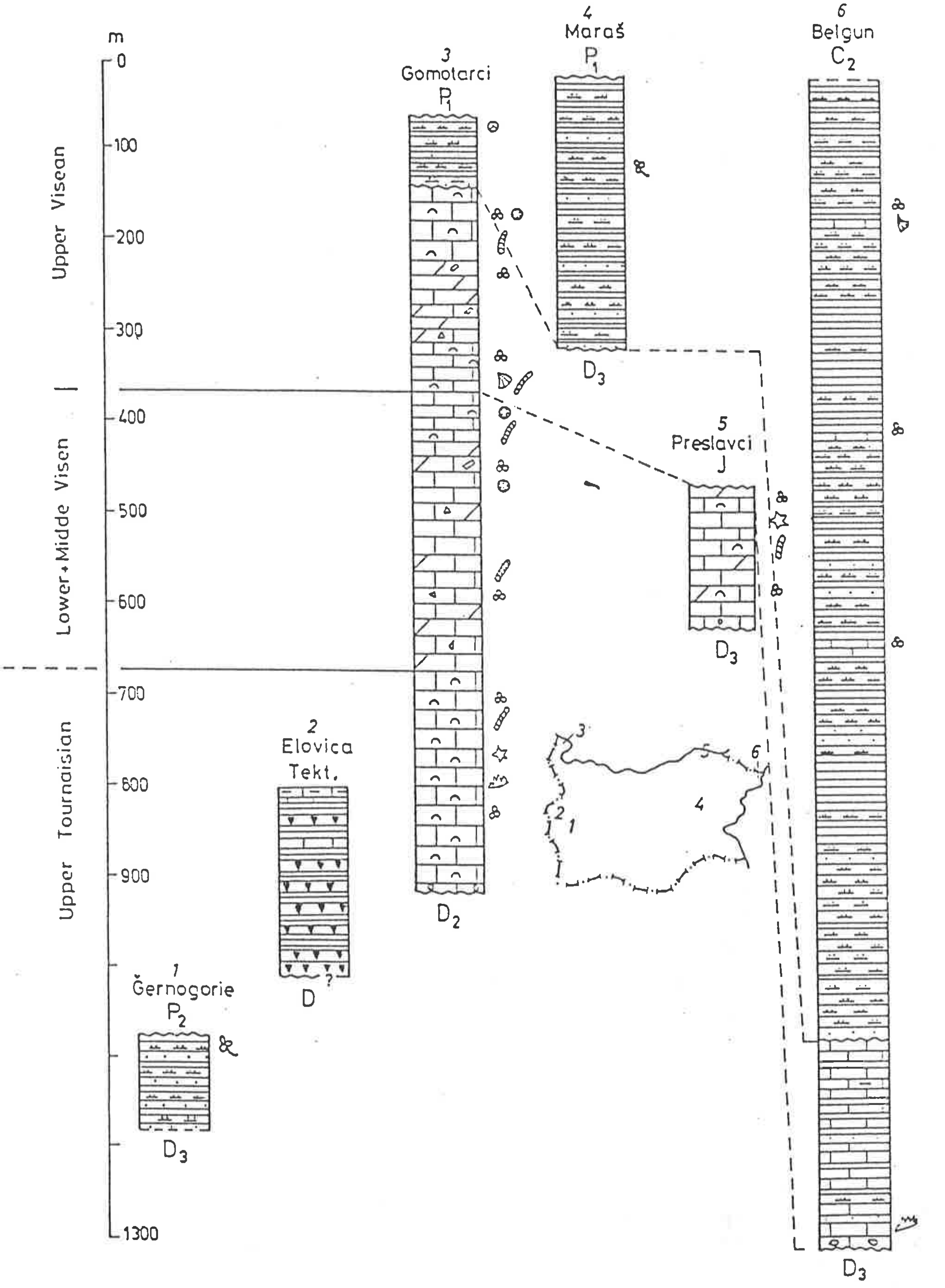
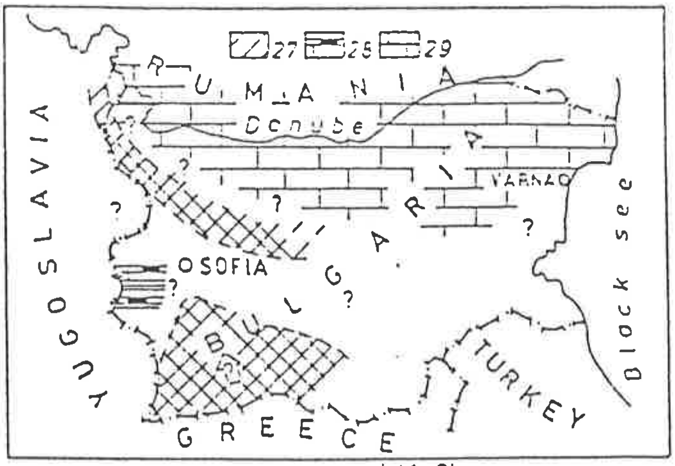
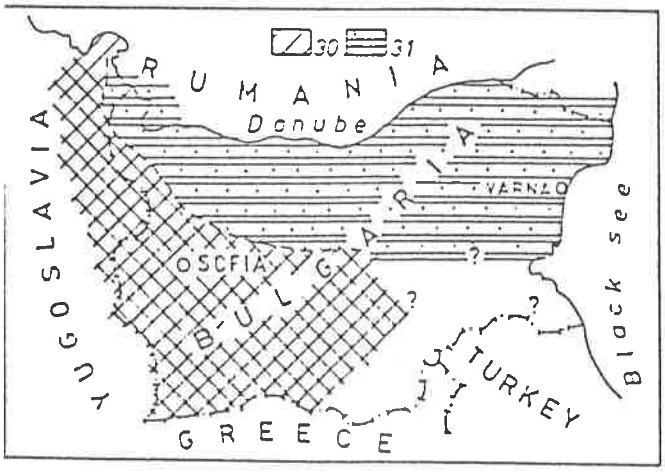


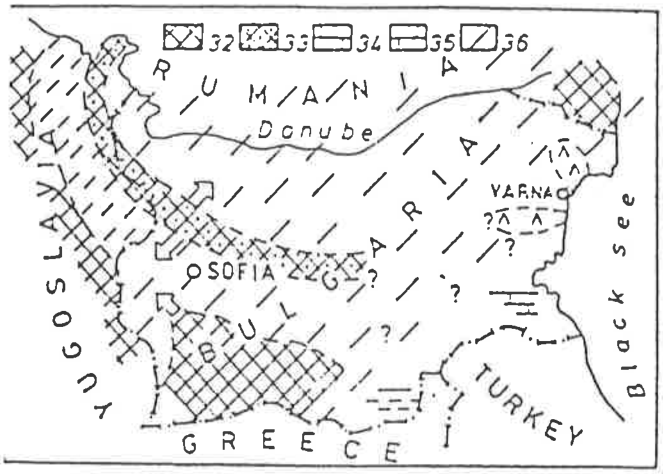
Fig.7: Colonnes lithologiques schématiques du Carbonifère en Bulgarie (lég. voir Fig. 1)



Upper Tournaisian (+ $C_1^{V(1-2)}$ for N-Bulgaria)



Upper Visean



Permian

Fig. 8. Schémas litho-paléogéographiques simplifiés de la Bulgarie pendant le Carbonifère inférieur et le Permien. Tournaisien supérieur: 27 - terre ferme, 28 - domaine de sédimentation silico-carbonato-argileuse; 29 - domaine de sédimentation carbonatée peu profonde. Viséen supérieur: 30 - terre ferme; 31 - domaine de sédimentation terrigéno-argileuse. Permien: 32 - secteur de sédimentation carbonatée; 33 - secteur proche d'une aire de sédimentation silico-carbonatée; 34 - domaines de sédimentation des bassins continentaux; 35 - secteurs de sédimentation lagunaire salifère; 36 - terre ferme surélevée (disposition de la dorsale varisque).

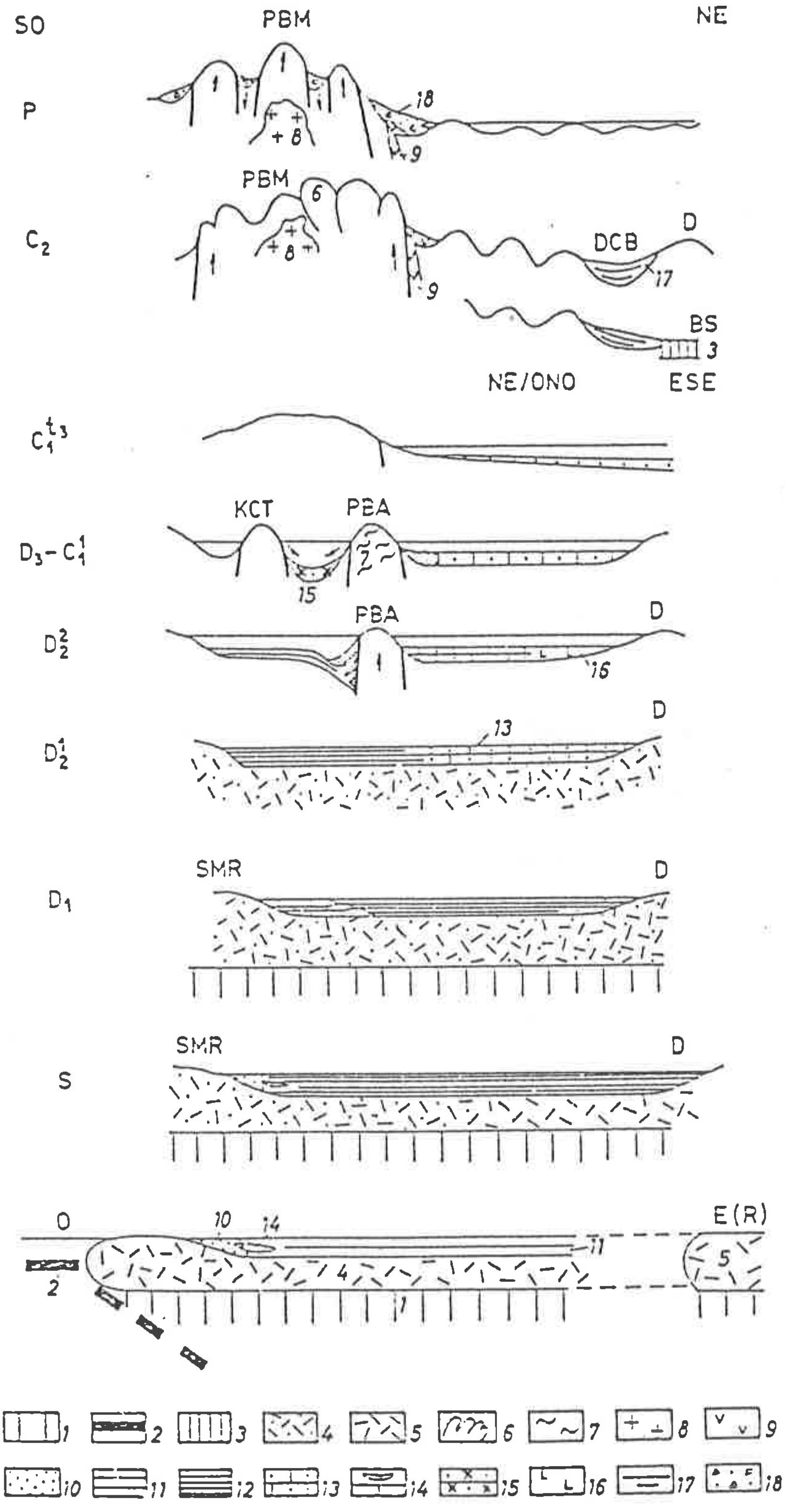


Fig. 9. Modèle néomobilistique de l'évolution paléozoïque de la Bulgarie. 1 - manteau; 2 - croûte océanique; 3 - croûte subocéanique; 4 - croûte subcontinentale; 6 - plis, chariages, prisme d'accrétion; 7 - métamorphites de degré bas de métamorphisme; 8 - roches intrusives; 9 - roches volcaniques; 10 - roches silico-clastiques fines (du talus et du plateau continental); 11 - sédiments pélagiques; 12 - pélagites des eaux profondes; 13 - dépôts de la plate-forme carbonatée mobile; 14 - roches carbonatées des eaux relativement profondes aux peu profondes; 15 - flysch; 16 - sédiments évaporitiques lagunaires; 17 - roches clastiques et argiloterrigènes continentales et transitives. D - paléoterre ferme de Dobrudja; KA - arc insulaire du Kraichté; KCT - fosse de Kutchaï-Sredna gora; PBA - arc paléobalkanique; MMKP - plateforme carbonatée mobile moésienne; PBM - système montagneux paléobalkanique; MP - plate-forme moésienne; DCB - bassin houiller de Kobroudja; BS - croûte subocéanique(?) de Tcherno moré (mer noire); SMR - système montagneux Serbo-Macedono-rodopienne; E (P) - Eurasie (plate-forme russe).

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<u>Note des rédacteurs</u>

Nous avons pris la liberté d'effectuer quelques corrections qui permettent de rendre certains passages plus compréhensibles, tout en laissant à l'auteur l'originalité de son style et sa responsabilité quant au texte.

PALEOZOIC GEODYNAMIC EVOLUTION OF SARDINIAN - CORSICAN AND CALABRIAN - PELORITAN CRYSTALLINE MASSIFS.

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Abstract

During the Paleozoic, in Sardinia two distinct stratigraphic-tectonic cycles developed: the first from lower Cambrian to lower Ordovician; the second from middle Ordovician to lower Carboniferous. The so-called Sardic (or Sardinian) phase is of middle Ordovician age. Beginning from this time, two opposed paleogeographic domains formed: an internal domain in northern Sardinia and an external domain in southern Sardinia. Therefore, the true Hercynian cycle began from the middle-upper Ordovician age. The Hercynian chain developed during upper Devonian-lower Carboniferous age. The chain, entirely ensialic, modelled by utilizing the lower Cambrian-lower Ordovician paleogeographic domains in the external zone and the middle-upper Ordovician-Devonian domains in the internal zone. The internal zone (northern Sardinia) is connected to the foreland (southern Sardinia) by a nappe zone (central Sardinia). Northern Corsica probably represents the backland of the chain overthrusting the more internal zone along intracontinental NE dipping shear zone (main shear zone). The chain forms a large arcuate belt with SW concavity. The Calabrian-Peloritan Hercynian segment was originally placed near central-southern Sardinia. The nappe zone of central Sardinia continued as far as the Calabria region thus completing the arcuate form of the chain. Several structural and stratigraphic data indicate that SW Sardinia represents a relative foreland and that another minor shear zone developed in southernmost areas (secondary shear zone). Some high grade metamophites, outcropping in widespread areas of central-southern Calabria, probably represent a emerged area (from middle Cambrian to lower Ordovician) originally placed in southern areas of Sardinia.

1. INTRODUCTION

The most recent data concerning Sardinian-Corsican and Calabrian-Peloritan Hercynian Massifs suggest a Paleozoic geodynamic evolution notably different from that already known (CARMIGNANI et al., 1981). In this paper the geodynamic evolution coherent with the new knowledge will be exposed and the main problems still unsolved will be discussed.

2. SARDINIAN-CORSICAN MASSIF

In the Sardinian-Corsican Massif two main Paleozoic sedimentary-tectonic cycles may be recognized (MINZONI, 1987, 1988b, 1989). The first cycle developed from lower Cambrian to lower Ordovician; the second from middle Ordovician to upper Devonian-lower Carboniferous. The so-called Sardic (or Sardinian) phase is of middle Ordovician age (BARCA et al., 1986). This phase produced contemporaneously compressional and tensional tectonics in adjacent areas (MINZONI, 1989). Both the age (between two distinct sedimentary cycles) and structural characteristics (compressional-tensional tectonics) indicate that Sardic phase represents a particular geologic "moment" in the Paleozoic geodynamic evolution.

3. THE FIRST CYCLE

The lower Cambrian-lower Ordovician sedimentary cycle is recognizable only in central-southern Sardinia. In this area the lower Paleozoic successions are well exposed and the Hercynian metamorphism never the greenschist facies (CNR, 1987). On the contrary in northern Sardinia the metamorphism is in the amphibolite facies so that it is very difficult to distinguish the Paleozoic successions.

Southern Sardinia is the only area where terrains of lower-middle Cambrian age are present. The succession consists of terrigenous and carbonate sequences. On the contrary, the middle Cambrian-lower Ordovician succession consists of entirely terrigenous sequences (micaceous sandstones, siltites and shales).

In southernmost Sardinia the lower Cambrian fossiliferous sequences overlie a non-fossiliferous succession containing lenses of conglomerates formed by erosion of acidic magmatites (MINZONI, 1980). The lower part of the non-fossiliferous succession is metamorphosed by alkaline granite: the so-called "Cataclastic Gneiss". The thermal aureole due to intrusion of Cataclastic Gneiss is deformed by the Hercynian tectonics. late Hercynian granitoids cut across the whole Paleozoic succession. The radiometric Rb-Sr dating method on whole rocks yielded an age of 427+ or - 34 Ma. (COCOZZA & al., 1977; FERRARA et al., 1978). More recently, the zircon U-Pb method yielded and age of 478+or-16 M.a. (DELAPERRIERE & LANCELOT, 1989). Both Rb-Sr and U-Pb ages conflict with structural-stratigraphic data. This discrepancy represents an open problem.

In central Sardinia the first cycle sedimentation began from middle Cambrian and, as in southern areas, the succession is entirely terrigenous. In this region the micaceous sandstones, siltites and shales overlie an acidic magmatite (the so-called "Porphyroid") showing the same lithologic-geochemical characteristics of the Cataclastic Gneiss of southernmost Sardinia (MAZZUCCHELLI & MINZONI, 1987). Therefore in central Sardinia an important hiatus (from lower to middle Cambrian) is probably present. The terrigenous sequences show different thicknesses in adjacent areas (growth faults; MINZONI, 1987): in some areas they reach 600-800 m; on the contrary in adjacent areas the same sequences are very thin or completely lacking so that the Caradocian terrains may directly overlie the Porphyroid.

4. THE SECOND CYCLE

Southern Sardinia is the only area where the middle Ordovician compressional tectonics are recognizable. The Sardic phase produced E-W trending folds not accompanied by penetrative schistosity or metamorphism. On the contrary, in central-northern Sardinia the Sardic phase is characterized by tensional tectonics and is accompanied by the outflow of abundant volcanics.

These data indicate that from the middle Ordovician two different paleogeographic domains began to form: a southern domain (central-southern Sardinia) and a northern domain (central-northern Sardinia). The opposition of two domains developed during the upper Ordovician. In fact, the upper Ordovician succession of central-northern Sardinia consists of very thick (500-1000 m) non-fossiliferous flyschoid sequences. On the contrary, in central-southern Sardinia the upper Ordovician sequences are very thin (50-100 m) and consist of mature sandstones with fossiliferous carbonate levels. In southern and northern Sardinia the Silurian-Devonian succession consists of the well known black shales and limestones with lenses of basic volcanics. Moving northward, the upper Ordovician-Silurian volcanics become more and more abundant. This means that the opposition of two paleogeographic domains continued during Silurian-Devonian times.

The exposed data indicate that starting from middle Ordovician a new sedimentary-tectonic cycle began to develop (MINZONI, 1985; 1987; 1988b; 1989). During lower Cambrian-lower Ordovician times the tensional tectonics developed very slowly and, especially during middle Cambrian-lower Ordovician, the same siliciclastic sediments formed in widespread areas. On the contrary, starting from middle-upper Ordovician, tensional tectonics developed very quickly and two very different paleogeographic domains shaped in central-southern Sardinia (external domain) and in central-northern Sardinia (internal domain). The sudden instability of the crust during middle-upper Ordovician times is evidenced by the outflow of abundant volcanics and by the sedimentation of thick flyschoid sequences. Therefore, starting from the middle-upper Ordovician the Hercynian "Geosyncline" began to form and the true Hercynian cycle began to develop. The Hercynian "Geosyncline" developed by transcurrent movements of the sialic crust. This geodynamic model unifies all geological data. Using this model we may connect the compressional-tensional tectonics (Sardic Phase) of central-southern Sardinia with the tensional tectonics of northern Sardinia (rifted area). In this framework the Cambrian-lower Ordovician succession does not belong to the true Hercynian cycle but probably represents a post-orogenic sedimentary cycle related to a pre-Cambrian orogeny.

5. THE HERCYNIAN CHAIN

The Hercynian orogenesis developed during upper Devonian-lower Carboniferous times (CARMIGNANI et al., 1981). It evolved homogeneously with a tectonic transport complexively trending NE-SW. However, lithostratigraphic metamorphic and structural differences make it possible to distinguish several zones: a SW, a central-southern, a central-northern and a NE zone.
In the SW zone the main Hercynian tectonics produced open folds accompanied by sub-vertical strain-slip cleavage in the higher structural levels. Towards the deeper levels the sub-vertical open folds progressively change to isoclinal recumbent folds and the strain-slip cleavage progressevely changes to a flow schistosity. In the same way the metamorphic recrystallisations become more and more important to reach the greenschist-biotite facies.
In central-southern zone several Hercynian nappes are present (CARMIGNANI & al., 1981). The nappes formed at the end of the main Hercynian phase through shear planes nearly paralleling the flow schistosity; therefore the nappes represent the most important structural results of Hercynian tangential tectonics. The flow schistosity was formed contemporaneously to isoclinal recumbent folds and materializes a synkinematic metamorphism in the greenschist facies. The polarity of sequences and the geometry of isoclinal recumbent folds indicate that the direction of overthrusting is towards W and SW. Therefore the nappes are of E and NE provenance.
Central-NE zone displays a structural style characteristic of a deeper subducted crust. There are enormous recumbent isoclinal SW verging folds with their inverted limbs often preserved (CARMIGNANI & al., 1979; MINZONI, 1988a). The same folds are accompanied by sub-horizontal flow schistosity and synkinematic metamorphism in the greenschist facies. Two tangential tectonic phases have been recognized (DESSAU & al., 1983).
The NE zone represents the deepest subducted crust. In this zone several tangential tectonics and several dynamic and static metamorphic events have been recognnized (CARMIGNANI & al., 1979b). Syntectonic recrystallizations related to older tangential phase (corresponding to main phase of nappe zone) do not go beyond the greenschist facies. The amphibolite facies was reached during a successive static metamorphic event. A second tangential phase deformed the static minerals and produced second generation of isoclinal recumbent folds. The axes related to main isoclinal structures show a rotation from N 140 in central and NE areas to N-S in SE and southern areas. Therefore the Hercynian chain forms a large virgation with SW concavity.
A paleogeographic reconstruction which considers the direction of movement of allochthonous units leads us to identify a relatively stable domain in SW Sardinia and a more complexely evolved domain in central-NE areas of the island. A detailed analysis of stratigraphic and structural characteristics of the various zones indicates that the Hercynian chain modelled utilizing both the lower Cambrian-lower Ordovician domains and the true Hercynian domains. More precisely, the Hercynian chain modelled utilizing lower Cambrian-lower Ordovician domains in central-SW Sardinia (nappe zone and foreland) but the middle Ordovician-Devonian domains in central-NE zone of the island (more internal zones of the chain). In fact, in central-SW Sardinia the nappes are

primarily made up of thick middle Cambrian-lower Ordovician terrigenous sequences (CARMIGNANI & al., 1978). This means that the overthrust surfaces formed by utilizing the middle Cambrian-lower Ordovician synsedimentary growth faults (inversion tectonics). The highest nappe of the nappe zone is of enormous dimensions and overlies several small nappes (at present outcropping in tectonic windows). Several of the small nappes consist of Porphyroid overlain by thin middle Cambrian-lower Ordovician-Silurian-Devonian sequences (CNR, 1987). This means that the lower Paleozoic structural highs of central Sardinia were crushed and dismembered into small nappes by the overlying greater nappe. On the whole, central-southern Sardinia display a structural style characteristic of superficial levels.
In the central-northern zone the enormous SW verging isoclinal folds are primarily made up of the upper Ordovician flyschoid sequences (MINZONI, 1988a). The NE area is characterized by high grade metamorphism typical of deeper subducted crust. These data indicate this zone was modelled parallel to the true Hercynian domains. Therefore the axis of the chain roughly corresponds to the axis of the Hercynian "Geosyncline". Both volcanics and terrigenous flyschoid successions of middle-upper Ordovician age are evidence of tensile movements that thinned the sialic crust. The NE rifted area corresponds to the portion of crust that was strongly fractured and deeply subducted during Hercynian orogenesis. This evolution pattern indicates that the NE zone, characterized by more complex metamorphic history and by polyphase structural evolution, represents the deepest outcropping part of the chain. In opposition, the less deformed central-southern zone represents the foreland of the Hercynian chain. This structural contrast reflects the opposition of the two paleogeographic domains that, starting from middle Ordovician, characterized the Hercynian "Geosyncline". The central zone, showing intermediate structural and metamorphic characteristics, represents the connection between the relative foreland and the axial zone. The nappes of central-southern Sardinia are closely related to the Cambrian-lower Ordovician paleogeographic domains. Complexively, moving towards NE, the true Hercynian domains played a more and more important role in building the Hercynian chain.
In northern Corsica, a slightly deformed Paleozoic succession overlies a "Basement" characterized by high grade metamorphites. Therefore northern Corsica probably represents the backland of the Hercynian chain (CARMIGNANI & al., 1979a). In this pattern we may consider northern Corsica as the crustal element overthrusting the axial zone along intracontinental NE dipping shear zones. This framework provides a coherent explanation for the nature of the overthrust, the style of the belt, the division into several metamorphic zones and the chronological relationship between crystallization and deformation. SW Sardinia probably represents a relative foreland of the chain. In fact the deeper structural levels outcropping in southernmost Sardinia are typical of a shear zone.

6. THE MAIN OPEN PROBLEMS

The age of the Cataclastic Gneiss and Porphyroid of central-southern Sardinia represents an open problem. As mentioned, the stratigraphic-structural data conflict with radiometric ages. In fact, in southernmost Sardinia the conglomerates, interbedded with the lower non fossiliferous sequences, derived by erosion of acidic magmatites. The underlying Cataclastic Gneiss is an acidic magmatite. Therefore the conglomerates probably formed by erosion of the Cataclastic Gneiss. In conclusion, we may suppose that Cataclastic Gneiss intruded (as laccolithic body) the lower part of the non fossiliferous succession and subsequently it was eroded to form conglomerate lenses. In central Sardinia the middle Cambrian-lower Ordovician succession overlies the Porphyroid. Thermal metamorphism is not present and dykes of Porphyroid do not cut across the overlying middle Cambrian- lower Ordovician sequences. On the contrary, several dykes linked to middle-uplper Ordovician magmatism clearly crosscut the above mentioned sequences. In some areas there is a clear stratigraphic polarity from Porphyroid to middle Cambrian-lower Ordovician sequences (NAUD & PITTAU DEMELIA, 1985). The Cataclastic Gneiss and Porphyroid are lithologically and geochemically very similar to each other but they are very different from middle

Ordovician volcanics (MAZZUCCHELLI & MINZONI, 1987). The former are intrusive magmatites and show a single geochemical character; on the contrary the latter are effusive volcanics and form a complete suite from rhyolite to basalt (MEMMI et al., 1983). Moreover the Cataclastic Gneiss-Porphyroid shows an alkaline affinity (DELAPERRIERE & LANCELOT, 1989) whereas the middle upper Ordovician volcanics show a clear calc-alkaline affinity (MEMMI & al., 1983). In conclusion, it is reasonable to consider the Cataclastic Gneiss-Porphyroid as a late orogenic magmatism related to a pre-Cambrian orogenesis.
The calc-alkaline affinity of middle Ordovician volcanics represents another open problem. Their geochemical characteristics indicate a late-orogenic magmatic-anatectic association. However, we do not have proof of a true middle Ordovician orogenic cycle. The compressional tectonics related to the Sardic phase probably represent an "epiphenomenon" related to transcurrent movements of the sialic crust. At present we do not know of a regional metamorphism referable to middle Ordovician tectonics. The middle-upper Ordovician calc-alkaline magmatism continues without interruption during Silurian times progressively assuming an alkaline affinity similar to that emplaced within lithosphere plates.
The area distribution of middle-upper Ordovician and Silurian volcanics and the sedimentary evolution in the different areas confirm that this event played a determinant role in modelling the paleogeographic domains. Nevertheless the nature of basic rocks and the lack of ophiolitic sequences indicate that the Hercynian orogenesis involved only sialic crust (MEMMI & al., 1983). This entirely ensialic evolution is not in contrast with the structural style of the Hercynian chain and is analogous to that of collisional chains along continental margins. The crustal subduction took place along multiple shear zones of collided continental margins rather than classical plate tectonics.

7. THE CALABRIAN-PELORITAN MASSIF

The characteristics of Calabrian-Peloritan Massif are:
- The oldest fossiliferous terrains are of upper Cambrian-lower Ordovician age (BOULLIN & al., 1984). These terrains are present in northern Calabria (where they reach the maximum thickness) and in Sicily. The upper Cambrian-lower Ordovician sequences of Calabrian-Peloritan Massif show the same lithology (micaceous siliciclastic sediments) as those of central-southern Sardinia.
- The youngest fossiliferous terrains are of Silurian-Devonian (and in some areas of Carboniferous) age.
- A terrigenous non fossiliferous succession (with subordinate acidic and basic volcanics) outcrops in widespread areas. This non-fossiliferous succession shows the same structural-metamorphic history of fossiliferous successions. Therefore we may reasonably suppose it is of middle-upper Ordovician age (MINZONI, 1988) and it was deformed, together with the fossiliferous terrains, by Hercynian tectonics.
- The main Hercynian tectonics produced isoclinal recumbent folds accompanied by a sub-horizontal flow schistosity and a synkinematic metamorphism. Both the deformation and the metamorphism increase toward deeper levels to reach the greenschist-biotite facies.
- In some areas there are high grade metamorphites that seem to have a structural history that is different from Hercynian metamorphites. It is possible that the high grade metamorphites represent a pre-Hercynian (pre-Cambrian ?) "Basement".

7. CONCLUSIONS

The Calabrian-Peloritan Massif was placed (before the opening of the Mesozoic Ocean) in this way the thick middle Cambrian-lower Ordovician terrigenous succession of SE Sardinia may be connected to upper Cambrian-lower Ordovician terrains of northern Calabria. At the same time, the nappes of SE Sardinia (constiting at their base of thick middle Cambrian-lower Ordovician siliciclastic terrains) probably continues as far as the northern areas of Calabria. Therefore we may complete, utilizing the Calabrian-Peloritan region, the virgation of the Hercynian chain in the Sardinian-Corsican Massif.

In the same way we may connect the deeper structural-metamorphic levels of southernmost Sardinia with the deeper levels of the Calabria region.
This geological framework seems to confirm that SW Sardinia represents a relative foreland of the chain. In fact the deeper levels outcropping in southernmost areas of the island and in central-southern Calabria are characteristic of a shear zone. In other words we may reasonably suppose that another minor shear zone developed south of southern Sardinia during Hercynian orogenesis. The abundance of late Hercynian granitoids in southern Sardinia (CNR, 1987) and in central Calabria seems proof of this hypothesis.
The "Basement" of the Calabrian-Peloritan Massif probably represents an emerged area (from lower Cambrian to middle Ordovician) placed east and south of southern Sardinia.

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METAMORPHIC PHENOMENA RELATED TO FORMATION OF EARLY CRETACEOUS NAPPES IN THE NORTH OF THE MARMAROSH MASSIF (EAST CARPATHIANS)

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Abstract

Local metamorphic changes produced by nappe formation in the Early Cretaceous are distinguished in the north of the Marmarosh Massif of the East Carpathians. Constructive metamorphism of Upper Paleozoic-Mesozoic rocks and retrograde greenschistous alterations in the gneissic-micaschistous complexes are typical. A possibility of relation of these changes with nappe formation is shown.

1. INTRODUCTION

The pre-Mesozoic metamorphic complexes of the Carpathian region are exposed in the West, East and South Carpathians as well as in the Nothern Apuseni Mountains (Fig. 1. A). These complexes form everywhere the basement of the main nappes containing also the Upper Paleozoic - Jurassic sedimentary cover. The age of metamorphic basement is Late Proterozoic - Middle Paleozoic. In addition to Variscan, Salairian and earlier metamorphic events known to occur in the north of the Marmarosh Massif in the East Carpathians, the massif is characterized by much later local metamorphic events which have partly affected the Upper Paleozoic - Lower Mesozoic rocks. Beside, there are some cases of the retrograde transformation of older rocks which took place probably in the Early Cretaceous. Some rocks of the sedimentary cover show also local constructive metamorphism.

Unlike regional metamorphic alterations which have been comprehensively and repetedly described, local metamorphic events related to nappe formation received little attention. Recently obtained information enables us to describe the metamorphic events which occured at the end of the Early Cretaceous. They must have been caused by the formation of tectonic nappes which account for the Marmarosh Massif structure (KHAIN et al., 1968; BERCIA et al., 1976).

The aim of this article is to analyse a possible relation between some cases of local metamorphism of rocks and the nappes formation which took place in the Carpathians in the end of the Early Cretaceous.

2. GENERAL GEOLOGY OF THE REGION

The Marmarosh Massif is the only zone of the East Carpathians in which pre-Mesozoic complexes are widespread. The Rakhov and Chywchyn Mountains in the Soviet Union, the Maramures and Bistritsa Mountains in Romania are northern parts of this Massif.The north-eastern boundary of the Marmarosh Massif contacts the Flysch area of the East Carpathians. The Early Cretaceous flysch complexes have developed near the Massif. The contact represents an overthrust of the north-eastern vergence. The south-western boundary of the Massif is overprinted by the transgressive-lying complexes of the Upper Cretaceous and younger age.

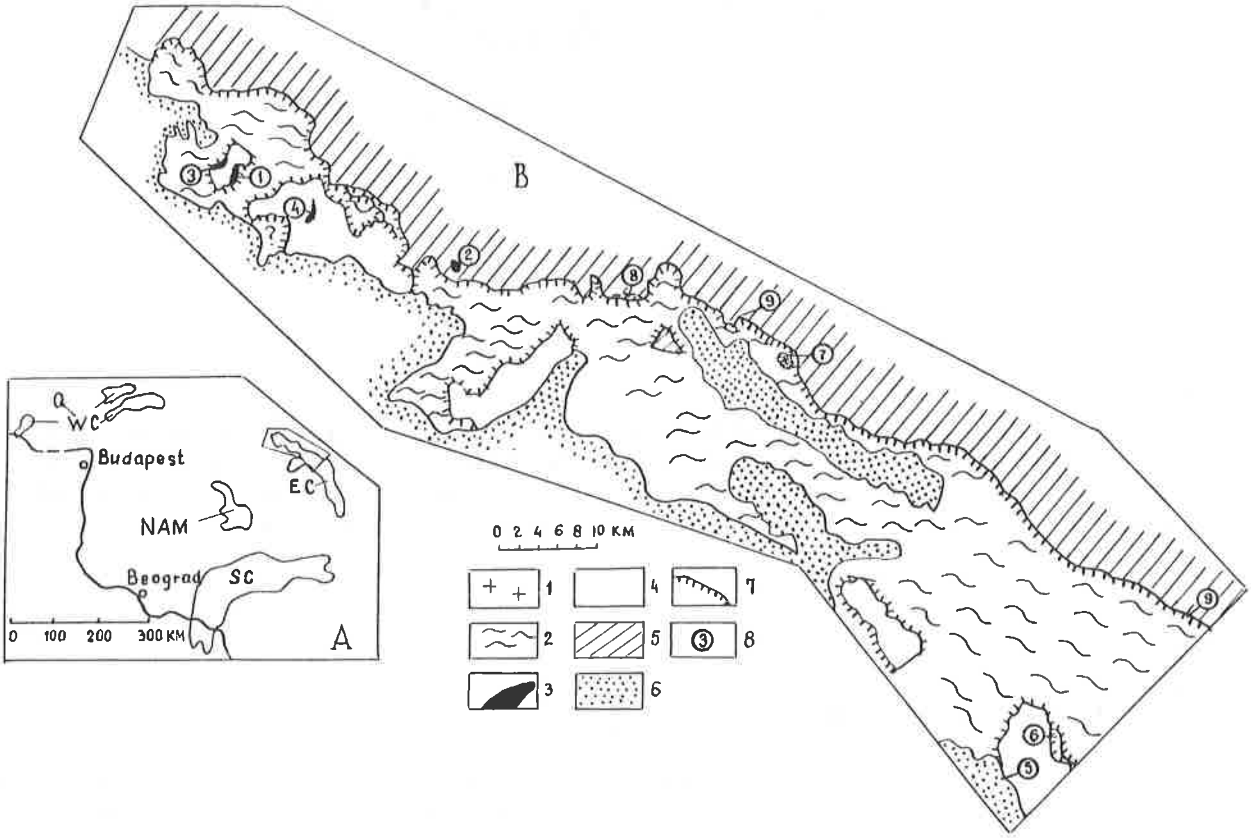

Fig. 1. A - The Position of the investigations area in the Carpathian Region: WC - the West Carpathians, EC - the East Carpathians (the Marmarosh Massif), NAM - the Northern Apuseni Mountains, SC - the South Carpathians.
B - The main tectonic units in the northern part of the Marmarosh Massif: 1 - Divnaya-Rarau, 2 - Delovetskaya-Tulghes, 3 - Rozisskaya-Stinisoara, 4 - Belopotokskaya-Bretila, 5 - Lower Cretaceous Flysch Zone, 6 - Post-Austrian sedimentary cover, 7 - thė boundaries between the tectonic units and zones, 8 - the number of the observation sites.

The main tectonic feature of the Massif is a nappe-fold structure that formed in the end of the Early Cretaceous. The major tectonic units differ not only in peculiarities of the Upper Paleozoic - Mesozoic sedimentary cover, but also in composition of pre-Mesozoic regionally metamorphosed complexes. Three major tectonic units (Belopotokskaya, Delovetskaya, Divnaya are distinghished in the Soviet part of the Massif (KHAIN & al., 1968; BURTMAN & RUDAKOV, 1979). Its analogues in the Romanian part of the Bretila, Tulghes and Rarau units correspondingly (Harta geologica..., 1968). The minor Rozisskaya tectonic scale lies between the Belopotokskaya and Delovetskaya units in the Rakhov Mountains; its analogue in the Romanian part of the Marmarosh Massif was called by Sandulescu (1985) the Stinisoara unit.
The para-autochthonous Belopotokskaya-Bretila unit has a gneiss-amphibolite-micaschist basement and a sedimentary cover including deposits of the Upper Paleozoic, Triassic and Jurassic. The existence of the Upper Permian argilites and the Dovgorun calcareous-pelitic sequence of the Upper Jurassic age is a feature most significant for our investigation. The age of both complexes was argued by macrofauna (Slavin, 1963).
The major part of the Rozisskaya-Stinisoara tectonic scale is represented by a bed of variegated phyllites and fine-grained quartzites. The major allochthonous Delovetskaya-Tulghes unit has a greenschist basement and preserved relics of the upper allochthonous (Divnaya-Rarau unit) containing micaschists, granitoids and augengneisses.
The nappes in the Marmarosh Massif have been formed between the Barremian and Late Albian time. It is proved by the existence of undersurface Neocomian deposits and

an Upper Albian (Vraconian) - Cenomanian complex overprinting the boundaries between the different nappes. Some later movements of the Marmarosh nappes are also presumed (KHAIN & al., 1968). Some Romanian investigators (BERCIA, KRÄUTNER, SANDULESCU) assume a Variscan age for the uppermost Rarau nappe in Romania (Divnaya unit in the Soviet Carpathians). However, the data of BURTMAN and RUDAKOV (1979) indicate that in the Chywchyn Mountains this nappe overthrusts Triassic and Jurassic deposits, providing the Austrian age of dislocations.

3. CONSTRUCTIVE (PROGRESSIVE) METAMORPHISM OF THE UPPER PALEOZOIC AND MESOZOIC ROCKS DURING NAPPE FORMATION

Progressive metamorphic changes occur in the sedimentary cover of the Lowermost Belopotokskaya-Bretila kunit and in the Lower Cretaceous flysch in the front of the Massif.

The structure of the Rozisskaya tectonic scale is a most illustrative example of progressive metamophic changes connected with nappe formation. It was identified for the first time in the Rakhov area of the Marmarosh Massif lying in the USSR (KHAIN & al., 1968). At first the Rozisskaya scale seemed similar to other tectonic units. Later it was found out that this unit occurred only in the north of the Marmarosh Massif, and unlike other units, it did not represent a relic of the primary independent zone. The variegated phyllites and fine-grained quartzites of the Rozisskaya tectonic scale are products of the dynamothermal transformation of Upper Permian variegated argillites and aleurolites which were mechanically withdrawn from the cover of the Belopotokskaya-Bretila unit during the Delovetskaya-Tulghes nappe overthrusting. The original rocks of these metamophic products are widespread in the cover of the lowermost unit in many places of the Massif. The change of initially pelitic rocks consists in the widespread sericite, or more rarely, chlorite development, and in the appearance of easily identifiable foliation, accounted for by the segregation of quartz grains into reciprocally parallel " interbeds". These changes are best observed when comparing the Rozis phyllite beds up to 200-300 m thick with unmetamorphosed Permian pelites belonging to the cover of the lowermost para-autochthonous unit which, just like the Rozisskaya scale, is known to exist only in the northern part of the East Carpathians (KHAIN & al., 1968; Harta geologica..., 1968). The idea of the Rozisskaya scale formation of rocks of the Belopotokskaya-Bretila unit cover is confirmed by the presence of similar Middle Triassic dolomites in both sheets of the cover. The Stinisoara unit situated at the right bank of the Vaser river (SANDULESCU, 1985) has a similar structure. But it is an isolated outlier tectonically overlying the Lower Cretaceous flysch trimming the Massif, whereas the Rozisskaya scale is squeesed between the major units of the Massif.

Nappe formation in the Early Cretaceous time is characterized by isolated cases of transformation of Upper Jurassic calcareous-pelitic rocks belonging to the Belopotokskaya unit cover into chlorite schists. One of the examples was recorded at the Dovgorun stream where it was reliably dated by ammonites and corals in lenses of metamorphosed limestones (SLAVIN, 1963). The essential metamorphogenic mineral is magnesian chlorite, optically similar to clinochlore. There are also sericite (paragonite) and more rare talc and tourmaline. The neoformed coarse-grained texture of schists is typical. The thickness of zones made up of chlorite schists does not exceed 10-20 m. In other places of the same unit, primary pelite rocks remained unmetamorphosed. Similar changes have been recently recorded by BALINTONI (1985) in the Mesozoic rocks of the cover of the lower unit near the Sukherezul-Mare stream, north of Yakoben town.

Another object that had undergone progressive metamorphic changes connected with nappe formation turned out to be terrigeneous rocks of the Lower Cretaceous flysch of the Massif close surroundings. The Massif is represented near the front by the Delovetskaya-Tulghes nappe. The primary sequence of the flysch complex consists of sandstones, aleurolites and argilites. The change of these rocks observed by the author at the left bank of the Chernyy Cheremosh river, at the right bank of the Mascotin stream in the Chywchyn Mountains, and in the north of the Romanian part of the Marmarosh Massif consists in phyllonitization and mylonitization of flysch rocks and

may be associated both with nappe formation in the Massif and with later north-eastward overthrusting of the Massif upon the flysch complexes. The metamorphogenic mineral assemblage exists only in argilites and is represented there by fine-grained sericite, chlorite and sometimes by minor kaolinite.
Anyway, the development of subhorizontal shifting of large masses of ancient regionally metamorphosed rocks dominates among local metamorphic events. It is to be noted that the intensity of metamorphic alterations in the Lower Cretaceous flysch was notably lower. That might have been due to the increased pressure, whereas during transformation of the para-autochthonous massif unit, rocks in the sedimentary cover have undergone recrystallization parallel to the appearance of new mineral associations and new structures.
The metamorphosing influence of nappe formation is also observed in the sedimentary cover which was not withdrawn from the basement of the Belopotokskaya-Bretila unit. In the Rakhov Mountains, in the upper course of the Belyy Potok and Bredetsel streams, the pelitic rocks of the Belopotokskaya unit cover are in some cases transformed into chlorite and sericite schists and phyllonitized with subhorizontal orientation of structures.

4. RETROGRADE TRANSFORMATION OF OLDER METAMORPHIC ROCKS

Retrograde metamorphic changes are expressed only in the gneissic-micaschistous basement of the Belopotokskaya-Bretila and Divnaya-Rarau units. Its major components are gneisses, different micaschists, para- and orthoamphibolites and granitoids. The original mineral assemblages of these rocks contain quartz, biotite, muscovite, amphiboles, garnet, hornblende, pyroxenes, potash-feldspars and different plagioclases. The metamorphic alterations connected with nappe formation are represented by changes of mineral assemblages and of rock structures. Latest garnet, hornblende and biotite chloritization as well as epidotization of potash-feldspar are observed here in the basement rocks, though some of the diaphthoric changes might not be related to the nappe formation. According to BALINTONI (1984) the nappe formation gave rise to the development of mylonitization and diaphthoresis in gneisses and micaschists in the paraautochthonon upper part directly under the allochthonous greenschist substrate. Such case was recorded at the Bistritsa right bank between Vatra-Dorney town and Chiokanest' village. Diaphthoric changes may similarly occur in the upper part of the metamorphic basement of the Belopotokskaya unit within the massif area lying in the USSR terretory. Locally the phyllonitization of gneisses is also developed here. The thickness of its zone amounts to several meters, the maximal depth of erosion in the Belopotokskaya unit basement is known to reach several hundreds of meters.
The so-called "Argestru series" is an example of most intensive metamorphic reworking connected with the nappe formation. It was for the first time identified as a stratigraphic subdivision in the Bistritsa Mountains (BERCIA, 1970) with proposed Late Paleozoic age (?). As to its tectonic position it was interpreted as a scale between two large nappes. Later BERCIA & al., (1976) recorded cases of transgressive position of the Argestru series on the older substrate, the latter made possible to date it as Carboniferous (?). The presence of biotite and hornblende in the rocks was considered due to deep metamorphism during the Alpine nappe formation. The recently available data have made BALINTONI (1984) change his conceptions about "Argestru series". The absence of stratigraphic contacts with the complex basement has been established, and the relations with gneissic-micaschistous substrate proved to be tectonic throughout the series. It was discovered that almandine, biotite, hornblende, muscovite and oligoclase in the rocks of the "Argestru series" are not new-formed, and represent relics of more ancient metamorphogenic mineral association. As a result among the rocks of this complex there have been identified deeply diaphtorized gneisses, amphibolites and quartzites, which have been parts of the complex metamorphosed in the conditions of almandine amphibolite facies. Quite logical would be an assumption that before the formation of nappes these rocks belonged to the Bretila series and it is wrong to

interpret the "Argestru series" as a stratigraphic subdivision. This rock association should be regarded as deeply diaphthorized material, which makes up a tectonic scale, 250-300 m thick (BERCIA, 1970) between two large nappes, involving the metamorphic basement. The Kirilen' scale "torn off from the Bretila series upper parts" has been formed in the same way near Sharul-Dorney village in the Bistritsa mountains (BALINTONI, 1985).

The diaphthoric changes recorded also in gneissic schist basement of the upper Divnaya-Rarau nappe might have been due to a certain extent to the nappes formation. Detached from the paraautochthonous Belopotokskaya-Bretila unit the diaphthorized basement is in some places moved as an independent structural element to the massif front. Such relations were observed by the author in the upper course of the Ryzhovatyy stream, at the Chernyy Cheremosh river left bank (RUDAKOV, 1969). A complex of rocks, belonging to the Belopotokskaya unit cover and basement has been revealed here directly beneath the Delovetskaya nappe greenschist basement. In addition to greenschist diaphthoresis occurred in garnet micaschists and in gneisses this area is characterized by subhorizontal plan-parallel structure clearly cutting the foliation of regionally metamorphosed rocks. The visible thickness of the alteration zone does not exceed 1-2 m. It is to be noted that in this area the Delovetskaya nappe has a wide, up to 9 Km, practically horizontal "cap peak" near the front with Lowe Cretaceous flysch beds developed under it and isolated scales torn off from the Belopotokskaya unit are recorded in some cases. The analogous element in Romania is the Stinisoara scale (SANDULESCU, 1985), also underlying the allochthonous greenschist basement but it was not identified near the massif front but in the erosion tectonic window, about 7 Km south-east of the Ryzhovatyy stream.

5. SUMMARY OF RESULTS AND OBSERVATIONS. DISCUSSION

The relation of the above-mentioned metamorphic changes to nappe formation is indirectly confirmed by the following evidence. First, locally metamorphosed rocks of the cover and the basement are developed either in the para-autochthonous upper parts or they make up intermediate plates, e.i. they are confined to the boundaries between the structural units composing the surface of tectonic nappes. Second, some diaphthorized rocks of the basement have been radiologically dated yielding close to the time of nappe formation. For example, the K-Ar age of 110+or-11 and 128+or-7 Ma has been determined for the amphibolites of the Rakhov Mountains (BOYKO & al., 1966). Similar ages have been obtained also in the Romanian part of the Massif. Third, it is to be noted that the sedimentary rocks of the cover of the main allochthonous Delovetskaya-Tulghes unit have not been affected by metamorphism.

1. The above-mentioned information indicates that some local metamorphic alterations of rocks in the north of the Marmarosh Massif of the East Carpathians can be connected with nappe formation. However, it needs to be underlined that not all retrograde transformations of older rocks and constructive metamorphic changes of the sedimentary cover are of a such origin.

2. The indicated metamorphic changes that turn out to occur near the surface of tectonic nappes and are evidently related to their formation, may be characterized by the following features (Fig. 2):

-a) the subject of the Early Cretaceous metamorphism was either the sedimentary cover, or an older metamorphic basement, or both elements together;

-b) metamorphism is distinguished in the upper parts of the para-autochthonous or in the sole of the gneissic allochtonous. There are also situations in which metamorphic changes affected the whole tectonic unit. Seven of nine possible combinations have been characterized by concrete examples. In addition to the above, local metamorphism has been recorded in the Lower Cretaceous flysch near the Massif front;

-c) the metamorphic alterations related to the formation of Early Cretaceous nappes in the north of the Marmarosh Massif are not distinguished in the greenschistous basement of the major allochthonous unit, for the thermodynamic conditions during this process were similar to the P-T conditions of regional metamorphism of this basement. However, the nappe sliding traces may be represented by small (not exceeding 10 m in thickness) phyllonitization and mylonitization zones in their lowermost parts, observed both in the Soviet and Romanian parts of the massif. In the Vatra-Dorney - Yakoben' region the greenschisct allochthon is divided into isolated scales with a superimposed plan parallel structure conformable to the nappe basement that can be traced in the lowermost part (up to 600 m thick) (BERCIA, 1970).

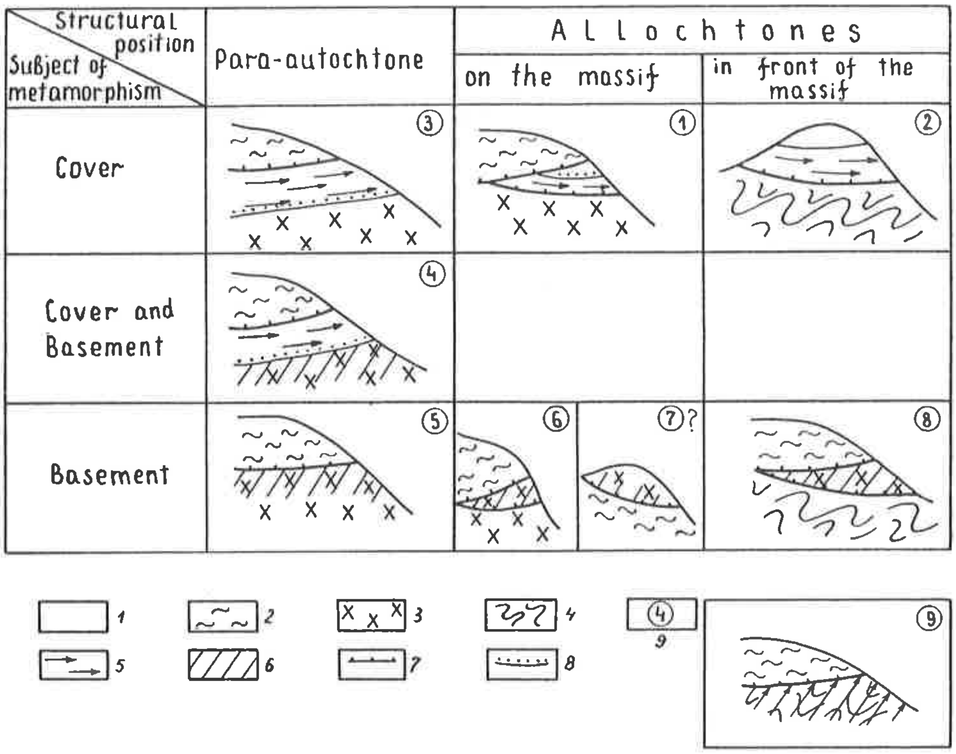

Fig. 2. The system of local metamorphic phenomena related to nappe formation in the north of the Marmarosh Massif. The progressive metamophic alterations (5) are located in the sedimentary cover (1) of the different Massif units and in the Lower Cretaceous flysch (4) from the massif framework. The retrograde transformations (6) are distinguished in the gneissic-micaschistous basement (3). The greenschistous basement (3) has no divisible metamorphic phenomena related to nappe formation. 7 - boundaries between tectonic units, 8 - stratigraphic contacts between the sedimentary cover and its metamorphic basement, 9 - numbers of the observation sites (according Fig. 1 B): 1 - the Rozis stream, 2 - Stinisoara stream, 3 - Dovgorun and Sukherezul-Mare streams, 4 - upper stream of the Belyy Potok and Bredestsel, 5 - Bistritsa river, 6 - Argestru stream and Kirilen' village, 7 - Chernyy Div and Rarau mountains, 8 - Ryzhovatyy and Stevioara-Mica streams, 9 -the Czywczyn Mountains and the north part of the Romanian Marmarosh.

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THE DEVONIAN BASIC INTRAPLATE VOLCANISM OF THE NORTHERN GRAYWACKE ZONE, EASTERN ALPS AND ITS RELATION TO CRUSTAL EXTENSION.

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Abstract

Detailed geological mapping, profiling as well as volcanological and geochemical investigations within the Paleozoic of Northern Graywacke Zone provided numerous new data concerning the basic volcanism. At the base of 410 geochemical analyses a new and far more detailed geotectonical model is presented. Within the strongly deformed and polyphase metamorphic Eastalpine Paleozoic realms different volcanological and geochemical rock provinces can be distinguished, not coinciding with the tectonical thrust sheets and slices formed during the Alpine orogeny.
The metabasalts are partly of Silurian, partly of Devonian age. For many sequences the age is still unknown. Most of the basic magmatites originated in an intraplate environment. Several phases of seamount growth and ocean island stages, as shallow water and subaerial conditions, are depicted by means of volcanological investigations. These rock types conincide with a transitional basalt and alkalibasalt geochemistry, enriched in immobile incompatible elements.
An exception is the basalt-sill complex of Maishofen, where the basic magmatites formed under deep water conditions. The geochemical composition of these rocks is ambiguous: an initial stage of back-arc spreading connected with an incipient, remote subduction zone is discussed for this peculiar region. Following recent mapping evidences, the age is assumed to be Carboniferous.
All plate tectonic models dealing with active plate margins during Silurian or Devonian times must be rejected for the Northern Graywacke Zone. In the authors' opinion there is no need to assume a Caledonian collisional stage. The widespread Upper Ordovician acid colcanism, as documented by the Blasseneck Porphyroids, is also not necessarly attached to a subduction zone. The hypothetical angular unconformity at the base of the Paleozoic sequences is undoubtedly older than Upper Ordovician. There is evidently no link to the Caledonian collision and Iapetus-closure. Therefore the term "Panafrican" should be used for Cambrian and Ordovician events in the Alps.
Crustal extension pertaining throughout almost the entire Lower Paleozoic is supported by stratigraphical, sedimentological, geochemical and volcanological data. Only in the uppermost parts of the rock sequences (Carboniferous ?) weak indications for subduction and collision are found.
Facies heterogeneity at a small scale is usual, biostratigraphical data and tectonically unaffected sequences are rare. So it is misleading to construct and publish audacious global models at the base of scarce data sets, as it was usual during the last 10 years of geochemical investigation within the Alpine Paleozoic.

1. INTRODUCTION

The northern Graywacke zone is one of the most important Paleozoic realms within the Eastern Alps (Fig. 1). It belongs to the Upper Austroalpine nappe system and can be regarded as the primary autochthonous basement which is, divided by an angular unconformity, transgressively overlain by the Permo-Mesozoic cover of the Northern Calcareous Alps. The Paleozoic rocks have undergone polyphase metamorphism during Variscan and Alpine orogeny, as well as different phases of nappe stacking, wrench faulting and extension under varying strain regimes. Within this complex mosaic pattern of dismembered Paleozoic units, stratigraphical sequences were mapped and in several cases also dated by fossils. The state of art concerning the stratigraphical record is summarized in SCHÖNLAUB (1979, 1982) and OBERHAUSER (1980).

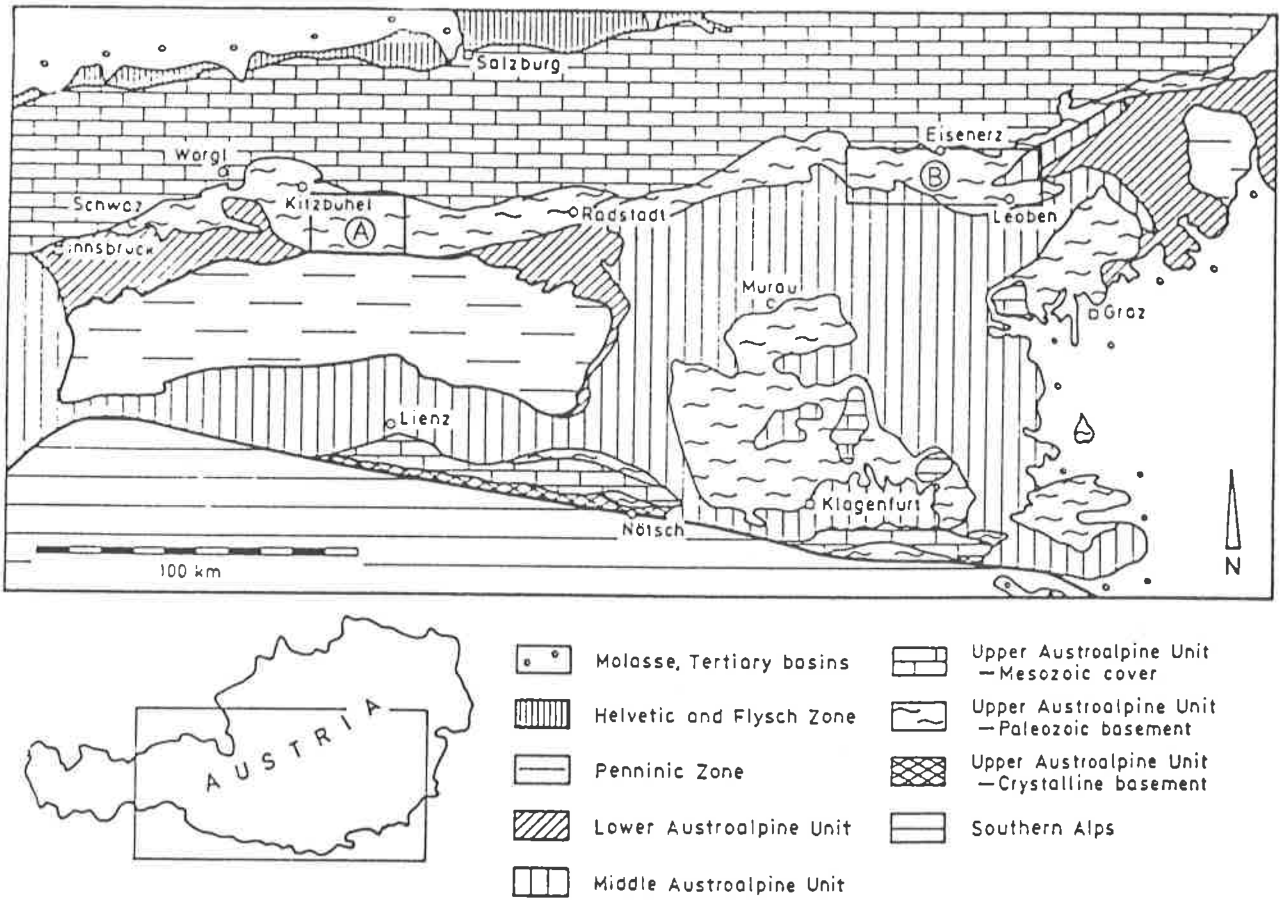


Fig. 1. Location of working areas within the Eastern Alps; A Western Graywacke Zone, B Eastern Graywacke Zone.

In the context of geotectonical modelling volcanic rocks are of special interest. This paper deals with the basic magmatic sequences. Within the Western part of the Graywacke Zone (area A, Fig. 1) they sometimes reach a maximum thickness of 400m and thus represent the thickest and best preserved Paleozoic metabasalts in the Eastern Alps.

At two localities within the Northern Graywacke Zone the age of these basic volcanites was determined with certainty by fossils:

- Kitzbühel Graywacke Zone (western part, area A - Fig. 1) as Devonian (HEINISCH et al., 1987).
- Eisenerz Graywacke Zone (eastern part, area B - Fig. 2) as Silurian (SCHÖNLAUB 1982).

All other ideas about the age of the metabasalts within the Northern Graywacke Zone, as found in many papers (MOSTLER 1968, 1984; HÖLL & MAUCHER 1976, LOESCHKE 1977, OBERHAUSER 1980, COLINS et al., 1980, FRISCH et al., 1984, TARKIAN & GARBE 1988), are based on assumptions and do not withstand critical review. Also within the regions mentionned above, only a minor part of sequences is dated up to now. So it can not be excluded, that basaltic rocks of different age but identical volcanological and geochemical character might exist within these areas.

A lot of plate-tectonic models has been published in the last years, either based on the basic volcanism of the Graywacke Zone or at least including it into the hypothetical ideas. The leading idea was the connection of this volcanism with and active plate boundary. Among others LOESCHKE (1977), FRISCH et al., (1984) and NEUBAUER

& FRISCH (1988) favoured the idea of subduction mechanisms and plate collision, COLINS et al., (1980) and MOSTLER (1983) established the model of a mid-oceanic ridge system with an mature sequence of oceanic crust. A complete summary of the different working hypotheses is given in SCHLAEGEL-BLAUT (1990) and HEINISCH (1986, 1988).
Aware of these contrasting ideas, detailed volcanological and geochemical investigations were carried out. Field mapping, detailed volcanological profiling and geochemical sampling covered all major outcrops in a statistical representative way and led to a large data base of 410 analysed samples. These efforts enabled us to develop a new, more reliable and much more detailed geodynamic interpretation of the basic magmatism. This paper was written under the purpose of giving a concise summary concentrating on the models. The complete data set and detailed discussions are published in SCHLAEGEL-BLAUT (1990).

2. VOLCANOLOGY

Though low grade metamorphism affected the rocks for several times, primary magmatic textures are quite well preserved in some cases. The comparison with modern volcanic sequences made it possible to identify many relics, as pillows, hyaloclastite layers, pumice lapilli, angular scoria and epiclastic debris.

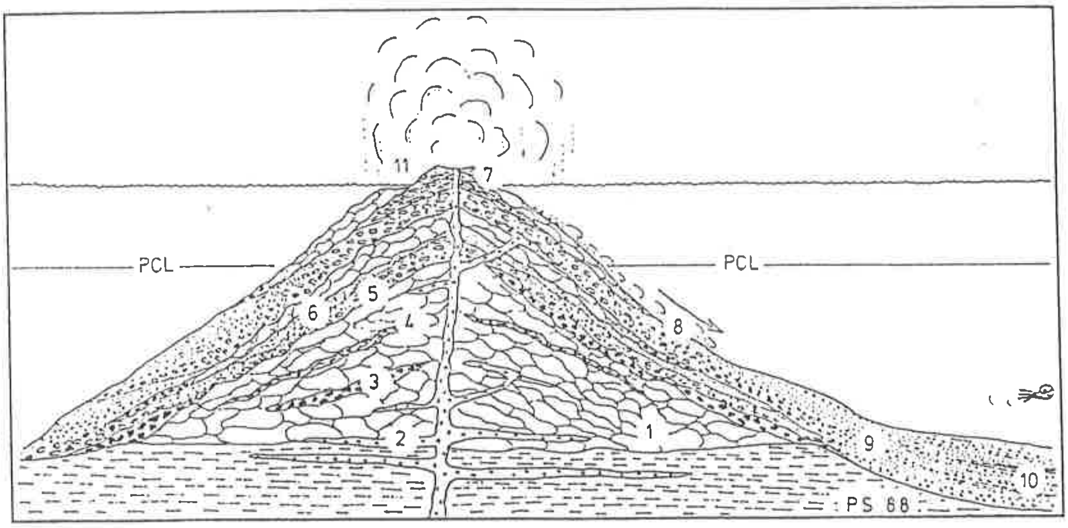


Fig. 2. Model of an island volcano; all characteristics mentioned here can be found within the basic magmatic sequences of the Northern Graywacke Zone. Explanation see text.

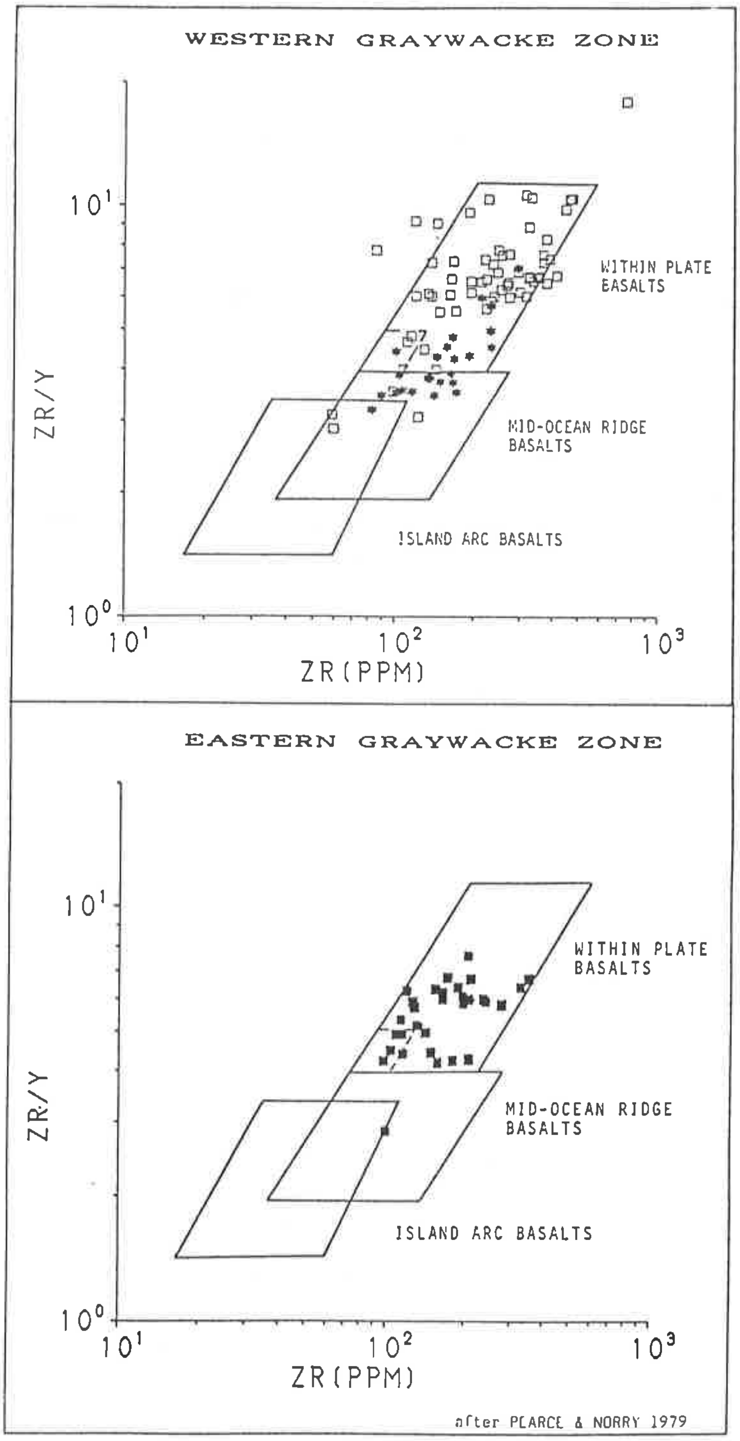


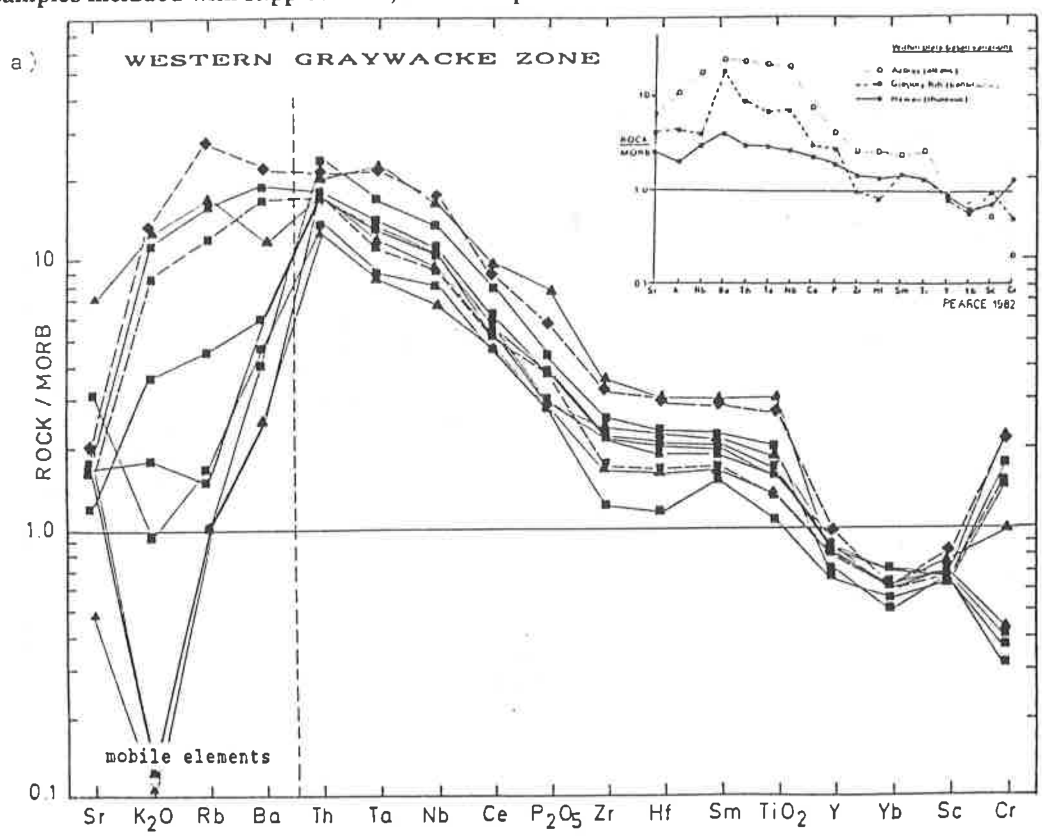
Fig. 3: Plate tectonic position of metabasalt lavas of the Northern Graywacke Zone (western part n=87; eastern part n=30).

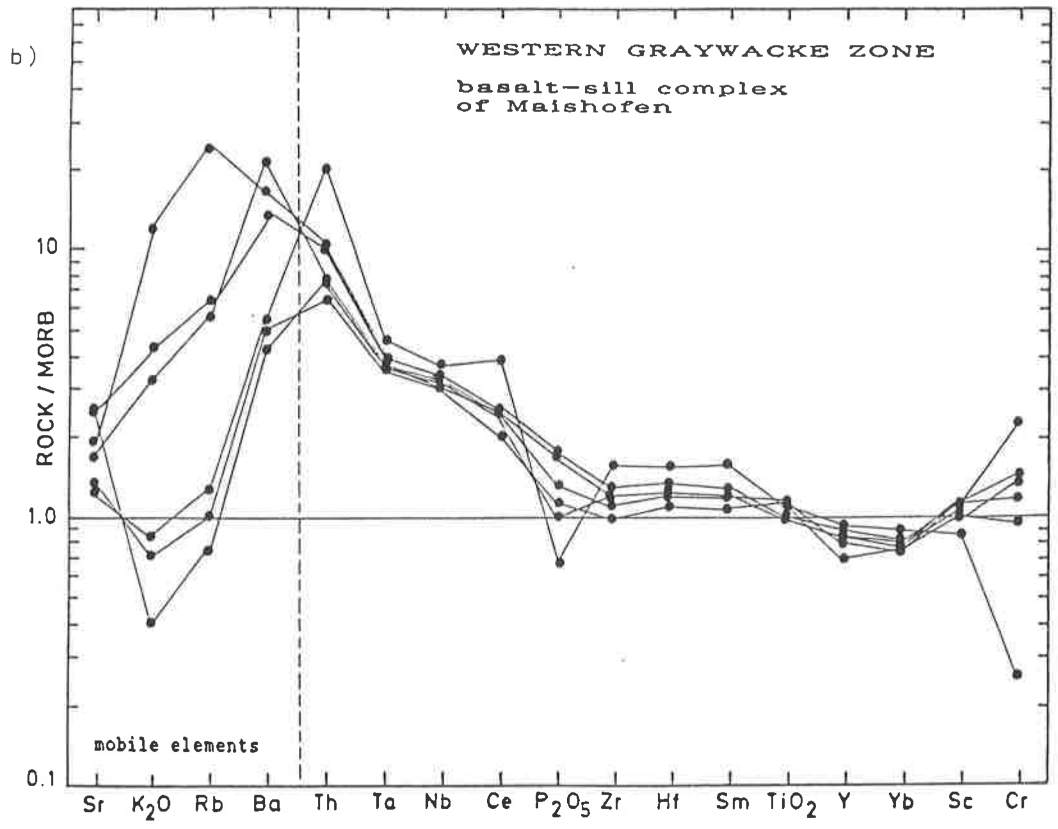
The combination of these results, derived from a lot of different profiles, enabled us to establish the model of island volcanoes (Fig. 2). Within the volcanic edifices, the following stages of evolution can be reconstructed:
- A pillow volcano is built up by sequences of pillow basalts (1) and gabbroic sills (2). Below the PCL (pressure compensation level, FISHER 1984) only small volumina of hyaloclastites (3) and pillow breccias (4) occur as pyroclastic rocks. In the Northern Graywacke Zone sequences of highly vesicular pillow basalts reach a maximum thickness of about 350m.
- By reaching the PCL, explosive volcanism becomes dominant. Different kinds of pyroclastic rocks erupt, as lapilli tuffs (5) and ash tuffs (6). The proximal scoria (7) and unsorted tuffs (8) prove a deposition very close to an distal position to the vent. At the slopes and within the basins between volcanic edifices, reworked volcanic debris (tuffites (8), (9), (10) is deposited. The pyroclastic sequences reach a maximum thickness of about 400m.
- Highly vesicular pumice lapilli and well rounded beach boulders (11) prove temporary subaerial stages.

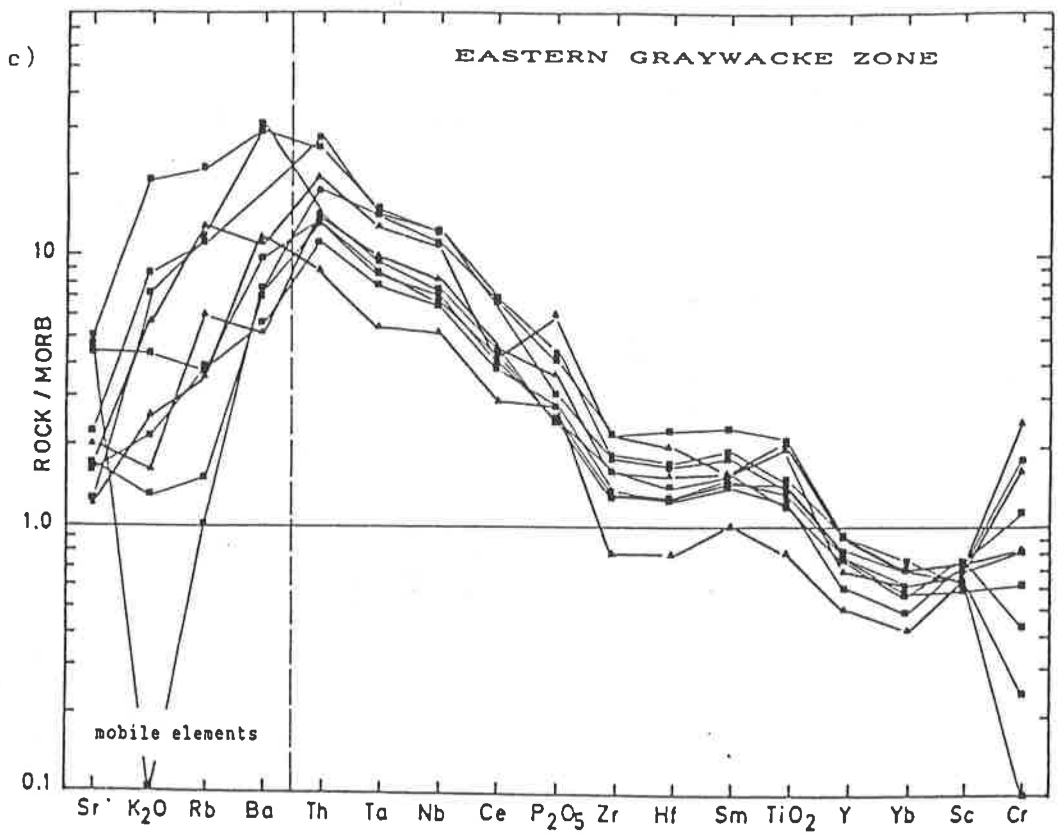
The occurence of widespread and thick pyroclastic sequences in the Northern Graywacke Zone shows that most of the eruptions took place above the PCL. This determines a maximum depth of water of 500m for extrusion of most of the basic volcanites (FISHER 1984, WILLIAMS & McBIRNEY 1979). So the scenario of growing seamounts, temporarily reaching a shoaling stage or even stable oceanic island stage is established.
In the Maishofen area, a small part of the Western Graywacke Zone, a different situation is found: Thick sequences of pillow lavas (with few or no vesicles), sheet flows and shallow level sub-sediment gabbroic sills are abundant (basalt-sill complex of Maishofen, SCHLAEGEL-BLAUT 1990); products of explosive volcanism are missing. The volcanological investigations prove an extrusion under deeper-water conditions, probably below the PCL. This result is in good concordance with the findings of COLINS et al., (1980). Further consequences are discussed within the ambit of geotectonical modelling (see Fig. 5).

3. GEOCHEMISTRY

Fig. 4. Element patterns of metabasalt lavas and some gabbroic sills of the Northern Graywacke Zone: a) western part with intraplate position n = 10, b) Maishofen area n = 6, c) eastern part n = 10 (Two highly altered samples included with stippled lines, further explanation see SCHLAEGEL-BLAUT 1990).




b)
WESTERN GRAYWACKE ZONE
basalt-sill complex
of Maishofen
ROCK / MORB
10
1.0
0.1
mobile elements
Sr K_2O Rb Ba Th Ta Nb Ce P_2O_5 Zr Hf Sm TiO_2 Y Yb Sc Cr


c)
EASTERN GRAYWACKE ZONE
ROCK / MORB
10
1.0
0.1
mobile elements
Sr K_2O Rb Ba Th Ta Nb Ce P_2O_5 Zr Hf Sm TiO_2 Y Yb Sc Cr

A total sum of 410 samples was analysed with XRF for 22 main and trace elements. In addition, 44 representative samples were selected for rare earth element analyses. The INAA was carried out by J. HERTOGEN (Univ. Leuven, Belgium). Further information is given in HEINISCH et al., (1988). The data presented here are restricted to 117 samples of metabasalt lavas and 26 REE patterns. The whole data set is published in SCHLAEGEL-BLAUT (1990).
All samples show petrographically and geochemically distinct alteration features. Therefore no mobile elements are used for the geochemical interpretation.
Most of the rocks are transitional basalt and alkali basalts, subordinate some tholeiites occur. All samples are enriched in immobiles incompatible elements (P, Zr, Nb), particularly in Ti and the light rare earth elements (LREE). Discrimination diagrams and typical element ratios (e.g. $Hf/Ta<2$; WOOD et al., 1979) lead to the conclusion that the rocks formed in a within-plate geotectonic position (Fig. 3). There is a good accordance with the typical element patterns of ocean-island basalts, e.g. strong enrichment of the elements Ti, P, Zr, Nb, Th and Ta relative to an average MORB composition (PEARCE 1982, SAUNDERS 1984, Fig. 4). As source rock lherzolitic mantle material, primarily enriched in incompatible elements is suggested (SCHLAEGEL-BLAUT 1990). The major part of the basic magmatites in the Northern Graywacke Zone belongs to these intraplate magmatites.
Apart from the volcanological features mentioned above, also the geochemistry of the basalt-sill complex of Maishofen differs clearly from the other localities within the western Graywacke Zone. The geochemical characteristics of these rocks are quite ambiguous, because they are less enriched in incompatible elements and show affinities to MORBs and intraplate basalts as well as to island-arc tholeiites (Fig. 3, 4).
Obviously, the basalt-sill complex is related to an independent magmatic event. Recent field work shows that this complex should be younger than the Lower Devonian intraplate volcanism (Carboniferous ?; HEINISCH & SCHLAEGEL 1989). Unfortunately, up to now no direct biostratigraphic dating of this sequence was possible.

4. GEOTECTONIC MODEL

The data presented here require a revision of the various plate tectonic models for the Alpine Paleozoic. The basic magmatism is not connected with the supposed Caledonian movements as favoured by LOESCHKE (1977), MOSTLER (1983, 1984) and FRISCH et al., (1984). It is related to crustal extension in the Devonian. The overwhelming majority of the examined rocks displays neither a relationship to an active consuming plate margin, nor to a mature ocean ridge. Only the basalt-sill complex of Maishofen does not match with this model.
In the new genetic model presented, the basic intraplate volcanism is correlated with volcanic edifices (e.g. seamounts, island volcanoes) in a shallow marine environment. This is in good accordance with the sedimentological data (HEINISCH 1986, 1988) which prove an adjacent passive continental margin. As in the Variscan belt of Central Europe in general, in the Alpine realm there is no evidence for a completely developed ocean. The Silurian-Devonian basic intraplate magmatism of the Graywacke Zone represents a phase of extensional tectonics and is probably in connection with mantle-plume mechanism.
A model for the geotectonic history of the basic magmatism of the Graywacke Zone is presented using data from the well studied Western Graywacke Zone (HEINISCH 1986). The following stages can be distinguished (Fig. 5).:

4.1 Lower Ordovician (?) till Upper Ordovician

Linked to extension of thick continental crust, clastic sediments are deposited in an epicontinental shallow marine basin. It is up to speculation, if this onset of crustal thinning is triggered by rising mantle plumes or by back-arc spreading mechanisms connected with a remote Panafrican subduction zone (HEINISCH 1988).
Near the Caradocian/Ashgillian border a widespread volcanic event occurs. Partly under subaerial conditions ignimbrites (Blasseneck Porphyroids) of mostly rhyolitic to alkalirhyolitic composition extrude over the entire Graywacke Zone and other Paleozoic

realms of the Alps (HEINISCH 1981). In some cases these extrusions formed large volcanic highs, in region distal from the fissure vents thin volcaniclastic intercalations within siliciclastic sediments occur. Afterwards, the porphyroid platforms were eroded, pertaining crustal thinning and subsidence enables the evolution of different facies zones.

4.2 Silurian to early Lower Devonian

Siliciclastic sediments accumulate to thick sequences. Submarine fans develop and huge amounts of debris are transported by turbidite mechanisms. This sedimentary basin is called Glemmtal-Unit (Fig. 5). Distal and proximal parts of these fans are interfingering within short distances. The sandstone composition and the lithology of breccia components (channel fillings) prove a continental source area. On top of a deeper marine rise, in a position remote from the marine fans, a thin condensed carbonate-lydite complex forms (HEINISCH et al., 1987).
Paleogeographically separated from the turbidite basin, on top of the former ignimbrite plateaus, a shallow water carbonate platform develops, mainly in the Uppermost Silurian and Devonian (Wildseeloder-Unit, Fig. 5).

4.3 Late Lower Devonian to Middle Devonian

Within a short time interval the submarine basic magmatism of within-plate type spreads out in the marine basin. Pillow volcanoes and thick pyroclastic sequences form volcanic edifices with temporary subaerial stages. Gabbroic sills and little intrusive bodies occur in the underlying rocks. The basic volcanites occur presumably within the distal turbidite facies environment. Finally also the condensed carbonate-lydite complex is covered and buried.

4.4 Upper Devonian to Lower Carboniferous (?)

Following the basic intraplate magmatism, the siliciclastic sediments of the proximal turbidite facies cover the whole area of Glemmtal-Unit. Coarsening-upwards sequences prograde into the distal facies realms. This change in sedimentation probably marks the end of the passive-margin situation.

4.5 Lower Carboniferous (?)

Increasing synsedimentary tectonic movements cause a pronounced submarine relief with a quick change in the depth of water. Marine highs are eroded. In a graben position under deep-water conditions, the basalt-sill complex of Maishofen develops.

Highly hypothetically these last events can be interpreted as the onset of subduction and reorganization of a passive continental margin to an orogenic deep-sea trench with flysch facies. Under this aspect, the basalt-sill complex of Maishofen could be seen as a part of an embryonal marginal basin with incipient back-arc spreading, as described for the Bransfield Strait by WEAVER & al., (1979). On the other hand a position far behind a subduction zone is possible, as reported for the West Philippine Basin by TARNEY & al., (1981). Even a continental rift system, like the Gulf of California (SAUNDERS & al., 1982) would fit to the data. The last alternative was proposed simultaneously to our investigations for the Western Graywacke Zone by TARKIAN & GARBE (1988). Keeping in mind that the geochemical patterns are ambiguous and the stratigraphical position is unknown, this question is open to speculation.

5. CONCLUSIONS

Focussing on the data from the Northern Graywacke Zone, there is no need to demand for a Caledonian collision in the Alpine Paleozoic as proposed by FRISCH & al., (1984), NEUBAUER & FRISCH (1988) or LOESCHKE (1989). During the time of interest, within the geological record no angular unconformity or even sedimentary hiatus occurs. Also there is absolutely no proof for the existence of a well-developed Ordovician, Silurian or Devonian ocean floor within the Alpine realm. Therefore these models must be rejected. In the investigated area a small-scale variability in volcanological and

geochemical characteristics occurs. So it must inevitably lead to confusion, if small data sets of unknown stratigraphy are used to establish a geotectonical model, claimed by some authors to be valid for the whole Alpine Paleozoic. Much work still must be done to separate data speculations.
Regarding the latest publications dealing with volcanites from the Eastern Graywacke Zone (LOESCHKE 1989), Magdalensberg-Series (LOESCHKE, in press), Eisenkappel area (LOESCHKE & SCHNEPF 1987) Eisenhutschiefer-Series of Gurktaler Alpen (GIESE 1989), Sausal Paleozoic (SCHLAMBERGER 1988) and the Paleozoic of Graz (FLÜGEL et al., 1988; FRITZ & NEUBAUER 1988) harmoniously the model of intraplate basaltic volcanism related to crustal extension is established for the time space of Silurian and Devonian. This idea was first proposed by KOLMER (1978) for volcanic rocks from Styria and then propagated by HEINISCH et al., (1988) based on data from the Western and Eastern Graywacke Zone. In our opinion at the moment this idea reflects the best-fitting picture with all available data at least for Silurian and Devonian time.

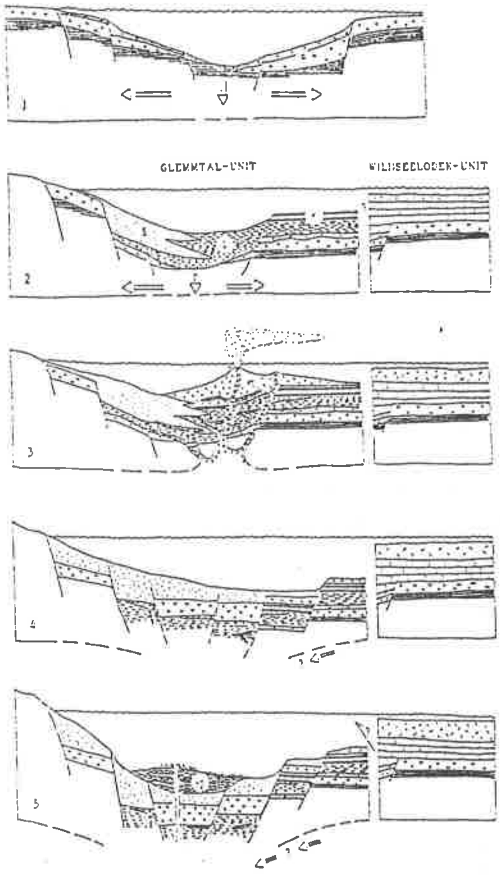


Fig. 5. Scenario of geotectonic history of the Western Graywacke Zone, including volcanological as well as stratigraphical and sedimentological data (HEINISCH 1986). P = Blasseneck Porphyroids, S = proximal turbidite sediments of Schattberg-Formation, L = distal turbidite sediments of Löhnersbach-Formation, K = condensed carbonate-lydite complex of Klinger-Kar-Formation, V_1 = basic intraplate volcanism, V_2 = basalt-sill complex of Maishofen.

The Ordovician events are more controversial (Stage 1 in Fig. 5). The Blasseneck Porphyroids again are taken as proof for an upper Ordovician active continental margin and continent-continent collision by LOESCHKE (1989). In the authors' opinion this cannot be the case, as angular unconformities are missing in the sedimentary record. The conglomerate layers found by NEUBAUER (1985) and interpreted as angular unconformity at the base of the Paleozoic sequences are at a wrong stratigraphic position, evidently older than the Porphyroids. The partly calc-alkaline chemistry of the upper Ordovician acid volcanism can also be understood as anatectic melting of thick continental crust that perhaps formed during a Panafrican collision. There is no time relation between the Alpine events and the Caledonian collision and Iapetus closure in Northern Europe. This did happen in the Late Silurian, followed by Old Red molasse sedimentation.
Therefore the term "Caledonian" should be avoided for Cambrian or Ordovician events in the Alps. We propose to use the term "Panafrican", as usual in other parts of continental crust linked to Gondwana at that time.
As far as we can see now, converging plate boundaries and collisional tectonics within the sedimentary record can not be proven until the Carboniferous age. The final result corresponds with the well-established Pangaea stage.

AKNOWLEDGEMENTS

We gratefully acknowledge the generous support by the Deutsche Forschungsgemeinschaft from 1983 to 1986. Detailed investigations in the Styrian part were enabled by a fellowchip of the ARGE Alpen-Adria. The Geologische Bundesanstalt Vienna continuously sponsored the field work in the Western Graywacke Zone.

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LATE PALEOZOIC AND MESOZOIC TECTONIC EVOLUTION OF THE MIDDLE EASTERN TETHYSIDES: IMPLICATIONS FOR THE PALAEZOIC GEODYNAMICS OF THE TETHYAN REALM

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Abstract

This paper reviews the late Palaeozoic and Mesozoic tectonics of the Middle Eastern Tethysides in terms of a new tectonic model, whose main tenet is to regard the late Palaezoic to late Triassic basements of the Pontide/Dzirula/Adzharia-Trialeti/Artvin-Karabagh/Sanandaj-Sirjan zones collectively as a NNE-facing Cimmeride magmatic arc, called the Podataksasi zone whose Jurassic-Cretaceous movement with respect to Eurasia caused much of the coeval deformation in Transcaucasia, southern USSR E of the Caspina Sea, and in Iran. The model suggests that the northern margin of Gondwana-Land in the Middle East was not a passive, Atlantic-type continental margin during the late Palaeozoic and the early and middle Triassic, but instead was an active, Pacific-type margin, formed by the Gondwania-ward subduction of the floor of Palaeo-Tethys. Successive calving of island arcs of this margin formed the Neo-Tethyan oceans. Very large magnitude strike-slip faults were responsible for disrupting the post-collisional geometry of the Middle Eastern Cimmerides. This disruption was largely responsible for the obliteration of the Podataksasi zone.

1. INTRODUCTION

The IGCP Project No. 276 is concerned with the "Palaeozoic geodynamic domains and their alpidic evolution in the Tethys. "To avoid any misunderstanding, I underline here that by alpidic what is meant here is the entire post-Variscan or post-Hercynian tectonics in its obsolete Bertrandian and Stillean sense. I have elsewhere emphasized (SENGOR 1984, 1985, 1986) that in this sense the term alpide has long outlived its usefulness, because it breaks down in regions where there had been no Variscan or Hercynian orogeny as in the Asiatic part of the Alpine-Himalayan system. Moreover, the destruction by subduction of Palaeo-Tethys (STÖCKLIN, 1974a, SENGOR, 1979, 1984) already commenced in the late Palaeozoic, and in such places as the Kuen-Lun even much earlier. Late Palaeozoic orogeny related to the subduction of Palaeo-Tethys had nothing to do with the coeval Hercynian events in Europe, North, Central and northern South America, and Northwest Africa and therefore was not Hercynian.
In order to denominate the non-Hercynian late Palaeozoic and early Mesozoic events in the Tethys sensu lato (cf. SENGOR, 1984) I used the term Cimmeride. I also emphasized that the term Cimmeride in my usage had no temporal connotation whatever, but was purely a genetic/geographic designation. It circumscribed the orogenic events that resulted from the destruction of Palaeo-Tethys. I contrasted it with the equally non-temporal term "Alpide" embracing only those orogenic phenomena related to the demise of Neo-Tethys. The totality of the orogenic development of the Tethys I called the Tethyside evolution, and its products the Tethysides. The following simple schema further elucidates this nomenclature:

	Ocean	Orogenic zone	
Tethys sensu lato «	Palaeo-Tethys	Cimmerides	» Tethysides
	Neo-Tethys	Alpides	

In the light of this discussion [1] it is clear that the IGCP Project No. 276 is actually concerned with the palaeozoic domains and their Tethyside and not only Alpide evolution.
The purpose of this contribution is to synthesize the late Palaeozoic orogeny events in the Middle Eastern Tethysides (Figs. 1 and 2) in terms of a new evolutionary model with a view to inspiring new field observations by pointing out particular problems and some promising avenues of approach to their solution. To this end I here review the entire late Palaeozoic (mostly Carboniferous and Permian) and Mesozoic tectonic evolution of the Middle East.
The problem that led me to this study was the significant structural, metamorphic, and igneous activity, producing recumbent folds and axial plane schistosity, Barrovian metamorphism, and calc-alkaline magmatism of early Carboniferous to in places middle Permian age in diverse tectonic units in the Middle Eastern Tethysides south of the Palaeo-Tethys suture. Their Iranian representatives in the Sanandaj-Sirjan Zone were originally interpreted by DEWEY et al., (1973) as related to extension, and in the light of the recent developments about the Cordilleran metamorphic core complexes I was originally tempted to regard the late Palaeozoic to late Triassic tectonic evolution of the Sanandaj-Sirjan zone in terms of a number of aligned metamorphic core complexes that had accompanied the opening of Neo-Tethys here. However, this interpretation seemed unsatisfactory for the following reasons:

a. Carboniferous granite emplacement in southeastern Turkey (Bitlis Massif), along the former northwestern continuation of the Sanandaj-Sirjan zone was not accompanied by extension.

b. Deformed and metamorphosed regions in the Sanandaj-Sirjan zone remained high, above sea-level, until the Jurassic. There was little coeval volcanism to suspect massive basalt underplating to generate thick crust during extension that would have led to long-lived high rift shoulders such as along the western Ghats (e.g. OLLIER & POWAR, 1985).

c. Deformation, metamorphism and igneous activity was long-lived (Carboniferous to late Triassic, i.e. about 140 Ma) in a narrow zone (about 100 Km.- wide, whose pre-Zagros collision width was probably not more than 200 Km.), in contrast to the much shorter-lived (ab. 45 Ma) Cordilleran metamorphic core complexes strewn in a wider belt (ab. 400 Km. width).

d. Major externide-type compressional deformation occured coevally in such regions as southwestern Turkey (Prof. YÜCEL YILMAZ, Pers. Comm., 1990) and Oman, then immediately adjoining the zone of deformation, metamorphism, and calc-alkaline magmatism in the Sanandaj-Sirjan Zone.

My solution offered as an alternative to what might be called the "extensional model" forms the subject of this paper.

In the Lausanne meeting Gerard Stampfli, Aymon Baud and Jean Marcoux disputed the presence of late Palaeozoic compressional deformation in Oman during the discussion of this paper. They argued that no difference in deformation between the pre-Permian rocks and those that are younger could be seen. They quoted the BRGM mapping results both in Jebel Akhdar (RABU, 1988) and in the Saih Hatat (LE METOUR, 1988) in support of their view. More recently ROBERTSON & SEARLE (1990, p. 15) also quoted RABU & al., (1990) as questioning "the reality of any "Hercynian" metamorphism and deformation" in the pre-Saiq basement of Oman. It is unfortunate that the BRGM workers are entirely misquoted both by MARCOUX, BAUD and STAMPFLI and by ROBERTSON & SEARLE (1990, p.15). First of all, RABU & al., (1990, P.50) clearly spell out the "pre-Permian (Hercynian?) deformation has had little effect, creating only open folds, lacking associated cleavage." Secondly both RABU (1988, pp.419-428) and LE METOUR (1988, pp.311-314) recognise and discuss in detail folding of late Palaeozoic age (which they regrettably call 'Hercynian'). What these two authors dispute (especially in the structurally more complicated Saih Hatat window by LE METOUR,1988) is the late Palaeozoic age of the schistosity and metamorphism studied by MICHARD (1982, 1983). Although I do not emphasize Oman in the following pages, I wish to point out here that recently MANN and HANNA (1990) reported NE-striking thrust faults and folds clearly truncated by a sub-Saiq) (Middle Permian) unconformity (see esp. their Fig. 4), thus further corroborating the results of MICHARD (1982, 1983). I take this to be strong evidence for pre-Saiq and post-Miaidin (Middle Ordovician) compressional deformation. The point made by STAMPFLI, BAUD, and MARCOUX that no difference in deformation between the pre- and post-Permian rocks is discernible - a point not previously made by any of the mapping geologists who worked on the pre-Permian rocks [2] - may result from the later reactivation of the pre-Permian structures involving major flattening as emphasized by MANN & HANNA (1990).
My own impression, gathered during a field excussion on 19th January, 1990 in the company of Dr. Samid Hanna, Dr. Peigi Wallace and Professor Robert Shackleton in the Jebel Akhdar area, was that the well-developed slaty cleavage seen in the Pre-Permian (especially in Mistal) units was conspicunously absent in the Saiq and later units. I thus differ from the BRGM workers only in considering the late Palaeozoic folding to have also produced a slaty cleavage (cf. MICHARD, 1982, 1983; MANN and HANNA, 1990).

2. SYNOPSIS OF THE LATE PALAEOZOIC AND MESOZOIC GEOLOGY OF THE MIDDLE EASTERN TETHYSIDES

In the Middle Eastern Tethysides (Fig. 1 and 2), the Palaeo-Tethyan suture in Turkey, S USSR, and N Iran form a fundamental line of demarcation separating 'Laurasian' tectonic elements in the N from 'Gondwanian' elements in the S (Fig.1).
In the Middle Eastern Tethysides two kinds of terrains are distinguished S of the main Palaeo-Tethyan suture: One kind shows significant late Palaozoic, dominantly Carboniferous, orogenic activity accompanied locally by calc-alkaline magmatism. Outcrops of this kind are grouped under a Podataksasi Zone (SENGOR, 1987a, 1990; SENGOR & al., 1988). The second kind has an Infracambrian to Middle Triassic platform cover with no significant late Palaeozoic orogeny, nor any appreciable magmatism. The two kinds of terrains are separated from each other mostly by steep faults commonly with evidence of Mesozoic and younger strike-slip displacement. In NW Iran, there seems to exist a somewhat intimate intermixing of these two kinds of terrains. In the Mesozoic these two kinds of terrains maintained their independence until the Jurassic. After the early Jurassic they became incorporated into newly-developed independent tectonic units. In the following paragraphs I review briefly the evidence for late Palaeozoic orogeny in the Middle Eastern Tethysides.

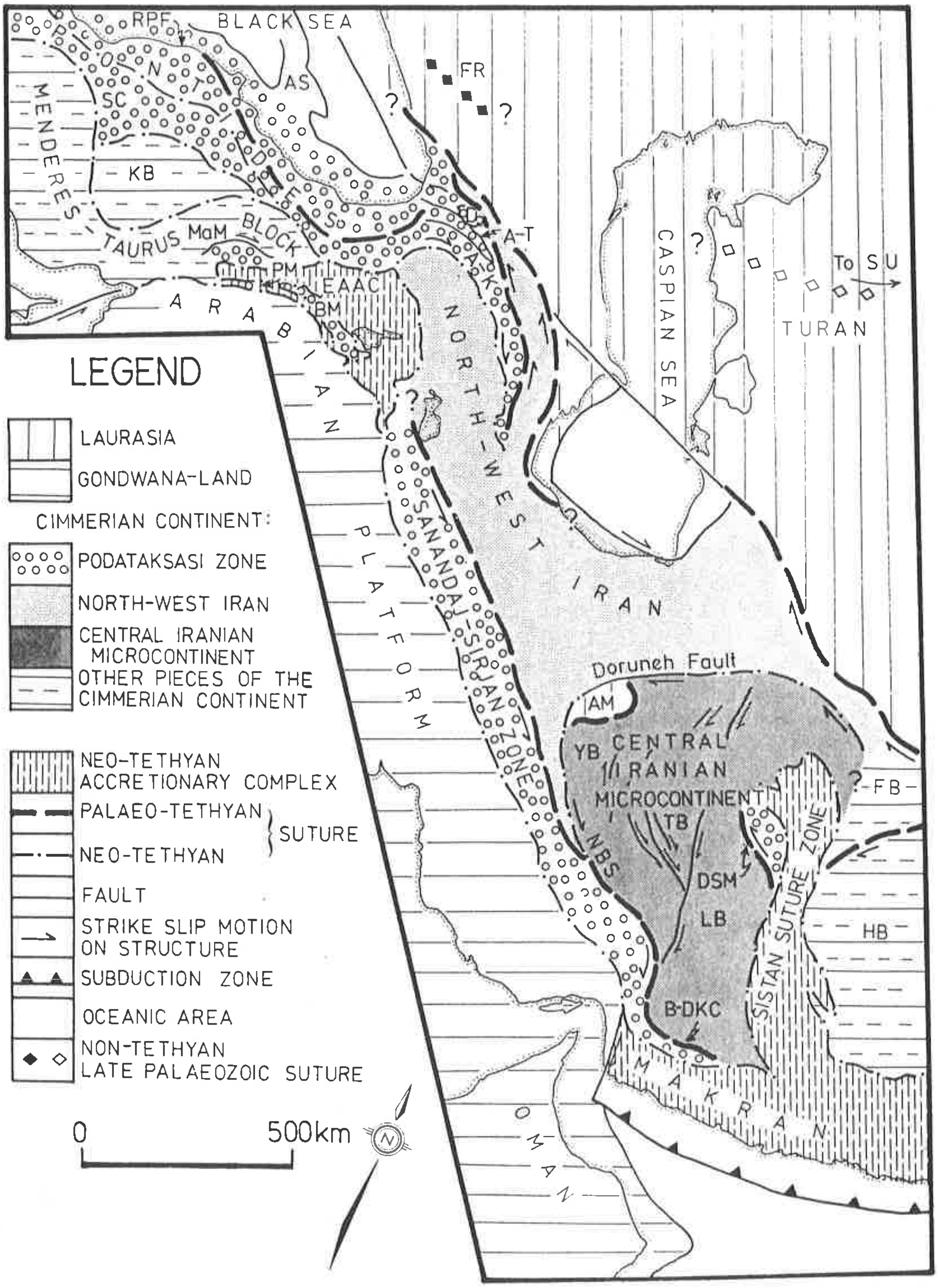


FIG. 1. Suture and blocks in the Middle Eastern Tethysides. In the Rhodope-Pontide fragment, the Palaeo-Tethyan suture is under the Podataksasi zone, as it was overthrust during the closure of Palaeo-Tethys. Key to abbreviations: A-K - Artv in-Karabagh zone, AM - Anarek Massif, AS - Andrusow Swell, A-T - Adzharia-Trialeti zone, B-DKC - Bajgan-Dur Kan Complex (including the Azava and Deyader complexes), BM - Bitlis Massif, D - Dzirula Massif, DSM - Deh Selm Metamorphics, EAAC - East Anatolian Accretionary Complex, FB - Farah Block, FR - Front Range, HB - Helmand Block (sensu SENGOR, 1984), KB - Kirsehir Block, LB - Lut Block, Mam - Malatya Metamorphics, NBS - Nain-Baft Suture, PM - Pötürge Massif, RPF - Rhodope-Pontide Fragment, SC - Sakarya Continent, SU - Sultan Uizdag, TB - Tabas Block, YB - Yazd Block.

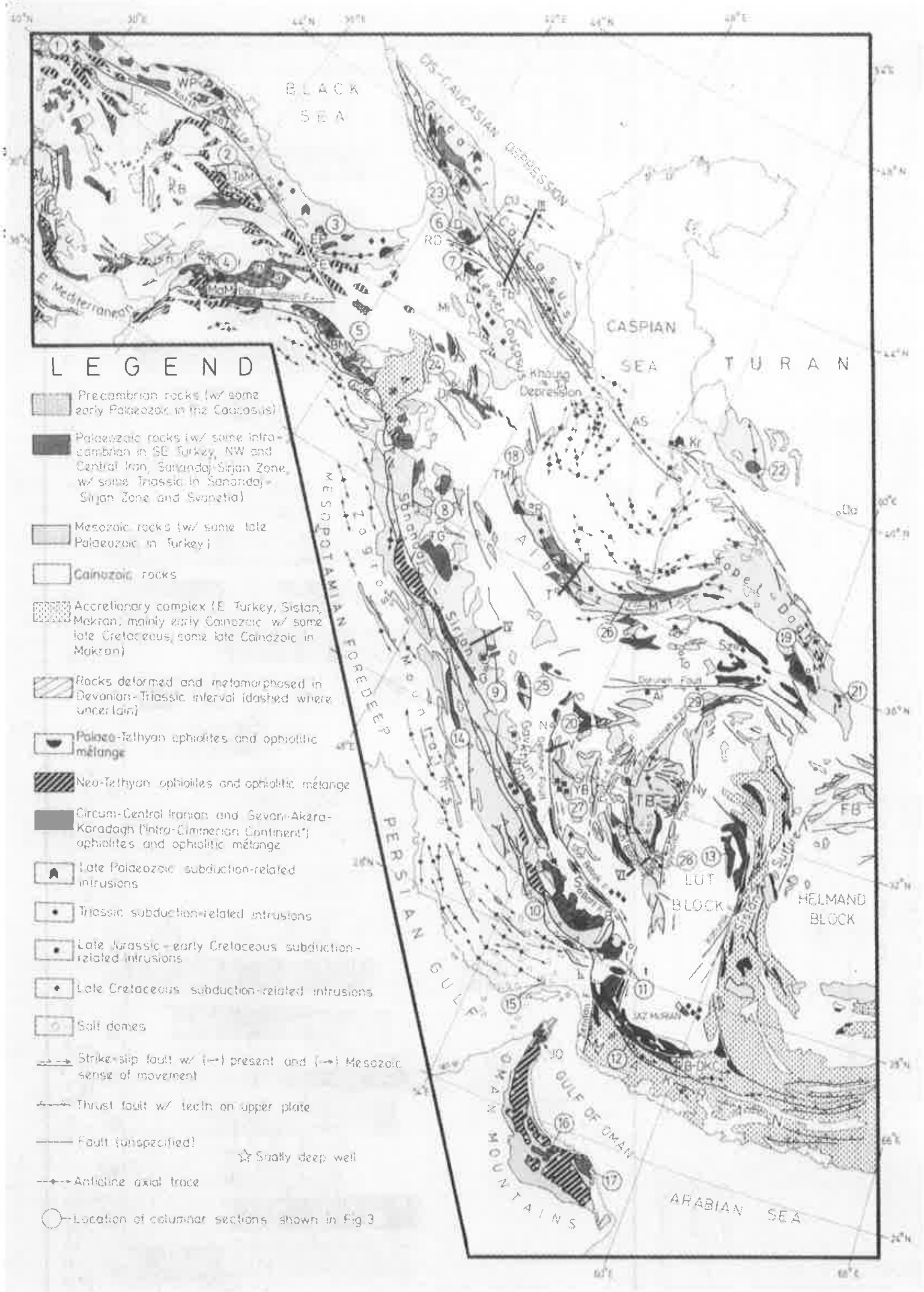


FIG. 2. Simplified geological/tectonic map of the Middle Eastern Tethysides after SENGOR (1990). Key to abbreviations: A - Ankara, Ai - Airakan, AS - Apsheron Sill, B - Baft, B-DKC - Bejgan-Dur Kan Complex, BM - Bitlis Massif, CFU - Chorchana-Utslevi Zone (see Fig. 4), D - Dzirula Massif, Da - Darvaz, Dj - Djulfa, EP - Eastern Pontides, FB - Farah Block, G - Golpaygan, I - Isfandageh, JQ - Jebel Qamar, KB - Kirsehir Block, Kh - Khrami Massif, Kr - Krasnovodsk, L - Loki Massif, M - Mashhad, MaM - Malatya Metamorphics, Mi - Mishkana Massif, N - Nain, Ny - Nayband, R - Rasht, S - Sirjan, SC - Sakarya Continent, Sh - Shirkuh, Sz - Sabzevar, T - Tehran, TB - Tabas Block, Tb - Tbilisi, TG - Tuzlu Gol, TM - Talesh Mountains, To - Torud, ToM - Tokat Massif, WP - Western Pontides, PB - Yazd Block.

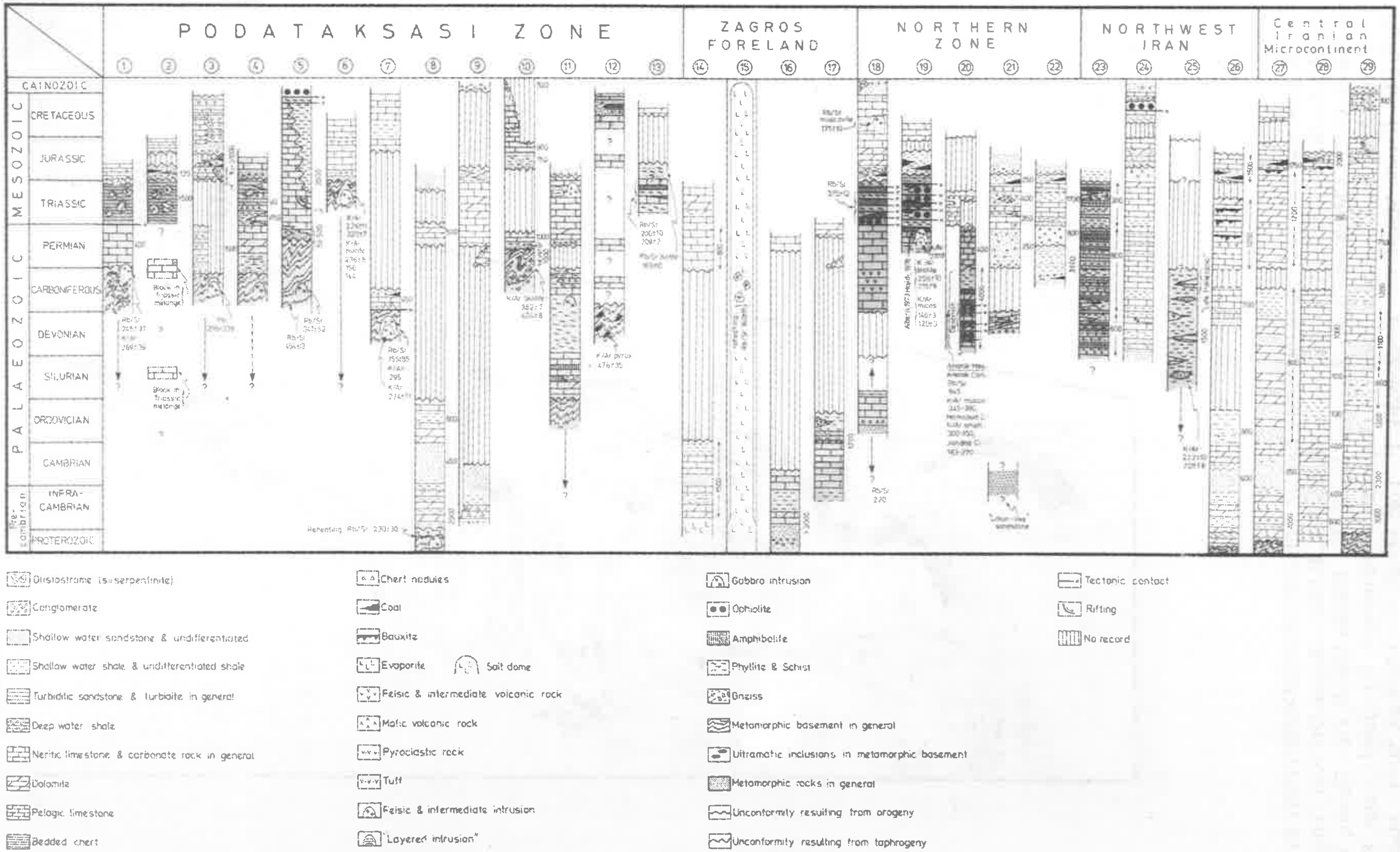

PODATAKSASI ZONE
ZAGROS FORELAND
NORTHERN ZONE
NORTHWEST IRAN
Central Iranian Microcontinent
CAINOZOIC
MESOZOIC
CRETACEOUS
JURASSIC
TRIASSIC
PALAEOZOIC
PERMIAN
CARBONIFEROUS
DEVONIAN
SILURIAN
ORDOVICIAN
CAMBRIAN
Pre-cambrian
INFRA-CAMBRIAN
PROTEROZOIC
Olistostrome (s=serpentinite)
Conglomerate
Shallow water sandstone & undifferentiated
Shallow water shale & undifferentiated shale
Turbiditic sandstone & turbidite in general
Deep water shale
Neritic limestone & carbonate rock in general
Dolomite
Pelagic limestone
Bedded chert
Chert nodules
Coal
Bauxite
Evaporite
Salt dome
Felsic & intermediate volcanic rock
Mafic volcanic rock
Pyroclastic rock
Tuff
Felsic & intermediate intrusion
"Layered intrusion"
Gabbro intrusion
Ophiolite
Amphibolite
Phyllite & Schist
Gneiss
Metamorphic basement in general
Ultramafic inclusions in metamorphic basement
Metamorphic rocks in general
Unconformity resulting from orogeny
Unconformity resulting from taphrogeny
Tectonic contact
Rifting
No record

FIG. 3. Simplified stratigraphic columns exhibiting the stratigraphic intervals from selected localities in the Middle Eastern Tethysides which are relevant to this study. Thicknesses of most of the units are shown by figures to the right of each column. With isotopic ages both the method and the mineral used are indicated. Where no mineral is mentioned, the age is one of whole rock. For the locations of the columns see Fig. 2. References for individual columns are as follows:

1- Chaput (1936), Ketin (1947, 1985), Öztunali (1973).
2- Alp (1972), Tekeli (1981).
3- Sengör and Yilmaz (1981).
4- Yilmaz et al., (1987).
5- Sengör and Yilmaz (1981).
6- Adamia and Belov (1984).
7- Abesadze et al., (1982), Adamia and Belov (1984).
8- Stöcklin (1968), Crawford (1977).
9- Thiele et al., (1968), Thiele (1973).
10- Ricou (1976), Berberian and King (1981).
11- Berberian and Berberian (1981), Davoudzadeh et al., (1986).
12- McCall and Kidd (1982), McCall (1985), Davoudzadeh et al., (1986).
13- Stöcklin et al., (1972), Crawford (1977).
14- Stöcklin (1968).
15- Crawford (1977).
16- Michard (1982, 1983), Mann and Hanna (1990).
17- Michard (1982, 1983), Mann and Hanna (1990).
18- Davies et al., (1972), Clark et al., (1975), Crawford (1977), Davoudzadeh (1986).
19- Alberti (1973), Majidi (1978), Alavi (1979), Lammerer et al., (1984), Davoudzadeh (1986).
20- Roman'ko and Morozov (1983), Davoudzadeh et al., (1986).
21- Ruttner (1984).
22- Slavin and Khain (1980).
23- Adamia et al., (1982), Adamia and Belov (1984).
24- Bonnet and Bonnet (1947), Kozur et al., (1975), Knipper and Sokolov (1976), Altiner et al., (1979).
25- Roman'ko and Sharkovskiy (1984).
26- Stöcklin (1968), Bradner et al., (1981).
27- Stöcklin (1968).
28- Stöcklin (1968).
29- Ruttner et al., (1968).

2.1 *Laurasia*

In the late Palaeozoic, the Laurasian parts of the Middle Eastern Tethysides were locus of intense orogeny related to the demise of two or more oceans. Remnants of the oldest one are seen in the front Range ophiolites in the Greater Caucasus (KHAIN, 1984) and in the Sultan Uizdag ophiolite complex in the Aral Sea region (GARETSKIY & al., 1972) whereas those of the younger one, the Palaeo-Tethys, are represented by the ophiolites and the accretionary complexes in N Turkey (SENGOR et al., 1980, 1984; YILMAZ and SENGOR, 1985), in the Caucasus (BELOV, 1989) and in N Iran, likely in the Talesh Mountains and near Mashhad (SENGOR, 1984) (Fig. 1).
Along the S accretionary margin of Laurasia in the Middle Eastern Tethysides N-dipping subduction did not end at the end of Palaeozoic. In both the Tuarkyr (Fig. 3, loc. 22) and in Aghdarband (Fig. 3, loc.21) one sees important compressional deformation in the latest Triassic to early Jurassic, which, in Aghdarband, was preceded by considerable Triassic arc magmatism in the Krasnovodsk and in Darvaz (Da. in Fig. 2), being either parts of a once-continuous magmatic arc above a N-dipping Palaeo-Tethyan subduction zone (SENGOR & al., 1988) or remnants of more than one arc in the accretionary basement of the Turkmenian SSR. BAUD & STAMPFLI (1989) argued that the marginal magmatic arc may have rifted away from the Laurasian margin in the late Scythian producing behind it a back-arc basin. From the evidence they present (breakup of an early Triassic carbonate platform, erosion of some fault blocks while others were depressed to receive a total of less than a kilometre thick turbidites and

deep-sea carbonates intercalated with volcaniclastics and tuffs), I have been unable to follow their view of a fully-grown back-arc basin (see their fig. 4), although some sort of an intra-arc (possibly strike-slip controlled as behind present Sumatra: CAMERON & al., 1980) extensional event seems clear from the record. W of the Caspian Sea in the Greater Caucasus, one does not see this early Mesozoic magmatic activity.

A number of outcrops in N Iran following the Palaeo-Tethyan suture (Figs. 2 and 3, locs. 18 and 19) exhibit strongly deformed and metamorphosed Palaeozoic sedimentary rocks that are structurally intercalated with ultramafics. The best-known of these is near Mashhad (Figs. 2 and 3, loc.19), where ALAVI (1979) mapped metamorphosed and partly serpentinized peridotites, dunites, gabbros, and thinly-bedded chert tectonically intercalated with low-grade metamorphic rocks including slates, marbles and quartzites. MAJIDI (1978) [3] also recognised here pillowed mafic volcanic rocks associated with hyaloclastics and radiolarian cherts, and a small outcrop of true ophiolitic mélange including blocks of ultramafics, schists, and marbles, floating in a matrix consisting of spilites, tuffs, pillow lavas, and 'green rocks'. The oldest unconformable rocks on this melange are *Orbitolina*-bearing Cretaceous limestones (MAJIDI, 1978, Fig. 25). The only fossils found in the metamorphic rocks are some plant remains, which A.F. de LAPPARENT thought may be related to Carboniferous forms (pers. comm. in MAJIDI, 1978, p. 55). Although DAVOUDZADEH & al., (1986) inferred that the entire assemblage was older than Permian, no rocks older than the Rhaeto-Liassic Shemshak Formation are seen to cover it anywhere (Fig. 3, loc. 19). Isotopic dating of a granite porphyry cutting the metamorphic assemblage has been so far inconclusive. MAJIDI (1978) obtained K/Ar ages of both late Permian and late Triassic, whereas Alberti (1973) found latest Jurassic to earliest Cretaceous ages. An applitic granite including the older plutons in the area has yielded a K/Ar age of 211 $\pm$ 8 Ma. (MAJIDI, 1978), thus supporting the older ages. Therefore, although DAVOUDZADEH & al. (1986) claimed that the metamorphic complex of Mashhad, including the ophiolites, was 'late Palaeozoic', available observations may be interpreted also in terms of a possible Carboniferous to Rhaetian bracket, as argued also by ALAVI (1979).

Recent observations by EFTEKHAR-NEZHAD & BEHROOZI (in press) near the town of Fariman south Mashhad showed the presence of Asselian to Murghabian neritic microfossils in limestones intercalated with the ophiolitic mafic lavas. Report of strong tectonization of the entire sequence here also suggests the presence of an ophiolitic mélange with Permian components.

The Shanderman-Asalem Complex of CLARK & al., (1975), represents a similar metamorphic assemblage. It includes actinolite-garnet-zoisite-muscovite schists and fine-grained gneisses with subordinate quartz and albite, and sheared, almost wholly serpentinized ultramafic rocks. It is located in the Talesh Mountains near the town of Rasht (Fig. 2 and 3, loc. 18). The age of these metamorphic rocks has been originally thought to be Precambrian (DAVIS & al., 1972; CLARK & al., 1975), but the oldest unconformable sequence covering them is the (here) Liassic Shemshak Formation. Rb/Sr dating of the schists yielded 375+or-12 Ma., and of the gneisses 382+or-47 Ma, suggesting middle and ?late Devonian metamorphism (CRAWFORD, 1977).

The serpentinities in places give the impression that were emplaced in the early Jurassic sequence, with which they are intimately sheared and brecciated (DAVIES, & al., 1972). This led both BERBERIAN & KING (1981) and SENGOR (1984) to interpret the Talesh ultramafics as remnants of Palaeo-Tethys, likely emplaced in late Triassic time, and thus unrelated to the older Shanderman-Asalem metamorphic complex. DAVOUDZADEH & al., (1986) retained the interpretation of DAVIES & al., (1972) and viewed the ultramafics as parts of the metamorphic complex that they interpret as Palaeozoic.

The lithology, structure, and the age of the Mashhad and Talesh outcrops, together with the Anarek Massif (see below), with which they can be compared best, suggest that they may be Palaeozoic accretionary complexes whose evolution in places (e.g. Mashhad and possibly Anarek) may have reached into the Triassic. They are similar to the accretionary complexes of the Paropamisus, W Hindu Kush, N Pamirs (SENGOR, 1984; BOULIN, 1988) and those of N Turkey (the Küre Nappe: SENGOR, 1984, YILMAZ & SENGOR, 1985) but the Turkish Cimmeride accretionary complexes reach further up into the earliest middle Jurassic (SENGOR & al., 1980, 1984). The Iranian occurences form a part of the same orogenic collage marking the suture of Palaeo-Tethys (SENGOR, 1984; DAVOUDZADEH & SCHMIDT, 1984).

2.2 *Cimmerian Continent and Gondwana-Land*

The N margin of Gondwana-Land to the S of Palaeo-Tethys, i.e. the Cimmerian Continent of the Triassic (SENGOR, 1979), has been regarded as an Atlantic-type, N- to NE-facing continental margin (ARGYRIADIS, 1974, 1978; STÖCKLIN, 1974a, 1980; JENNY, 1977; STAMPFLI, 1978; BERBERIAN & KING, 1981; TAPPONNIER & al., 1981; BOULIN, 1988) with the exception of N Turkey (SENGOR & al., 1980, 1982, 1984; BERGOUGNAN & FOURQUIN, 1982). In Iran, most writers since GANSSER (1955) have underlined the absence of late Palaeozoic orogeny S of the Kopet Dagh/ Aghdarband ranges (e.g. ASSEROTO, 1963; FLÜGEL, 1964; HARRISON, 1968, pp. 182-183; STÖCKLIN, 1960, 1968, 1974b, 1977; STÖCKLIN & al., 1964, 1965; see also the numerous doctoral dissertations completed on northern and central Iran in the 1960's and 1970's which came out of the Swiss Federal Institute of Technology (Zürich) and the University of Geneva) in contrast to such earlier workers as STAHL (1911) and FURON (1941) and insisted that since the early Palaeozoic the first orogeny was of early Jurassic age corresponding with the collisional obliteration of Palaeo-Tethys (cf. SENGOR, 1984).

Recently a number of writers have indicated that the absence of important late Palaeozoic and early Triassic orogeny in Iran is not supported by accumulating new field data (THIELE & al., 1968; THIELE, 1973; BERBERIAN & BERBERIAN, 1981; BERBERIAN & KING, 1981; ROMAN'KO & MOROZOV, 1983; ROMAN'KO & SHARKOVSKIY, 1984; DAVOUDZADEH & al., 1986; DAVOUDZADEH & WEBER-DIEFENBACH, 1987; SENGOR, 1987a, 1990; SENGOR & al.,1988). In the following pages I summarize these data from Iran and its vicinity. The descriptions follow the major tectonostratigraphic entities employed here, viz. the Podataksasi Zone, the Central Iranian Microcontinent, and NW Iran (Fig. 1) which together form parts of the Cimmerian Continent in the Middle East.

2.1.1 Podataksasi Zone

Region originally belonging to Gondwana-Land in the Middle Eastern Tethysides in which there is record of late Palaeozoic orogeny (Fig. 2, loc.1 through 13) have been designated under the name Podataksasi Zone derived from the names of the Alpide tectonic units containing them, viz, the Pontides, Dzirula, Adzharia-Trialeti, Artvin-Karabagh, and Sanandaj-Sirjan[4] (Fig. 1).

In the Isfandageh region (Fig. 2 and 3, loc. 11) early Permian turbidites and middle and late Permian shallow water limestones with basal conglomerate sit across an unconformity on 5000 m. of intensely deformed Devonian and Carboniferous schists, phyllites, calc-schists, metadiabases, and metagabbros constituting the Khaju unit (DAVOUDZADEH & al., 1986). The Permo-Triassic carbonates in this region were themselves deformed and metamorphosed, and intruded by the Triassic Sikhoran layered intrusion that is covered by Jurassic sedimentary rocks as are the Permo-Triassic metamorphic rocks (SABZEHEI, 1974). DAVOUDZADEH & al., (1986) and DAVOUDZADEH & WEBER-DIEFENBACH (1987, p.139) concluded, that at least two episodes of orogeny are seen in the Isfandageh area: the first was late Carboniferous, and the second was late Triassic.

Another region, in which Paleozoic and Triassic deformed and metamorphosed rocks including carbonates, pelitic and psammatic rocks, and mafic volcanics plus ultramafics are encountered lies to the W of Sirjan, near Kor-e Sefid (Figs. 2 and 3, loc.10: BERBERIAN & KING (1981); DAVOUDZADEH & WEBER-DIEFENBACH (1987)). Isotopic data (Fig. 3, loc.10) yield early Devonian and early Carboniferous cooling ages. The early Devonian age likely stems from an earlier Palaeozoic basement. SABZEHEI & BERBERIAN (1972), VIALON & al., (1972) and RICOU (1976), have recognised two syntectonic episodes of regional metamorphism of lower amphibolite

grade could be older and identical with the early Carboniferous event and the second late Triassic and retrograde.
Farther N along the Sanandaj-Sirjan zone, near Golpaygan and Mahallat (Figs. 2 and 3, loc 9) THIELE & al., (1968) and THIELE (1973) have shown the presence of a significant middle to possibly late Permian folding phase that created ENE to NE trending close folds. The folded rocks include a late Proterozoic to Palaeozoic succession containing from bottom to top, the correlatives of the Kahar (Precambrian), Soltaniyeh (Infracambrian), and Zaigun-Lalun (infra-Cambrian to Lower Cambrian) formations and a 100 m.-thick massive dolomite. THIELE (1973, p.494) discovered in this dolomite *Pseudohuangia sp.* and *Ipsiphyllum sp.* indicating an Asselian to Murgabian (Lower to Middle Permian) age. He noted in one place a sharp unconformity between this dolomite and the late Proterozoic to Ordovician succession, a condition very similar to the situation in Oman, with the exception that here the unconformable dolomite begins with the Lower Permian *(Pseudohuangia)*. In this region the main folding postdates the Lower to Middle Permian dolomite, but is older than the unconformable calcareous shales, sandy limestones, sandstones, and conglomerates containing the calcareous algae *Permocalculus cf. plumosus* ELLIOT and *Pseudovermiporella*, suggesting an Upper Permian age (THIELE, 1973). The main folding in the central part of the Sanandaj-Sirjan zone is therefore middle Permian age, with probably a forerunner, whose stratigraphic bracket is Middle Ordovician to Lowest Permian. In the western Golpaygan area, DAVOUDZADEH & WEBER-DIEFENBACH (1987) noted volcanistics, flysch-type shales, and sandstones with *Pseudoschwagerina*-bearing (i.e. Lower Permian) limestone interlayers, and metadia-base and andesite flows indicating Permian calc-alkaline magmatism and clastic sedimentation.

Regions of intense late Palaeozoic orogeny S of the main Palaeo-Tethyan suture are present both farther N and NW and S and SE of the Sanandaj-Sirjan zone. To the S, they are located in the Bajgan Complex (McCALL & KIDD, 1982, Figs. 2 and 3, loc. 12), made up of pelitic and psammitic schists, calc-silicates, limestones, and marbles. One of the ultramafic bodies yielded a K/Ar age on pyroxene of earlier middle Ordovician, interpreted as time of crystallization [3] (Fig. 3, loc.12). McCALL (1985) found many inliers of unmetamorphosed, locally slightly recrystallized Permian limestones containing *Pseudoschwagerina sp., Schwagerina sp., Schubertella sp.*, and *Climacammina sp.* He thus considers the metamorphic rocks at least as pre-Permian. Continuity of outcrop with insignificant interruptions between the Bajgan Complex and the metamorphic rocks of the Isfandageh area (Fig. 2, locks. 11 and 12), the similarity of the sequences and metamorphism in the two regions, and the existence of unmetamorphosed Permian rocks in the Bajgan area, led McCALL & KIDD (1982), DAVOUDZADEH & al., (1986), and DAVOUDZADEH & WEBER-DIEFENBACH (1987) to view the Bajgan metamorphics and their oriental correlatives in the Deyader and the Azava complexes (shown as parts of the Bajgan - Dur-kan complex in Fig. 2) as equivalents of the late Palaeozoic metamorphic rocks in the Isfandageh area. They thus assigned them a late Carboniferous age.

To the N and NW, it is not possible to follow the Sanandaj-Sirjan zone continuously beyond S of Lake Urmieh (about the position of loc. 8 in Fig. 2). Regions exhibiting similar late Palaeozoic evolution are separated from it by younger ophiolitic sutures (e.g. the Sevan-Akera-Karadagh ophiolites in the Lesser Caucasus and the Neo-Tethyan accretionary complex in eastern Turkey, EAAC: Fig. 1) and by large areas in which no trace exists of any Palaeozoic orogeny following the Pan-African tectonism (e.g. locs. 23 and 24 in Figs. 2 and 3). Evidence for notable late Palaeozoic Gondwanian orogeny N of the northern termination of the Sanandaj-Sirjan zone is seen in the Lesser Caucasus and in the Dzirula Massif in Transcaucasia.
In the Dzirula Massif (Figs. 2 and 3, loc. 6 and Fig. 4) two different rock associations exist whose mutual contacts are probably large offset strike-slip faults. The 'crystalline basement' outside the Chorchana-Utslevi Zone (Fig. 5) (ABESADZE & al.,1982) is made up of high grade gneisses, amphibolites, and migmatites predating cross-cutting 'early Variscan' quartz diorites and 'late Variscan' microcline granites. K/Ar cooling

ages on these metamorphic rocks are also late Carboniferous (ADAMIA & al., 1983). K/Ar dates suggest that the latest possibble time of granite generation was mainly late Carboniferous to possibly earliest Triassic (Fig. 3, loc. 6). Small syenite plutons ('Rikotites'), yielded 236+8, 156, and 144 Ma K/Ar ages on biotite, suggesting middle Mesozic disturbance.

The highly sheared and tectonically interleaved rocks of the Chorchana-Utslevi zone contain, in addition to gabbros of uncertain age, serpentinites, and metavolcanics suggesting a dismembered ophiolite, Lower Cambrian marble slivers (*Archaeocyathus sp., Coscinocyathus caucasicus* VOL.) and Upper Silurian to Upper Devonian phyllites containing a rich palynomorph flora. The deformation and metamorphism of these rocks were post-Devonian. The phyllites yielded a late Carboniferous Rb/Sr age (Fig. 4) probably indicating the time of the greenschist metamorphism (ABESADZE & al., 1982), which was nearly contemporaneous with the Dzirula Massif granites. After the metamorphism of the phyllites and the intrusion of the granites, important strike-slip faulting of unknown sense and magnitude affected these rocks (Fig. 4). The time of this deformation is bracketed between the youngest granites (?late Palaeozoic ?earliest Triassic) and the unconformable Liassic shales and sandstones (ADAMIA & al.,1982), and was likely late Triassic-earliest Jurassic. The original palaeogeographic affiliation of the metasediments associated with the dismembered ophiolite is unknown, but comparison with some of the N Iranian occurences suggests that it may be Laurasian.

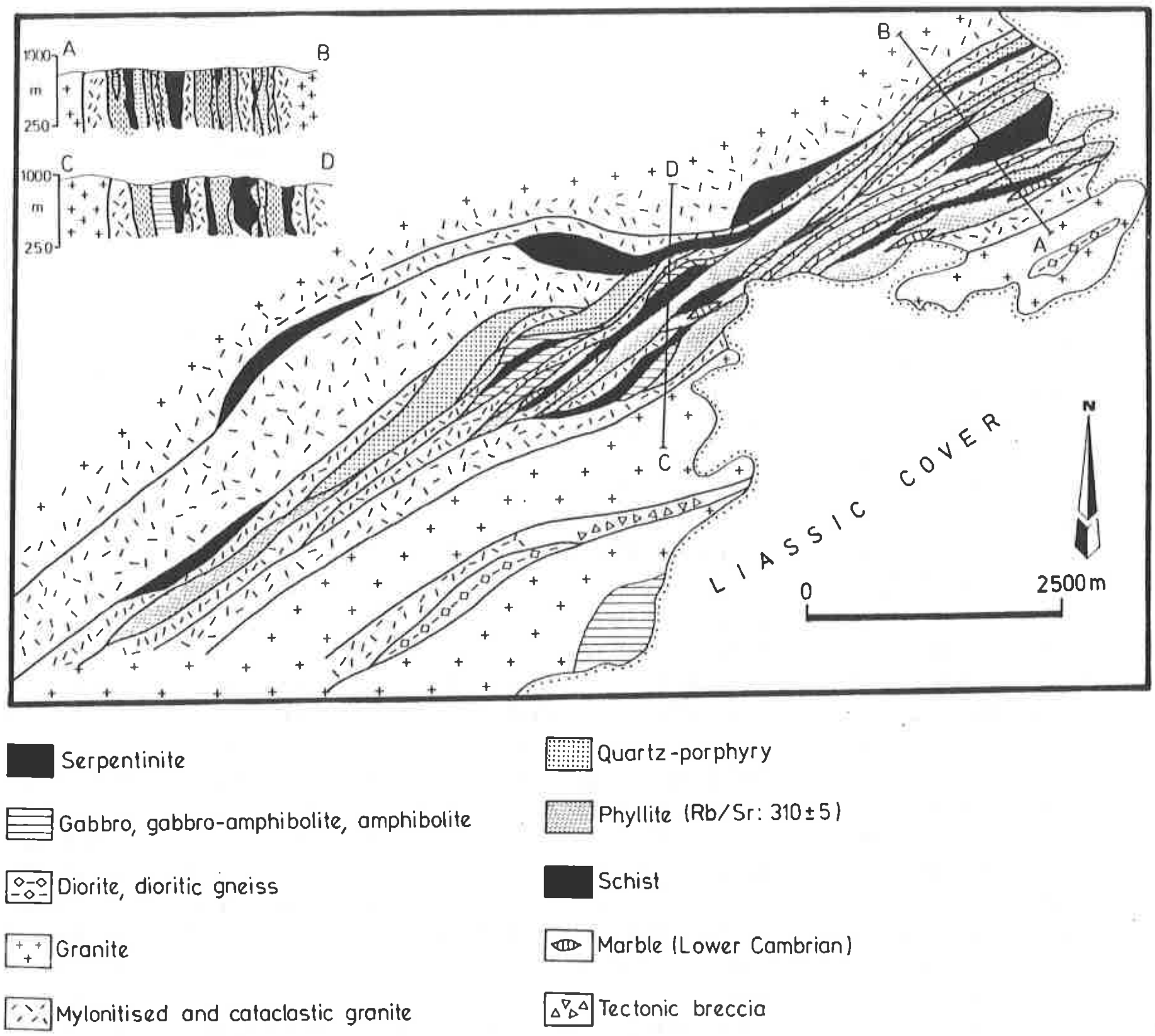


FIG. 4. Simplified geological map and cross-sections (A-B and C-D) of the Chorchana-Utslevi shear zone of the Dzirula Massif (for location see Fig.2) redrawn from ADAMIA & al., (1982).

Evidence of similar events exists farther S in the Khrami Salient (Fig. 2, loc. Kh), where an unmetamorphosed Carboniferous sedimentary and volcaniclastic succession is also preserved in fault contact with metamorphic and plutonic rocks. The metamorphics are andalusite-sillimanite grade quartzo-feldspathic gneisses with subordinate micaschists. The metamorphism is older than the intrusion of mainly S-type granites of (?middle) Carboniferous age (Fig. 3, loc.7). The sedimentary rocks, containing throughout the section numerous tuff horizons, range from late Visean to Namurian marine limestones, clastics to pyroclastics, to middle Carboniferous paralic to terrestrial clastics and pyroclastics exhibiting a Euramerian flora [6] (ADAMIA & BELOV, 1984). Petrologic and geochemical evidence suggest that the Khrami volcanic rocks are the eruptive equivalents of the granites (ADAMIA & al., 1982). In the Chochiani river gorge a small tectonic sliver containing gabbro-diorites, olivine gabbros and serpentinities crop out, which may be correlative with the disrupted ophiolites of the Chorchana-Utslevi zone in the Khrami Massif (ADAMIA & al., 1982).

In the Loki Massif (Fig. 2, loc.1) to the S of the Khrami Salient, greenschist and locally epidote-amphibolite facies micaschist and rare marbles crop out. K/Ar muscovite ages of 287+or-15, 319+or-11, and 333+or-11 Ma from them probably indicate late Carboniferous cooling (Abesadze et al., 1982). Granitic rocks intruding the Loki metamorphics have K/Ar and Rb/Sr ages around 338 to 325 Ma, corresponding with a later early Carboniferous intrusive episode, which may have been coeval with the peak of the metamorphism.
Evidence from Transcaucasia and the Lesser Caucasus thus indicates a period of metamorphism (and accompanying penetrative deformation) and both plutonic and volcanic activity ranging from later early carboniferous to late Carboniferous to possibly later, similar to what we have seen at the S extremity of the Sanandaj-Sirjan zone. Similar coeval events probably generated the Sub-Permian unconformity and the early Permian calc-alkaline volcanics in the middle part of the Sanandaj-Sirjan zone, and of the deformation and metamorphism of the Bajgan Complex.
Gondwanian late Palaeozoic orogeny is seen in diverse Alpide tectonic units in Turkey that originally may have constituted a single entity (SENGOR & al., 1980, 1984) are located WSW of the Transcaucasian and Lesser Caucasian late Palaeozoic inliers. This basement consists of a possibly early Devonian quartzo-feldspathic gneissic host rock cut by the minimum-melting composition Gümüshane granite (YILMAZ, 1975 and pers. comm., 1988) that yielded a Pb whole rock age of latest early Carboniferous to earlier late Carboniferous (Figs. 2 and 3, loc. 3). A composite section consisting, from base to top, of neritic limestones, clastic sedimentary (KETIN, 1951) contains in places Gzelian-Asselian (latest Carboniferous-earliest Permian) marine fossils (KESKIN, 1986) and in places Stephanian (latest Carboniferous) Euramerian plant fossils (WAGNER, pers. comm., 1987) and forms the cover to the pre-mid-Carboniferous basement.
To the S of locality 3 (Fig. 2) and across the East Anatolian Accretionary Complex (Fig. 1), is the Bitlis Massif (Figs. 2 and 3, loc. 5), consisting of several pre-late Cretaceous to middle Eocene back-arc basin basalts and sedimentary rocks (SENGOR & YILMAZ, 1981). Its pre-Cretaceous basement contains two major sequences: One is a pre-Permian basement consisting of amphibolite grade gneisses (felsic metavolcanics), amphibolites, metavolcanics, metatuffs, and metaagglomerates intruded by two granite intrusions of early Carboniferous age (HELVACI & GRIFFIN 1984). The whole assemblage is uncomformably covered by the next sequence of quartzites, garnet-biotite micaschists and marbles, whose ages are thought to be Permian to late Cretaceous on the basis of local correlations (SENGOR & YILMAZ 1981). HELVACI & GRIFFIN (1984) have shown that the age of eruption of the gneissic metavolcanics in the basement is Ordovician (Fig.3, loc.5). Therefore, the metamorphism of the basement at locality 5 (Fig. 3) is bracketed between late Ordovician and early Carboniferous and probably correlative with the metamorphism in basement of the Eastern Pontides (SENGOR & YILMAZ 1981). YILMAZ & al., (1987) have documented a very similar sequence in the Engizek Mountains (Figs. 2 and 3, loc.4).

Numerous authors since Kellog (1960; see also BASTUG, 1976; YAZGAN & al., HEMPTON, 1985) have maintained that the great similarity in the stratigraphic succession between Arabian platform and the Bitlis-Pötürge crystalline complexes indicate that they formed parts of a single tectono-stratigraphic entity until at least the Triassic. SENGOR & YILMAZ (1981) and YILMAZ & al., (1987) presented lithological and sequence data to show that a similar resemblance existed between the Bitlis-Pötürge fragment and the eastern Pontides plus the Sakarya Continent (OKAY's, 1989; 'Sakarya Zone'). This ties the westernmost parts of the Podataksasi Zone directly to Gondwana-Land. This connexion is supported by the recently-discovered latest Carboniferous-early Permian marine carbonate sequence in the eastern Pontides (KESKIN, 1986) that resembles more the Gondwanian fully-marine Permo-Carboniferous in Turkey than the paralic to terrestrial (and largely non-existant) Permo-Carboniferous of the southern fringe of Laurasia.
The basement of the Sakarya Continent (Fig. 1) was discovered by KETIN (1947,1985) in the Uludag Massif (Figs.2 and 3, loc. 1) where he mapped quartzo-feldspathic gneisses, amphibolites and marbles. The age of this intensely deformed association that is similar to the Bitlis Massif basement, unfortunately remains unknown. The marbles may have been originally unconformable on their gneissic and amphibolitic basement (as in Bitlis), but younger deformation has erased the original relationship. The Uludag marbles are most likely part of the Permian carbonate platform cover seen elsewhere on the Sakarya Continent (SENGOR & YILMAZ 1981). In the Uludag Massif we thus maybe looking at two orogenic events of late Palaeozoic age: the first possibly premarble (i.e.pre-Permian) and the second possibly Permo-Triassic (pre-Jurassic in any event).
The localities 1, 3-7, and 9-12, all of which are situated to the S of the main Palaeo-Tethyan suture, and some of which (4, 5, and 9) exhibit clear Gondwanian Palaeozoic stratigraphy, show that a significant episode of multiple orogenic deformation, Barrovian metamorphism, and subdution-related magmatic activity characterized large regions near the N margin of Gondwana-Land in early carboniferous through Permian time. Orogeny was especially widespread in Carboniferous time, but, in places, lasted at least into the middle Permian.

2.2.2 Arabian Platform

The obvious place to look first for the place of the Podataksasi Zone in Gondwana-Land is the Arabian Platform. An Arabia-wide pre-Permian lacuna reaches a Silurian, but locally Ordovician of Devonian lower limit in different areas along the Arabian Platform's visible parts closer to its Palaeozoic margins (e.g. Figs. 2 and 3, loc. 14). In these regions sedimentation began with Lower to Upper Permian red sandstones containing a possibly mixed Cathaysia and Gondwana-Land flora (WAGNER, 1962; ARCHANGELSKY & WAGNER, 1983; CTYROKY, 1973). ZIEGLER (1990) pointed out that floras assigned to the 'tropical or Cathaysian realm' (biomes 1,2 and 3 in ZIEGLER, 1990) include the 'Euramerian' floras (biome 2) and that palaeobotanists link all localities from western North America, Europe and Arabia together. ZIEGLER's (1990, fig. 4) Kazanian world map indeed shows the biome 2 (Euromerian flora) both in Spain and in Central Europe and in Arabia (from the Khuff Formation after El-Khayal and Wagner, 1985).
Starting with the late Permian, shallow water carbonate sedimentation with Arabian margin. At least locally, the Upper Permian sedimentary rocks (widespread limestones and dolomites, evaporites, and subordinate clastics of the Dalan Formation) were deposited in extensional basins (SZABO & KHERADPIR, 1978). Known Permian volcanism on the Arabian Palaeozoic margin is confined to some basalt flows W of Dehhid in the S part of the Zagros Mountains (BERBERIAN & KING, 1981) and to some of the internal units of the Hawasina nappes in Oman (BECHENNEC, 1988). Clearly by late Permian time extension had begun affecting the Arabian Palaeozoic margins, at least along the present-day Zagros, because evidence for Permian extension is unknown in SE Turkey.

Carboniferous record from the Arabian margin is known from Precambrian igneous rocks that came up in salt diapirs (Figs. 2 and 3, loc 15). These range from rhyolites, ignimbrites, through porphyritic felsic and mafic lavas to dolerites and dolerite gabbros. CRAWFORD (1977) obtained from four samples of similar petrography a whole rock isochron of 340+or-15 Ma (Visean), possibly corresponding with a Carboniferous event correlative with the early Carboniferous metamorphism in the Kor-e Sefid area (Fig. 2, loc. 10), and other contemporaneous orogenic events in the Podataksasi Zone, suggesting that the latter was near Arabia in the early Carboniferous at least.

2.2.3 Interior Iran

The relationship of the Podataksasi Zone with the rest of Iran S of the main Palaeo-Tethyan suture zone is not easy to establish because of widespread Tertiary and Quaternary cover, abrupt facies change, and complex deformations that took place in the Mesozoic and the Cainozoic. A zone of depressions (Lake Urmieh, Tzulu Gol, Gavkhuni, and the Jaz Murian) effectively covers the contact region between the Sanandaj-Sirjan Zone and interior Iran (Fig. 2). The very linearity of this zone (see also BERBERIAN, 1981, Fig.2) between Lake Urmieh and Isfandageh suggests that a strike-slip system may be partly responsible for it. Along its S extension, the Gavkhuni is delineated by some remarkably straight faults (Abadeh, Dehshir, Shar Babak, and Baft faults, Fig. 2), some of which exhibit Quaternary right-lateral strike-slip (BERBERIAN, 1981). Farther N along the boundary young faulting is not observed. STÖKLIN (1974b), however, inferred that steep faults separate the Sanandaj-Sirjan zone from interior Iran and affect a part of the Mesozoic and all older rocks (Fig.5, sec.IV). BERBERIAN and KING (1981) and DAVOUDZADEH & WEBER-DIEFENBACH (1987) emphasized the 'geosynclinal' facies of the Palaeozoic rocks of the Sanandaj-Sirjan zone in contrast to the platform facies of much of interior Iran.

Interior Iran, separated from the Sanandaj-Sirjan Zone by a belt of steep, straight faults reaching well into the Mesozoic, consists of two distinct tectonic provinces. The NW Iran tectonic province includes those regions S of the main Palaeo-Tethyan suture and N of the Doruneh Fault and reaches into Nakhichevan, Armenia, and Svanetia (Fig. 1). The Central Iranian Microcontinent (Takin, 1972) is the terrain delimited by the Sistan, Nain-Baft, and Makran ophiolitic belts and the Doruneh Fault plus the Sabzevar ophiolites (Fig. 1, STÖKLIN, 1974a, 1977; BERBERIAN & KING, 1981; DAVOUDZADEH & al., 1981; DAVOUDZADEH & SCHMIDT 1984).

The Central Iranian Microcontinent itself is separated into three major sub-blocks by long, E-concave, N-S, right-lateral strike-slip faults (Fig. 1). The easternmost of these blocks, the Lut (roughly equivalent to HARRISON's, 1968, 'East Iranian Quadrangle'), has been the most stable since the late Precambrian and constitutes the only genuine 'median mass' in Iran. It is demarcated from a Tabas block by the Nayband Fault (Fig. 2) along which the Jurassic 'Shotori Horst' (STÖCKLIN & al., 1965) has been a constant facies divider through time. The Tabas block is demarcated from the larger and internally inhomogeneous Yazd block by the curved faults of Kuh-e Kalsaneh and kuh-e Banan (Fig. 2). The Posht-e Badam Fault my delimit and eastern sub-block of the Yazd. The Tabas and the Yazd blocks together constitute HARRISON's (1968) 'Tabas Wedge' (see his fig.35).

These large N-S faults formed early in the history of Central Iran. They have been persistent facies dividers since the Infracambrian (STÖCKLIN, 1968; BERBERIAN & KING, 1981) and repeatedly reactivated in various roles throughout the Phanerozoic. Many of them, or at least considerable stretches along them, are now active and nucleate earthquakes (BERBERIAN, 1981).

In contrast to the Podataksasi Zone, the Central Iranian Microcontinent was a quiet platform throughout the Palaeozoic until the late Triassic-early Jurassic (DAVOUDZADEH & WEBER-DIEFENBACH 1987). In it, the late Precambrian Pan-African orogeny (960-600 Ma. in Central Iran) was succeeded by a platform regime, in which alternating shallow-marine, lagoonal, and continental deposits were deposited (e.g. Fig. 3, locs. 27, 28, 29). There are important lacunae in the Palaeozoic record in much of Central Iran, the most widespread of which being the earlier middle Devonian

('Eifelian Hiatus') and the Stephanian (late Carboniferous) gaps (STÖCKLIN, 1968, BERBERIAN & KING, 1981; WEDDIGE, 1984a, 1984b, DAVOUDZADEH & WEBER-DIEFENBACH, 1987). The total Infracambrian to Middle Triassic thickness, between 3 and 4 Km is surprisingly constant, (STÖCKLIN, 1968). There are a number of regions within the Central Iranian Microcontinent, however, whose late Palaeozoic/early Mesozoic geology deviates considerably from the rest of Central Iran. These are discussed in the following paragraphs.

The Tabas block is one region that differs somewhat from the rest of Central Iran with respect to its Palaeozoic evolution. In it RUTTNER & al., (1968) discovered probably one of the thickest and most complete Palaeozoic sections in the entire Middle East, reaching an average of more than 8 Km (Fig. 3, loc.29; see also RUTTNER, 1980, Plate III). BRATASH (1975) noted total thicknesses in places more than 10, and in one locality 14 Km. The 'Eifelian hiatus' is not present here but is represented instead by a regressive facies of clastics and evaporites. Similarly, the Upper Carboniferous, absent elsewhere in Central Iran, is present in this area. It consists of shallow water carbonates (SARDAR Fm.: see DAVOUDZADEH & WEBER-DIEFENBACH, 1987, Fig. 1).

Both the igneous and the structural history of the Tabas block is more lively throughout the Palaeozoic than in the rest of the Central Iranian Microcontinent. Mafic and intermediate volcanism characterized this region during the Palaeozoic, but not abundantly (BERBERIAN & KING 1981, BRATASH (1975) showed that the strongly subsident character of the Tabas block, that was formely thought to be confined to the Shirgesht - Ozbak-Kuh Mountains in the extreme N of the block, is now known to have characterized the entire block in the Palaeozoic and the Mesozoic until the Cretaceous. BRATASH (1975) emphasized the presence of early Triassic or possibly even earlier 'inversion' events (?compressional) in the N part of the Tabas block in the Kashmer area.

Another locality, whose Palaeozoic history deviated from the rest of Central Iran in a different way is the Saghand region (Fig. 2, wust E of the Posht-e Badam Fault; DAVOUDZADEH & WEBER-DIEFENBACH, 1987, Fig. 1, N of their loc.8). CRAWFORD (1977) obtained a reliable Rb/Sr age of 315+or-5 Ma on a single biotite from the late Proterozoic Saghand metamorphic complex, suggesting a disturbance of earlier late Carboniferous age. From whole rock and biotite analyses CRAWFORD (1977) also found evidence for a younger event at 240+or-15 Ma (late Permian to earliest Triassic). BERBERIAN & KING (1981) commented that this 'event' did not fit with any orogeny in Central Iran, but a similar age has been obtained from the Precambrian volcanics in the Takab region (Figs. 2 and 3, loc 8) suggesting a similar rejuvenation.

Two other regions stand out in the Central Iranian Microcontinent because of their very different geological evolution. Unlike the two cases just discussed however, they represent now-displaced pieces of tectonostratigraphic entities from outside Central Iran. One of these is the Deh Salm metamorphic complex (Figs. 2 and 3, loc. 13). In it STÖCKLIN & al., (1972) discovered a sequence made up of an alternation of schists and marbles with subordinate amphibolites associated with the marbles in its lower half, and phyllites and quartz micaschists in places including graphite in its upper half. The whole sequence is strongly tectonised and metamorphism increases from upper greenschist facies in the W to amphibolite facies in the E. REYRE & MOHAFEZ (1970, p. 983) reported Rb/Sr ages of 206+or-10 and 209+or-2 Ma corresponding with a possibly latest Triassic to earliest Jurassic interval. A Rb/Sr age on biotite from the eastern high grade rocks yielded a middle Jurassic age of 165+10 Ma (CRAWFORD 1977). BERBERIAN (1977) suggested that the Deh Salm metamorphics were formed during the late Triassic-early Jurassic Palaeo-Tethyan ocean closure. This view is accepted by most later workers (e.g. DAVOUDZADEH & SCHMIDT (1984) in interpreting the Deh Salm metamorphics as a displaced fragment of the Sanandaj-Sirjan Zone, i.e. a part of the larger Podataksasi Zone.

The final region with a deviant geological evolution in the Central Iranian Microcontinent which I discuss here is the Anarek region (Figs. 2 and 3, loc.20, BERBERIAN & KING, 1981; DAVOUDZADEH & al., 1981; DAVOUDZADEH & SCHMIDT, 1984), also known as the Anarek-Khvor massif (ROMAN'KO &

MOROZOV, 1983). It is made up of a sequence of limestones, schists, dolomites, amphibolites, and limestone succession (REMAN'KO & SCHMIDT, 1983, Fig. 2). The lower sequence has structurally intercalated slices of ultramafic rocks found in NNE-vergent schuppen in the Anarek region was in greenschist facies. The Anarek metamorphic complex is similar to those near Mashhad and in the Talesh Mountains and also to the metasediments and ophiolites of the Chorchana-Utslevi zone in the Dzirula Massif.

Its age is thought to be pre-middle Triassic, because DAVOUDZADEH & SEYED-EMAMI (1972) reported clasts of the metamorphic rocks from the Baqoroq conglomerate of late Anisian-early Ladinian age belonging to the nearby Nakhlak Group (see below). Isotopic ages of the various parts of the metamorphic complex form a wide spectrum ranging from late Precambrian to early Jurassic (Fig. 3, loc. 20), which is not precise enough to enable one to assign the Anarek massif unambiguously to any one of the tectonic units in Iran and surrounding regions.

By contrast, the younger, and likely overlying, Naklak Group of sedimentary rocks ranging from Lower to Middle Triassic greatly resembles the Triassic sequence known from Aghdarband (Fig.2, loc. 19) and suggests that the Nakhlak Group is its correlative belonging most probably to the same DAVOUDZADEH & SCHMIDT, 1984; RUTTNER, 1984).

The palaeozoic evolution of what SENGOR (1990) called NW Iran (Fig.1) is very similar to that of Central Iran, although it is harder to interpret owing to sparse outcrop and because the relation of well-studied outcrops to each other are covered by younger rocks (Fig. 2, locs. 24, 25, and 26). In NW Iran, an Infracambrian to Middle Triassic platform cover contains no evidence of late Palaeozoic orogeny and little, mainly mafic volcanism. The total thickness of the plateform cover in it is, as in Central Iran, around 3 Km or so, with the notable exception of the Alborz Mountains containing a Palaeozoic thickness exceeding 7 Km. (Fig. 3, loc.26), resembling that on the Tabas block. The volcanic history of the Alborz Palaeozoic is also similar to that of the Tabas block in being livelier than elsewhere in NW Iran (BERBERIAN & KING, 1981). A Mesozoic Pb-Zn-Cu mineralization that occurs in Permian, Triassic, and Jurassic rocks is common to the Tabas (FRIEDRICH, 1960) and Alborz (BRADNER & al., 1981) regions, and underlines the similarities between the two areas.

Neither the Alborz (Fig.2, loc.26) nor the Azerbaijan (Fig.2, loc.24) outcrops display any evidence of strong Mesozoic metamorphism. ROMAN'KO & SHARKOVSKIY (1984) reported, however, from the Kuh-e Dom area (Fig.2 and 3, loc. 25) a late Triassic to early Jurassic low grade greenschist metamorphism together with polyphase penetrative deformation similar to that seen in the S part of the Sanandaj-Sirjan zone. How this metamorphic region may relate to the unmetamorphosed Palaeozoic section in Soh to its immediate W (and to other outcrops in NW Iran) remains unknown.

Near Torud is yet another isolated metamorphic outcrop in NW Iran (Fig.2, loc. To). Its predominantly clastic and carbonate rocks with subordinate volcanics contain Lower-Middle Devonian or possibly Silurian corals. Here the age of metamorphism could be anything between late Devonian and Triassic according to THIELE (1973) and may be even late Jurassic according to HUSHMANDZADEH & al., (1978). Analogy with the Kuh-e Dom region suggests a late Triassic event, but DAVOUDZADEH & WEBER-DIEFENBACH (1987) consider it late Palaeozoic.

Nearly complete Palaeozoic to at least Upper Triassic record exists farther to the NW both in Iran and in Transcaucasia. In the classical Djulfa region in NW Iran and Nachichevan (Figs. 2 and 3, loc. 24), for example, is a Silurian to Middle Jurassic platform cover containing no angular unconformities (BONNET & BONNET, 1947; SCHIKALIBEILI, 1984). In this succession ALTINER & al., (1979) have documented a Lower Griesbachian (Lowermost Triassic) hiatus that was manifestly a result of uplift owing to limited stretching. The Precambrian basement of this zone is exposed about 200 km farther to the NW in the Mishkan zone (Fig.2,loc. Mi; age 660 Ma: ABESADZE & al., 1982).

In the Svanetia region farther N (Figs. 2 and 3, loc.23), ADAMIA & BELOV (1984) documented a complete, nearly volcanic-free (exceptions are Devonian tuffs and tuff breccias of andesitic composition and Middle Carboniferous rhyolitic and dacitic tuffs) sedimentary succession reaching from palaeontologically dated Middle Devonian to Upper Triassic. Lias is in 'abnormal',probably unconformable contact with this Dizi Series (KHAIN, 1975) that had been intensely deformed before the Liassic shales were laid down. The Dizi Series represents a region free of any orogenic deformation between the early Devonian and the latest Triassic and thus resembles other sequences in NW Iran, except that it is now separated from them by the fragments of the former Podataksasi Zone viz. the Dzirula, Khrami, and the Loki massifs. In a single cross-section extending from Svanetia to about Takab (i.e. between locs. 23 and 8 in Fig. 2) therefore the remnants of the former Podataksasi Zone and NW Iran tectonic units are repeated at least twice as follows:

1- Svanetia	»	(NW) Iran
2- Dzirula Khrami Loki	»	(Podataksasi Zone)
2- Mishkan Djulfa	»	(NW Iran)
3- Takab	»	(Podataksasi Zone)

3. DISRUPTION OF THE MIDDLE EASTERN CIMMERIDE OROGEN

Region characterized by a late Palaeozoic orogeny, especially the Podataksasi zone, are also commonly characterised by late Triassic to early Jurassic orogeny that results from continental collision.

Throughout the Podataksasi Zone, Upper Triassic or various levels of Lower to Middle Jurassic continental molasse type deposits sit on older rocks across an unconformity (fig.3, locs. 1-4, 6-8, and 10-11) caused by intense folding, thrusting, and/or strike-slip faulting. The record in the localities 6, 7, 8 and 13 in the Podataksasi Zone in Fig. 3 shows that some of the areas in which there is stratigraphic evidence for Triassic and Jurassic deformation also contain isotopic evidence for it. Areas of known or suspected late Palaeozoic deformation in N and interior Iran (locs. 18,19,20 and 25 in Fig.3.) also show isotopic evidence for it. Areas of known or suspected late Palaeozoic deformation in N and interior Iran (locs. 18, 19, 20 and 25 in Fig.3) also show isotopic evidence for Triassic and Jurassic deformation.

By contrast, in Central Iran Upper Triassic and Jurassic rocks sit on older rocks locally disconformably or across commonly low-angle unconformities formed only by tilting (e.g. STÖCKLIN & al., 1965, Fig 24; STÖCKLIN, 1968). Only locally, and mainly along the main divider faults of Nayband, Kuh-e Kalsaneh, Kuh-e Banan, and Posht-e Badam is there any folding and metamorphism of late Triassic-early Jurassic age in Central Iran (e.g. RUTTNER & al., 1968, Figs. 26 and 27; STÖCKLIN, 1968, Fig. 4; HAGHIPOUR & al., 1977,pp.61-63; BERBERIAN & KING, 1981). But these events were not associated with magmatism in Central Iran, save for a few places around the Central Iranian Microcontinent such as the granites of Airakan and Shirkuh (Fig.2).

The distribution of two types of unconformities under the Upper Triassic or Lower Jurassic is now not as straightfoward as one would expect and implies that a simpler original geometry was later redistributed (Fig.1). After the accretion of the Cimmerian continent into Laurasia, the internal disruption of the former evolution was dominanted by strike-slip faults and that resulted in the 130^0 counter-clockwise rotation of the Central Iranian Microcontinent (DAVOUDZADEH & al., 1981), the transport of the various parts of the Podataksasi Zone into their current places, and the opening of the

Slate-Diabase Zone, Sevan-Akera-Karadagh, and the oceans around the Central Iranian Microcontinent (SENGOR, 1987a, 1990; SENGOR & al., 1988).

Immediately after the accretion of the Cimmerian Continent onto Laurasia, extensional deformation started along its suture. By the Sinemurian, the Slate-Diabase Zone in the Greater Caucasus (Fig.1) began forming astride the Palaeo-Tethyan suture as indicated by fault-controlled subsidence, resulting in the deposition of conglomerates, sandstones, and shales in small basins, along with rhyolitic-dacitic volcanism. BERIDZE's (1984) fig. 27 shows interfingering of turbidite fans that were shed from both the N and the S, suggesting that the Slate-Diabase Zone remained narrow through the early Jurassic.

Along the Sevan-Akera-Karadagh suture in the Lesser Caucasus, early Jurassic stretching formed narrow grabens at the present sites of the Mishkan and Karabagh zones (Fig.2 between locs.Mi and Tb). The Saatly deep well in the Khoura Depression (Fig.2) was sunk through some 4.5 Km.-thick early and middle Jurassic dacites, basalts, andesites,diorite porphyries, and tuffs (ABDULLAEV & GUSEINOV, 1984), suggesting important contemporaneous stretching. In a number of places along the Alborz Mountains, there are tholeiitic 'within plate' basalts immediately underlying the Shemshak Formation. This indicates that collision probably predated the stretching only by a very short interval (BERBERIAN & KING, 1981; BRANDNER & al.,1981). These latest Triassic to early Jurassic extensional events were contemporaneous with compressional events in other places in the Cimmerian Continent. In the Saghand region (W of the Posth-e Badam fault in Fig.2), for instance, Jurassic rocks are unconformable on folded, faulted, and imbricated Upper Triassic metamorphic rocks (HAGUIPOUR & al., 1977).

In the middle to late Jurassic, stretching continued in the Slate-Diabase Zone with increased vulcanicity including tholeiitic and andesitic basalts. The Sevan-Akera-Karadagh grabens stretched further and sea-floor spreading commenced in the resultant through (KNIPPER & SOKOLOV, 1976; GASANOV, 1986). In the Alborz Mountains this extension was represented by basaltic volcanism (e.g. BRANDNER & al., 1981). In the Kopet Dagh basin, extension-related subsidence started in the Bathonian and accelerated progressively to the end of the Jurassic, reaching its peak in the Berriasian (PROZOROVSKIY, 1985).

The interval through the middle Jurassic to early Cretaceous was probably also characterized by extension in the oceans around the Central Iranian Microcontinent, although no Jurassic rocks have yet been identified in them. The presence of early Cretaceous deep marine evironments in the Sabzevar region is clear, however (Fig. 2, Lindenberg et al., 1984).

While extension was going on in the Slate-Diabase Zone in the Greater Caucasus, the Sevan-Akera-Karadagh zone the Lesser Caucasus, the Alborz and Kopet Dagh in N Iran and in the S USSR and in parts of the periphery of the magmatism were active elsewhere in the Middle Eastern Tethysides. From the S part of the Sanandaj-Sirjan Zone, BERBERIAN & KING (1981) reported late Jurassic-early Cretaceous low-to medium-pressure metamorphism and compressional deformation accompanied by andesitic volcanism and intrusions ranging from gabbro to granite with K/Ar ages between 118+or-10 and 164+or-4 Ma (BERBERIAN and BERBERIAN, 1981). Other intrusions of similar age are also present farther along the Sanandaj-Sirjan Zone possibly implying that the subduction of Neo-Tethys may have started already under the Sanandaj-Sirjan Zone by the late Jurassic to early Cretaceous interval (BERBERIAN & BERBERIAN 1981).Granites of this age are also present in the Central Iranian Microcontinent. Near Airakan (Ai in Fig. 2) and in the W part of the Yazd block (Shirkuh granite) plus along the E margin of the Lut block, they may suggest subduction was going on in much of the Sanandaj region itself, fast subsidence resulted in the deposition of 2-3 Km. of shales of Barremian-Aptian age, probably showing local extension. Other structural evidence for crustal shortening of late Jurassic-early Cretaceous age in Iran comes from the Ardakan region, where Jurassic rocks are strongly folded, display slaty cleavage, and are unconformably covered by Albian limestones (HAGHIPOUR & al., 1977).

In the late Cretaceous shortening began to dominate the deformation in the Middle Eastern Tethysides S of the Caspian Sea - Black Sea axis. In both the Slate-Diabase

Zone and the Sevan-Akera-Karadagh oceans S-vergent compressional deformation commenced (KHAIN, 1975). Ophiolite obduction onto the NW Iranian margin from the Sevan-Akera-Karadagh ocean began possibly in pre-Albian time but definitely by the Cenomanian and was completed by late Coniacian (KHAIN, 1975; KNIPPER and SOKOLOV, 1976; GASANOV, 1986). Early Cretaceous through Palaeogene calc-alkaline magmatism was abundant in areas N of the Sevan-Akera-Karadagh suture from NE Turkey to well into the W Alborz (LORDKIPANIDZE, 1980, Figs. 12, 19 and 35).
The onset of the compressional deformation in the Alborz Mountains also took place in the late Cretaceous, although the main deformation was in the Palaeocene. It was largely amagmatic and was accompanied in the few areas only by an incipient metamorphism. The late Cretaceous was a time of stretching N of the Alborz Mountains. It was at this time that the rifting of the S Caspian Sea and the Black Sea (since the Aptian) began (APOL'SKIY, 1974; GÖRÜR, 1988) in an overall right-lateral strike-slip regime independent of the earlier extension related to the left-lateral motion of the Podataksasi fragment, and the Kopet Dagh basin went on subsiding.
the post-collisional, late Triassic to mid-Cretaceous tectonic evolution of the Middle Eastern Tethysides thus displays and irregular and confusing picture when viewed in the conventional, simplistic tectonic scenarios involving concertina-style ocean closing and opening episodes (e.g. JENNY, 1977; BERBERIAN & KING, 1981; BAUD & STAMPFLI, 1989), one in which extension and compression occurred contemporaneously in different parts of the orogen and alternated in the same places in time. A horst-graben pattern formed in the Triassic-early Jurassic in interior Iran, and another, independent one in the late Jurassic-early Cretaceous. Some of these grabens contains more than 5 km. of Jurassic i.e. more than three times thicker than the thickest Jurassic section in the Zagros (STÖCKLIN 1968). Repeated and generally contrasting kinds of deformation along old fault zones, especially in interor Iran and in the S USSR within our area of interest (Fig.2), has been constantly stressed by most workers in this region (e.g. STÖCKLIN, 1968; BERBERIAN, 1981; BERBERIAN & KING, 1981; DAVOUDZADEH & SCHMIDT, 1984 for Iran, KRAVCHENKO & al., 1976; SENGOR 1984; PROZOROVSKIY, 1985 for the Turkmenian SSR). Many of them also realised that most of these repeatedly activated fault zones have exhibited, at different times in their history, considerable strike-slip movement (e.g. RUTTNER & al., 1968; APOL'SKIY, 1974; BERBERIAN, 1981; SENGOR, 1987a), but direct evidence for Mesozoic wrench faulting is present only along the borders of the Tabas block (RUTTNER & al., 1968; BERBERIAN, 1981) and in the Dzirula Massif (ADAMIA & al., 1982; SENGOR, 1987a, 1987b). Indirect evidence is plenty, however, and comes mainly from a combination of shape of fault trace, cross-sectional aspect of fault zone, and palaeomagmatism, supplemented in places by the existence of inferred pull-apart basins. A large number of the major fault zones dividing the Middle Eastern Tethysides into separate tectono-stratigraphic units have many, if not most of the characteristics of strike-slip faults.
RUTTNER & al., (1968) mapped a minimum of 40 to 50 km. of right-lateral offset along the Kuh-e Kalsaneh (Fig.2) by the displacement of the Permo-Triassic facies in the Shirgesht region. Along the S continuation of the same fault zone, there are a number of spindle-shaped Palaeozoic outcrops probably entrained along the fault zone. Farther S, cross section VI in Fig. 5 traverse the fault zone. It exhibits a fan- (or 'palm-tree') shaped cross-section with faults of mainly reverse separation with one exception (fault market 'N'). The age of the initiation of faulting here must be earlier than a certain Cretaceous.
The Dehshir - Shar Babak - Baft fault system on which Recent (and active) right-lateral strike-slip has been observed is another example (BERBERIAN, 1981). Mesozoic strike-slip on this structure has been inferred, owing to an 130^0 counter-clockwise rotation of the entire Central Iranian Microcontinent (DAVOUDZADEH & al., 1981; SOFFEL & FÖRSTER, 1985). Cross-section V in Fig. 5 supports this inference as it shows a fault zone of inconsistent vergence defining two fans side-by-side along which dominant separation is normal with three exceptions (faults marked 'R').
A third example comes from locality 21 (Fig.2), where RUTTNER (in press and pers. comm. 1988) mapped a zone of steep faulting with inconsistent vergence whose cross-

sectional aspect is more of a tulip than a palm tree. He thought that considerable left-lateral strike-slip must have occured along this fault zone (Fig. 5, cross-section I).
The Sanandaj-Sirjan zone is demarcated from the rest of interior Iran either by strike-slip fault zones (Abadeh, Dehshir, Shar Babak, and Baft faults) or by long, linear depressions likely controlled by wrench faults. Cross-section IV in Fig. 5 indicates a style of faulting resembling the cases shown in Fig. 5, cross-sections V and VI. I interpret it as half a palm tree plus another steep zone of faulting caused by significant strike-slip. The cross-section IV in Fig. 5 shows that some of the faults delimiting the Sanandaj-Sirjan Ranges to the NE were active solely in the Mesozoic.
The Alborz and the late Cretaceous to later Cainozoic Greater Caucasus (cross-sections II and III in Fig. 5) both have palm tree-shaped cross-sections and both display an anastomozing fault pattern with characteristic palm tree cross-sections. This is all the more remarkable, since "the Alborz Belt is not an orogenic belt of its own, but only a partial belt of a much vaster orogenic region, which includes all of Iran..." (STÖCKLIN, 1974b, p.220, also 1960). Although the Greater Caucasus does not show alternating normal and reverse separations in any given cross-section, the Alborz does (in cross section II, Fig 5, faults marked 'N' have normal separation). The Greater Caucasus has already been interpreted as a dominantly transpressional orogenic belt (e.g. APOL'SKIY, 1974).
In Transcaucasia and in Turkey the strong late Cainozoic shortening with accompanying widespread volcanism (DEWEY & al., 1986) have largely destroyed the earmarks of older, Mesozoic wrench systems. That they existed is surmised however, from their remnants as in the case of the Chorchana-Utslevi Zone in the Dzirula Massif. Neither the magnitude of offset, nor the original orientation of the fault zone segments preserved in such tiny fragments can be estimated from these random pieces, however, because of the probable rotations around vertical axes they have suffered since (cf. JACKSON & McKENZIE, 1984). All these remnants can tell us is the presence of strike-slip faults in the past.
In the following speculative model much use is made of such strike-slip faults without detailed justification for each owing to space limitations, and also because for a number of the wrench faults postulated below such independent justification, other than obvious map and cross-sectional aspects, is not yet available.

4. LATE PALAEOZOIC AND MESOZOIC EVOLUTION OF THE MIDDLE EASTERN TETHYSIDES

The tectonic model presented here is based in addition to the data reviewed above, on the palaeomagnetic evidence provided by WENSINK (1979, 1982; 1983). WENSINK and VARECAMP (1980), WENSINK & al., (1978), and SOFFEL & FÖRSTER (1984). The main features on the model are the recognition of the Podataksasi Zone as a once coherent arc, the 130^{o} counter-clockwise Mesozoic rotation of the Central Iranian Microcontinent, and their union through extensive strike-slip faulting through the Mesozoic.

4.1 Late Palaeozoic

In the Carboniferous and early Permian, the NE margin of Gondwana-Land between Turkey and Australia was flanked intermittently but for long distances by a continental margin arc. It included the Podataksasi Zone in the W, the Alborz and Tabas regions in the Middle, and Central Pamirs (SHVOLMAN, 1978) in the E. In the Podataksasi Zone and in the Central Pamirs, it was well-developed and long-lived (in the Podataksasi Zone early Carboniferous to late Triassic, in the Central Pamirs early Permian to Triassic). In the Alborz and in the Tabas block, however, its presence is suggested by mainly mafic with subordinate intermediate volcanism, and by early deformation in the Saghand region, in the northernmost Tabas block (Kashmar region: BRATASH, 1974), and in Djulfa (ALTINER & al., 1979). If ROMAN'KO & SHARKOVSKIY's (1984) and DAVOUDZADEH & WEBER-DIEFENBACH's (1987) views about late Palaeozoic

metamorphism and deformation in the NW-Iran block is right, these may be considered as possible manifestations of the late Palaeozoic arc in NW Iran.
The reconstruction of NE Gondwana-Land in Carboniferous and early Permian time is achieved by placing all the rifted bits of continental material shown in Fig. 6A (i.e. parts of the Cimmerian Continent in the Middle Eastern Tethysides), except the E Qangtang block, back into their original locations. The way this should be done is as follows: In Turkey, record of Triassic orogeny (e.g. locs.2 and 4 in Fig. 2) occurs on both sides of Neo-Tethyan sutures. Any reconstruction must bring these areas into some sort of continuity (SENGOR & al., 1984; YILMAZ & al., 1987). Once the Rhodope-Pontide fragment, the Sakarya continent, and the Kirsehir block are put back into their former locations following SENGOR & al., (1984), the W end of the Podataksasi Zone becomes fixed. Its eastern continuation in Transcaucasia (D, AT/AK in Fig. 6A), the Sanandaj-Sirjan Zone, the Makran and the eastern Lut are then placed against Arabia. Because NW Iran and Central Iranian Microcontinent have late Proterozoic Pan-African basements and Gondwanian Palaeozoic cover similar to those of Arabia, they must have had a position more 'internal' than the Podataksasi Zone with respect to Gondwana-Land. As the Alborz and the Tabas regions are interpreted as former passive continental margins turned into continental margin arcs, I align them with the Podataksasi Zone and put NW Iran and Central Iranian against Gondwana-Land as the SE prolongation of the Arabian Platform (see also SENGOR, 1987a, 1990; SENGOR 1988).
Because shortening continued well into the Permian in a number of places in the Podataksasi arc it must have remained a compressional arc *à la* DEWEY (1980), at least in places, between early Carboniferous and a certain early Permian. In the early Permian it started to switch character and was already extensional by the mid-Permian (except in Golpayan area and in Turkey?). This extension may have ended the compressional deformation in Oman and led to the beggining opening of Neo-Tethys all along the Zagros, as suggested by the Dehbid and Omani basalts and normal faulting in the Kuh-e Dinar area (SZABO & KHERADPIR, 1978). In Oman, the Hawasina basin (Hamrat Duru basin of BECHENNEC, 1988) localised the stretching and BECHENNEC's (1988) Baid platform was probably torn away from Arabia as a part of the Podataksasi arc.
In SENGOR (1990) I discussed the back-arc basin interpretation of Omani Neo-Tethyan basins more fully and underlined, in comparison with such active back-arc basins behind rifted ensialic arcs as the Tyrrhenian and the Japan Sea, that the 'intraplate' geochemical signatures of the Hawasina volcanics were <u>not</u> diagnostic of tectonic environment. Recently, however, ZIEGLER & al., (1990) described the world's first occurence of metacarbonalites in the metamorphic sole of the Semail ophiolite nappe in the Dibba Zone. ZIEGLER & al., (1990) maintain that carbonatites only occur in intracontinental rift zones or, rarely, in oceanic islands. Their finding may thus be seen to provide decisive support for the 'intracontinental rift' interpretation of the initial Hawasina rifting in Oman.
Indeed, I know of no carbonatite flows associated with *modern* back-arc basins. However, PELL & HÖY (1989) have recently described carbonatites, associated syenites, and alkaline ultramafic rocks of Devonian-early Carboniferous age from the Foreland Belt of the Canadian Cordillera associated with back-arc rifting behind a volcanic arc that is known in southern British Columbia (MONGER & PRICE, 1979; p. 786 and fig. 10a). It thus seems possible to accommodate the Omani carbonatites and the associated alkalic rocks in a back-arc environment.

4.2 Triassic

By the early Triassic (Fig. 6A), intracontinental stretching behind the Podataksasi arc had created a back-arc basin complex between it and N Gondwana-Land from Turkey to Oman. In Turkey, this back-arc basin contained an intra-basin platform, the Kirsehir block, reminiscent of the present Yamato bank in the Sea of Japan. In Oman, BECHENNEC's (1988) Baid platform began disintegrating by middle to late Triassic

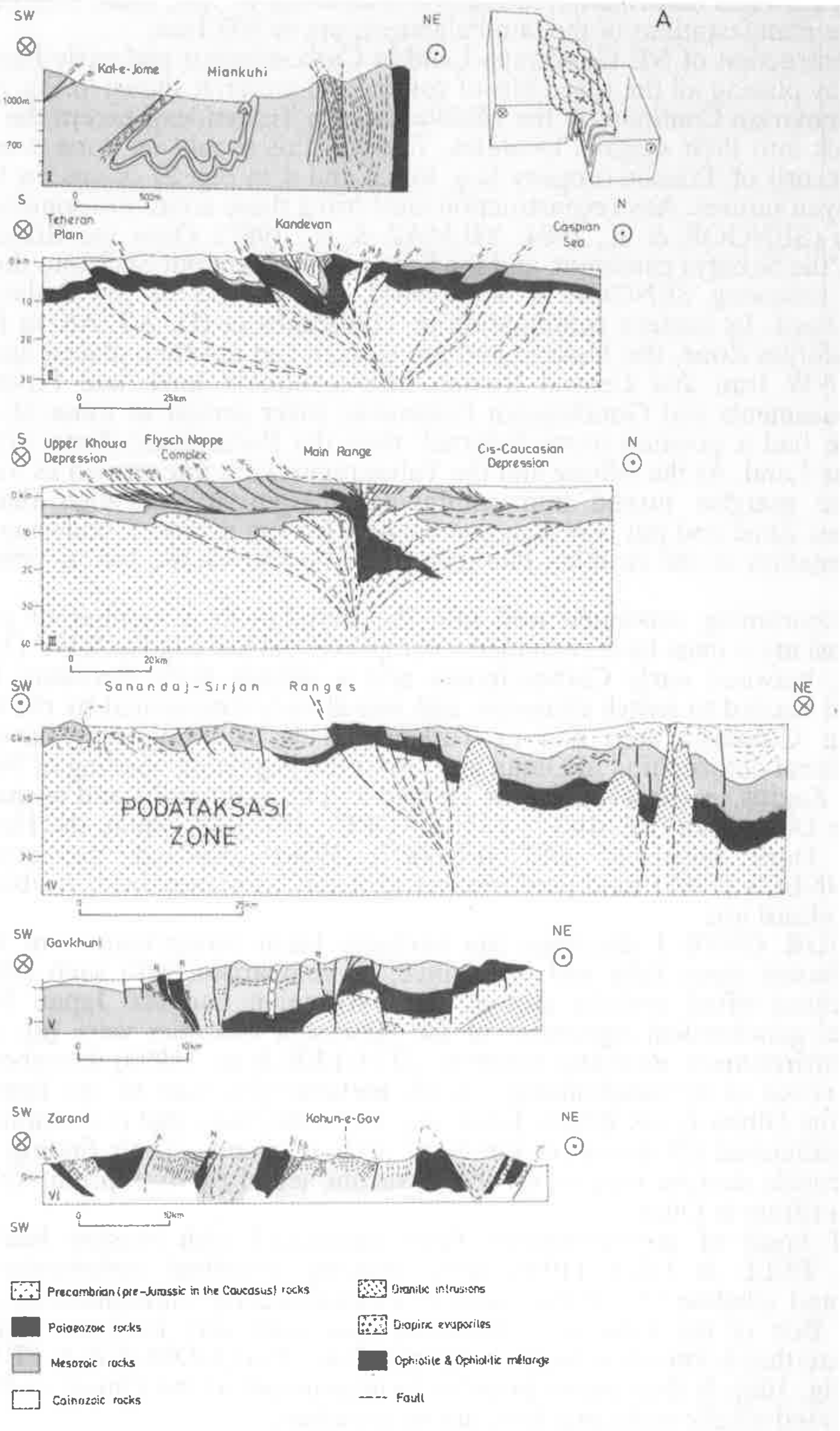


FIG. 5. Selected geological cross-sections from the Middle Eastern Tethysides exhibiting the important, inplaces dominant 'strike-slip-like' profiles, after SENGOR (1990). In all secctions, except in IV, note the inconsistency of the vergence across the strike. For location see Fig. 2. Inset A was redrawn after LOWELL (1972) to illustrate the mechanism that is inferred to be largely responsible for the generation of the across-sectional styles shown in this figure.

time, the locus of extension having jumped northeastwards, and the main Neo-Tethyan basin started forming. Many of the Oman exotics were seamounts in a growing back-arc basin, similar to the Magnaghi, Vavilov, Marsili, and Usticha seamounts in the Tyrrhenian Sea and were left on the Arabian side as the locus of extension migrated away from Arabia. As in the Tyrrhenian Sea, some of these seamounts probably had an oceanic substratum (e.g. Glennie 1974), while others may have been constructed on continent (e.g. BECHENNEC,1988).
Fig. 6B displays the Rhaetian situation in the Middle Eastern Cimmerides. The Podataksasi arc and NW and Central Iran were accreted to Laurasia along the main Palaeo-Tethyan suture *as single block*, but some coeval wrench faulting is suggested in Central Iran in the Saghand area (HAGHIPOUR & al., 1977) and the Shotori Range respectively (STÖCKLIN & al., 1965). Molasse deposition (the right-lateral movement of the Yazd, Tabas, and Lut blocks were simultaneously disrupting this simple pattern (Fig. b 6B). Following the collision in Iran, Palaeo-Tethys became reduced to Eastern Mediterranean-type remnant oceans in N Turkey and in Afghanistan. In both regions, large accretionary flysch/mélange wedges resembling the present-day Eastern Mediterrannean Ridge (LE PICHON, 1983) were evolving (Fig. 6B). The N Turkish remnant of Palaeo-Tethys, which was being filled up by the Akgöl (N Turkey), Tauridian (Crimea), Nalbant (N Dobrudja), and Lipacka (Strandja Mountains) flysch complexes, shut by the later early Jurassic possibly by an initial right-lateral motion of the Cimmerian Continent with respect to Laurasia, as indicated in Fig. 6B. Between the Caucasus and Afganistan, Palaeo-Tethys closed along two subduction zones of opposing polarity. The initial right-lateral motion of the Cimmerian Continent was probably localised along this suture and may have caused the shearing seen in the Chorchana-Utslevi Zone (Fig. 6B, loc. D, SENGOR, 1987a). The same movement probably also caused the re-Shemshak basalt erruptions in N Iran.

4.3 Jurassic

An important changes in the evolution of the Middle Eastern Cimmerides possibly as early as the Pliensbachian was the onset of the independent left-lateral *coastwise transport* (cf. BECK 1988), of the Podataksasi arc along a line coincident with the Palaeo-Tethyan suture, but passing S of the NW and Central Iranian fragments. The rifting of the Slate-Diabase zone in the Greater Caucasus may have been a pull-apart event. The stretching that followed the collision in the Lesser Caucasus (Sevan-Akera-Karadagh rifting), in the Khoura basement (volcanics in the Saatly well), and in N Iran (Shemshak basins) may represent secondary extension related to the strike-slip motion of the Podataksasi zone (Fig. 6C and C'2). The counter-clockwise rotation of Iran commenced in the Jurassic (SOFFEL & FÖRSTER, 1984) and was probably accomplished by the initial rifting of the Nain-Baft ocean, the western segment of the Circum-Central Iranian Microcontinent oceans.
The rotation of Central Iran was probably caused by the coastwise transport of the Podataksasi Zone and by Central Iran's being 'pinned' in the NE by the already collided Farah block in early Jurassic time creating a shear couple (Fig. 6C). This interpretation brings and attractive solution to the otherwise hard-to-explain independent rotation of the Central Iranian Microcontinent.
Fig. 7 shows a possible present-day analogue of the isolation and beginning rotation of the Central Iranian Microcontinent. The Ordos block in Central China is surrounded by the Helan Shan and the Shansi Graben System that formed in response to the strike-slip motion on the Altin Dagh fault. The motion of the Qaidamu northeastwards with respect to North China seems to be ripping the Ordos away from the Alxa region much like the Podataksasi Zone ripping Central Iranian Microcontinent away from Northwest Iran.
Throughout the Jurassic, the Slate-Diabase Zone, Sevan-Akera-Karadagh, and Nain-Baft oceans continued to expand. In the Sevan-Akera-Karadagh ocean, sea-floor spreading began in middle to late Jurassic time and extension probably started in the Kopet Dagh basin in the Middle Jurassic. Extension in the Kopet Dagh probably also

resulted from secondary extension related to the strike-slip motion of the Podataksasi Zone (Fig.6C'2).
Compressional deformation, metamorphism, and arc-related magmatic activity in Iran and in Caucasia increased in the middle to late Jurassic. This is ascribed partly to the onset of subduction both under the E and S margin of the Lut block and under the SW margin of the Sanandaj-Sirjan Zone. Evidence for these events is provided by the Sorkh-Kug granite in the central Lut (Fig. 6C), and isotopic ages atound 170 Ma indicating metamorphism in the Sanandaj-Sirjan Zone and the Deh Salm region in eastern Lut (BERBERIAN & KING, 191; DAVOUDZADEH & SCHMIDT, 1987).
The late Jurassic subduction along the E and S sides of the Lut block and along the S part of the Sanandaj-Sirjan Zone throughout the late Jurassic and the early Cretaceous and arrived at the S margin at the Rhodope-Pontide fragment by the Albian (Fig. 6D, GÖRÜR, 1988).

4.4 Cretaceous

In the Early Cretaceous, convergent plate-margin acivity increased in the Middle Eastern Tethyside (Fig. 6D). It also witnessed a regression from wide areas of NW Iran, Central Iranian Microcontinent, and the Sanandaj-Sirjan Zone. The time of most widespread emergence (late Neocomian: STÖCKLIN & STUDEHNIA, 1977, fig. 4) is not coincident with the time of lowest early Cretaceous sea-level (early Neocomian: see HAQ et al., 1988, fig. 15), although the effects of a world-wide low stand on the major early Cretaceous regression in Iran cannot be denied. Especially in northern Iran, however, the steadily rising sea-level between the Neocomian and the Cenomanian is not reflected in the regional stratigraphy, further underlining the partial independence of the early Cretaceous regression in Iran from the world-wide stand displacements. Palaeomagnetic evidence shows that continuing rotation of the Central Iranian Microcontinent (SOFFEL & FÖRSTER, 1984) was accompanied by right-lateral strike-slip along its main divider faults. Accumulation of large thicknesses of Barremian to Aptian clastics, to the W of Posht-e Badam, suggests the presence of pull-apart basins along the Yazd-Tabas boundary fault zone respectively.
Subduction migrated northwards along the Sanandaj-Sirjan Zone and probably reached the easternmost E Pontides, where the 'Lower Basic Series' probably represent its earlier products (SENGOR & al., 1984). Ophiolite obduction in the Sevan-Akera-

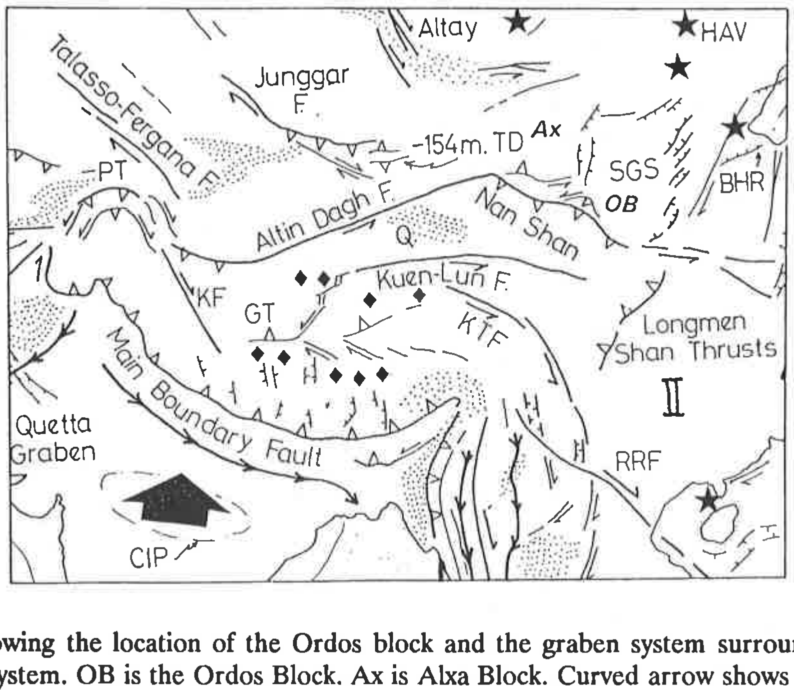


FIG. 7. Map showing the location of the Ordos block and the graben system surrounding it. SGS is the Shansi Graben System. OB is the Ordos Block. Ax is Alxa Block. Curved arrow shows the presumed sense of the rotation of the Ordos Block.

Karadagh ocean may have started in pre-Albian time, although the final emplacement occured in the Cenomanian and Coniacian. The E Lut subduction zone stayed active probably through the early Cretaceous, but stopped by the Aptian. *Orbitolina* limestones of this age unconformably cover its products in the Shah Kuh and in the Deh Salm metamorphics. Probably the Sabzevar-Sistan subduction zone started its activity coevally or immediately following the demise of the E Lut subduction zone, although the first evidence for the former is the Upper Cretaceous (Campanian) flysch and subduction-related volcanism (TIRRUL & al., 1983; BAROZ & al., 1984; LINDENBERG & al., 1984). I view the Sistan and the Sabzevar subduction zones as parts of the same convergent margin that formed along the W edge of the combined Farah and Helmand blocks (Fig. 6D and E, SENGOR, 1984; BOULIN, 1988). This 'larger' Sistan subduction zone was probably generated by the westerly motion of the united Farah/Helmand blocks along the Palaeo-Tethyan suture (Fig. 6D).

During the Campanian the Eurasia/Africa relative motion became very nearly purely convergent (SAVOSTIN & al., 1986). This change also coincided with increased convergent margin activity, in *and* actual convergent strain across, the Middle Eastern Tethysides (Fig.6E). This compressive regime continued into the Palaeocene, and the "Eocene everywhere overlies various older formations with pronounced angular unconformity" (STÖCKLIN, 1968, p. 1244).

By the end of the late Cretaceous, the Central Iranian Microcontinent had nearly completed its counter-clockwise rotation, suggesting that the Podataksasi Zone had also ended its coastwise strike-slip movement. One reason why this movement stopped may have been because slivers of the Podataksasi Zone (the Karabagh zones) were shaved off the main body during its coastwise transport under increasing N-S compression and inserted sideways along strike-slip faults into the closing Sevan-Akera-Karadagh ocean as shown in Fig. 6E. This sideways insertion explains the repetition of parts of the Podataksasi Zone and of NW Iran in one cross-section as discussed above. The 330+or-42 Ma-old eclogite-amphibolites associated with the Sevan-Akera ophiolites (MELIKSETIAN & al., 1984) maybe remnants of the Palaeo-Tethyan subduction complex dragged into the Sevan-Akera-Karadagh ocean and mixed with its own much younger ophiolites. The complicated late Cretaceous strike-slip faults postulated here receive a weak support results also from the palaeontology. VÖRÖS (1988) pointed out that the distribution of early and middle Jurassic (Pliensbachian: PROZOROVSKAYA & VÖRÖS, 1988, fig. 3, and Bajocian: ibid, fig. 4) brachiopod provinces in the Caucasus displays a similar 'interfingering' as the tectonic units (see his fig. 9). The Dzirula Massif (Podataksasi) displays a 'Mediterranean' ('Tethyan') province that normally occurs north of the Dzirula. The weakness of the palaeontologic support results from the ill-defined and commonly non-diagnostic nature of brachiopod provincialism from the viewpoint of palaegeography (although some brachiopod specialists disagree: e.g. Ager, 1988).

Immense quantities of flysch wereadded to the Neh and Ratuk accretionary complexes in the Sistan region (TIRRUL 1983) and to the Dowlatabab unit in the Sabzevar zone (LINDENBERG & al., 1984) (Fig. 6E).

The compressional structures dominating much of Central and NW-Iran, Transcaucasia, and the S Pontides contrast sharply with extensional structures in the Black Sea, Riou and Khoura depressions, and the South Caspian Sea. In Iran, the dominantly compressional southern areas were separated from the dominantly extensional northern areas by the Symmetrical Alborz Ranges. Following APOL'SKIY (1974) I view the South Caspian ocean as pull-apart basin along a continuous acestral Herat/Alborz/Greater Caucasus right-lateral strike-slip zone (Fig. 6E'). Both SHIKALIBEILI & GRIGORIANTS (1980) and BERBERIAN (1983) marshalled abundant data to show that it is a young, late Mesozoic (SHIKALIBEILI & GRIGORIANTS, 1980) or even early Cainozoic (BERBERIAN 1983) structure. [7] The Khoura basin is the westerly prolongation of the South Caspian pull-apart and probably is related to it geneticaly, having formed as a deep through in the Oligocene (TAGIYEV, 1984).

The rifting of the Black Sea commenced in the Albian by extensive normal faulting that remained active until the Cenomanian in N Turkey (GÖRÜR, 1988). Its easterly continuation, the Adzharia-Trialeti trough, accumulated 6 km. of mildy alkaline to alkaline basaltic rocks during the early Palaeogene (LORDKIPANIDZE & al., 1977). The Caspian-Black Sea strike-Slip system shifted the entire Cimmerides orogenic collage in the Middle Eastern Tethysides for a minimum of 500 km. (as judged from the

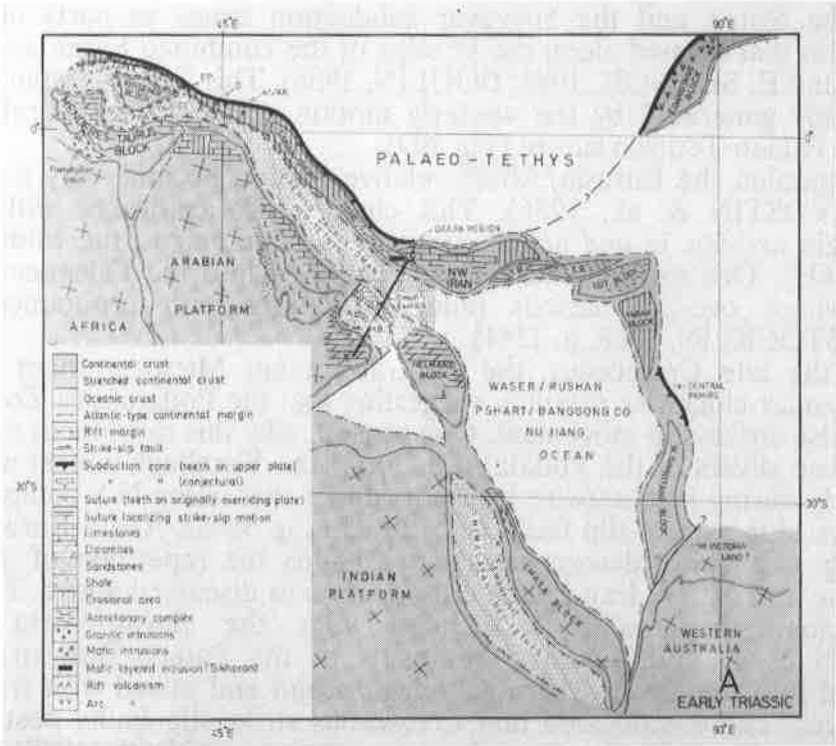
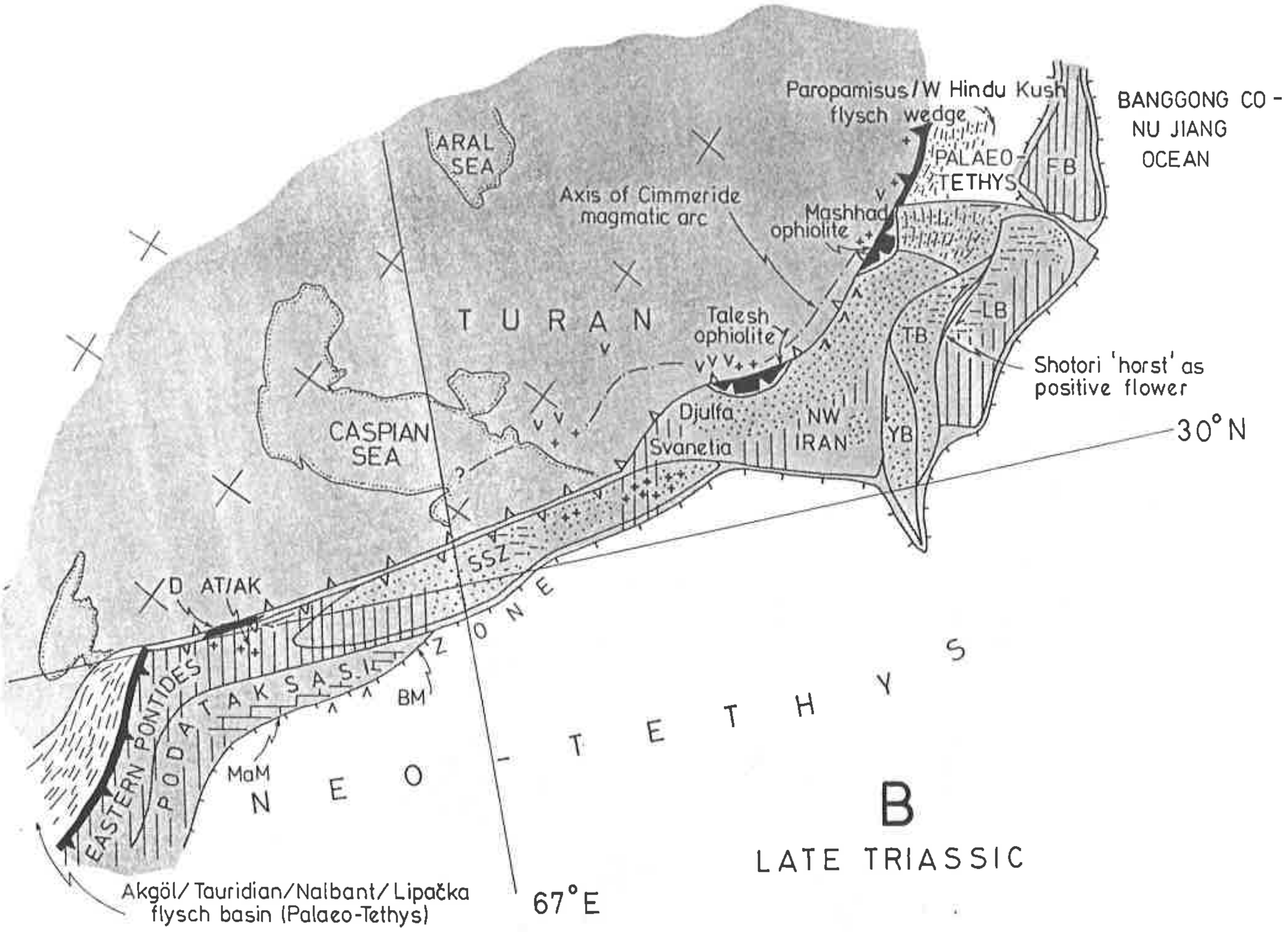
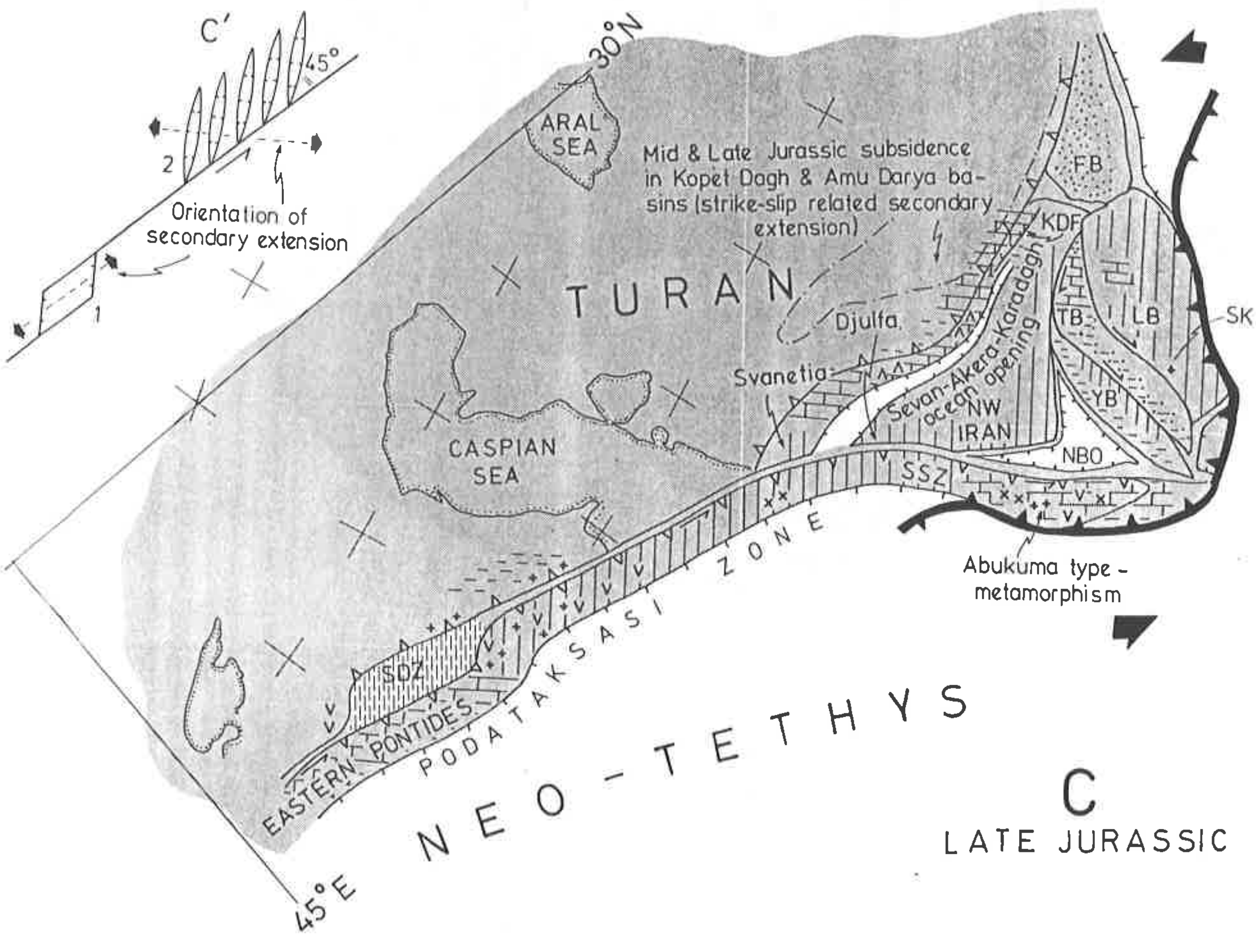
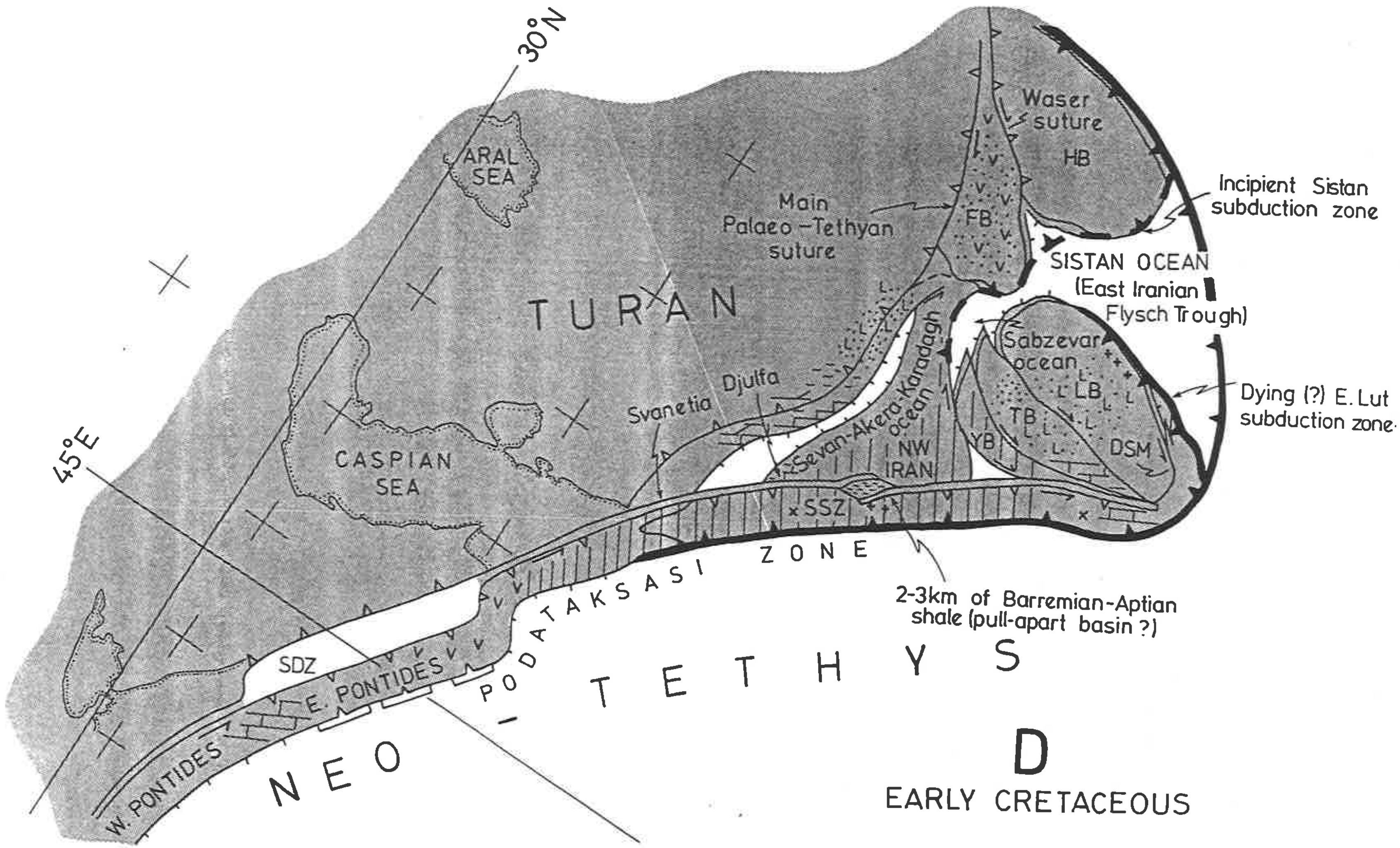


FIG. 6. Palaetectonic maps showing the tectonic evolution of the Middle Eastern Tethysides from the early Triassic to late Cretaceous after SENGOR (1990). Key to abbreviations: AT/AK - Adzharia/Trialeti and Artvin/Karabagh zones, BM - Bitlis Massif, D - Dzirula Massif, DSM - Deh Salm Metamorphics, FB - Farah Block, HB - Hawasina Basin (in A), Helmand Block (in B-E) KDF - Kopet Dagh Fault (normal), LB - Lut Block, MaM - Malatya Metamorphics, NBO - Nain/Baft Ocean, SDZ - Slate-Diabase Zone, SO - Sabzevar Ocean, SSZ - Sanandaj-Sirjan Zone, TB - Tabas Block , YB - Yazd Block (LB+TB+YB = Central Iranian Microcontinent).

A. Palaetectonic map of the South-Central Tethysides in early Triassic time. Wavy symbol at the southern extremity of the Sanandaj-Sirjan Zone shows contemporary orogenic deformation and metamorphism.
B. Palaeotectonic map of the northern and middle parts of the Middle Eastern Tethysides in late Triassic time.
C. Palaeotectonic map of the northern and middle parts of the Middle Eastern Tethysides in the late Jurassic.
D. Palaeotectonic map of the northern and middle parts of the Middle Eastern Tethysides in early Cretaceous time.
E. Palaeotectonic map of the northern and middle parts of the Middle Eastern Tethysides in the late Cretaceous. Here notice especially how splinters of the Podataksasi zone (Dzirula, Adzharia-Trialeti basement and the Artvin-Karabagh zones) are inserted sideways into the closing Sevan-Akera-Karadagh ocean.

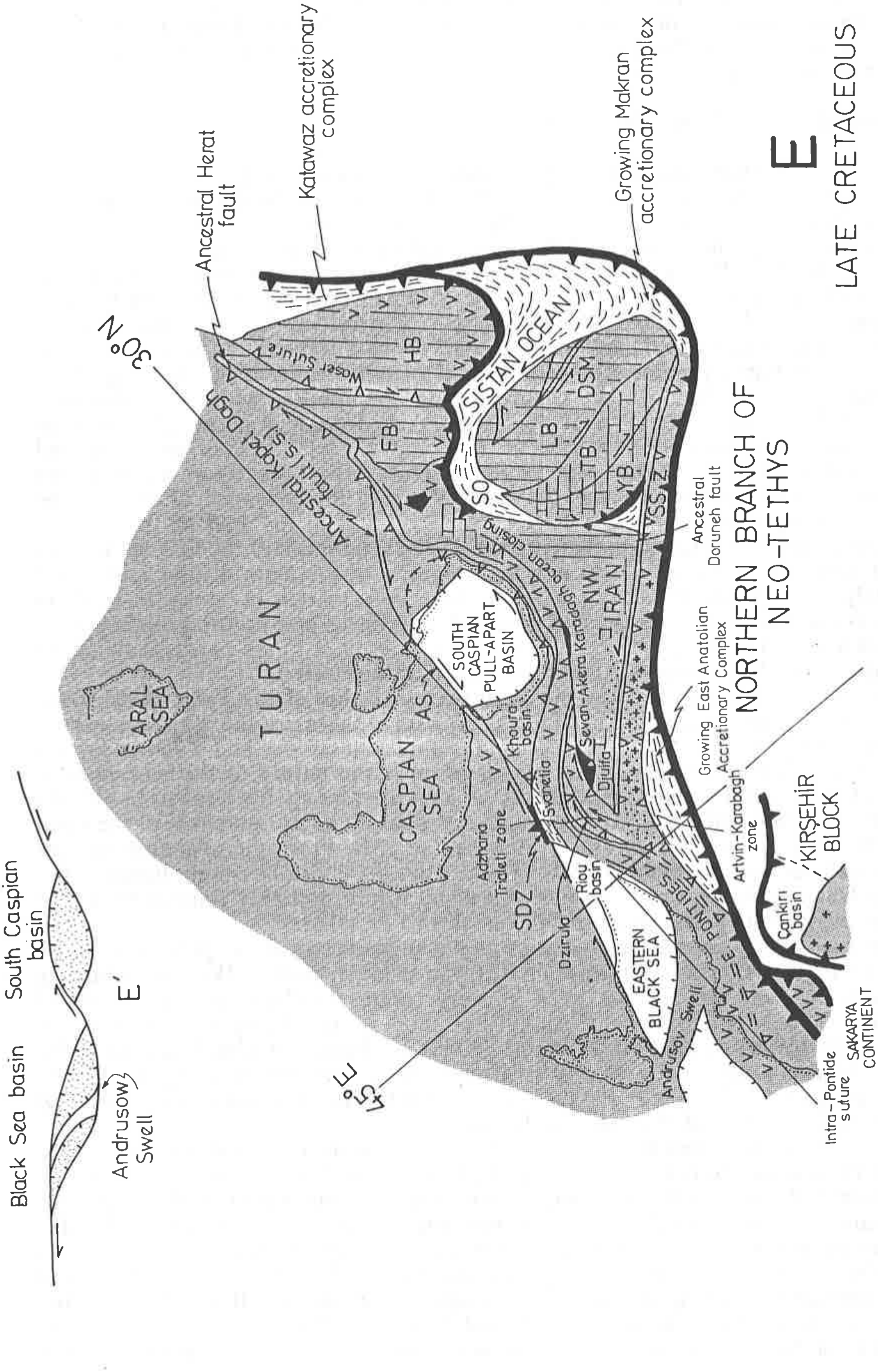

Paropamisus/W Hindu Kush
flysch wedge
BANGGONG CO -
NU JIANG
OCEAN
ARAL
SEA
PALAEO-
TETHYS
FB
Axis of Cimmeride
magmatic arc
Mashhad
ophiolite
LB
T U R A N
Talesh
ophiolite
TB
Shotori 'horst' as
positive flower
CASPIAN
SEA
Djulfa
NW
IRAN
YB
Svanetia
30°N
SSZ
D
AT/AK
ZONE
EASTERN PONTIDES
PODATAKSASI
BM
MaM
NEO-TETHYS
B
LATE TRIASSIC
Akgöl/Tauridian/Nalbant/Lipačka
flysch basin (Palaeo-Tethys)
67°E


C'
45°
2
1
Orientation of
secondary extension
30°N
ARAL
SEA
Mid & Late Jurassic subsidence
in Kopet Dagh & Amu Darya basins (strike-slip related secondary
extension)
FB
KDF
T U R A N
Djulfa
TB
LB
SK
Svanetia
Sevan-Akera-Karadagh ocean opening
YB
NW
IRAN
CASPIAN
SEA
NBO
SSZ
ZONE
Abukuma type -
metamorphism
SDZ
EASTERN PONTIDES
PODATAKSASI
NEO-TETHYS
C
LATE JURASSIC
45°E



D

EARLY CRETACEOUS


Black Sea basin
South Caspian basin
Andrusow Swell
E'
Ancestral Herat fault
Katawaz accretionary complex
Growing Makran accretionary complex
E
LATE CRETACEOUS
30°N
45°E
TURAN
ARAL SEA
CASPIAN SEA
AS
SOUTH CASPIAN PULL-APART BASIN
Ancestral Kopet Dagh fault (s.s.)
Waser Suture
HB
FB
SO
SISTAN OCEAN
DSM
LB
TB
YB
SSZ
Ancestral Doruneh fault
Sevan-Akera Karadagh ocean closing
NW IRAN
Khoura basin
Svanetia
Djulfa
Adzharia Trialeti zone
SDZ
Riou basin
Dzirula
EASTERN BLACK SEA
Andrusov Swell
PONTIDES
Growing East Anatolian Accretionary Complex
NORTHERN BRANCH OF NEO-TETHYS
Artvin-Karabagh zone
KIRŞEHIR BLOCK
Çankırı basin
Intra-Pontide suture
SAKARYA CONTINENT

present width of the oceanic area in the South Caspian basin) westwards with respect to stable Eurasia in latest Cretaceous to about Eocene time (Fig. 6E).
Because by the end of the Mesozoic era the Sanandaj-Sirjan Zone, and with it rest of the Podataksasi Zone, reached more-or-less their present positions with respect to other units in Eurasia, I here end my account of the history of the Middle Eastern Tethysides. The purpose of this and the previous sections was simply to justify the unorthodox reconstruction depicted in Fig. 6A.

5. DISCUSSION AND CONCLUSIONS

The data reviewed above and graphically displayed especially in Fig. 3 underlines most importantly the late Palaeozoic and earliest Mesozoic unity of the various elements of the former Podataksasi zone. The recognition of the former coherence of this zone constitutes the key to the solution of the major palaeo-tectonic riddles of the Middle East. The entire late Palaeozoic and Mesozoic evolution of this part of the world, including such enigmatic events as the middle Permian opening of the Omani Neo-Tethys and the Mesozoic counter-clockwise rotation of the Central Iranian Microcontinent are readily explained in the framework of this model. In fact, its strength lies in the explanation it provides for such diverse tectonic events as the opening of the Slate-Diabase zone in the Greater Caucasus and the modest width it maintained throughout its life-span, the opening of the Sevan-Akera-Karadagh ocean later than the various branches of Neo-Tethys in eastern Turkey, the pre-Shemshak extension and basalt extrusion in the Alborz, the curious across-strike repetition of different tectonostratigraphic entities in Caucasia and NW Iran, the presence of bits of the Sanandaj-Sirjan zone to the south (Bajgan Complex) and to the east of the Central Iranian Microcontinent (Deh Salm metamorphics) the separation of the Central Iranian Microcontinent into three sub-blocks and the longevity of the main divider faults, and the allegedly isolated late Palaeozoic compressional deformation in Oman, in addition to the two events mentioned above. This model clearly solves many more problems than it generates and that is why I considered it useful to present it to the participants of the IGCP Project No. 276 as a working hypothesis to be further tested.
The main problem it faces concerns the westerly connection of the Podataksasi Zone and the places that spring to mind in which to look for it there are the Rhodope Massif in Bulgaria and Greece, the Serbo-Macedonian Complex in Greece and Yugoslavia, and the Pelagonian Massif in Greece. Another problem is the nature of the basement of Turan (Turkmenian SSR) and here a significant step has already been taken by BAUD & STAMPFLI (1989) who distinguished an independent island arc system N of the main Palaeo-Tethyan suture in N Iran and S Turkmenian SSR. From the available meagre subsurface data in the Turkmenian SSR, I suspect that its entire basement between the Sultan Uiz Dagh ophiolite and the Talesh-Mashhad suture probably consists mainly of subduction-accretion material with possibly one or two island arcs.
One of the main conclusions of this study is that we must overcome our prevalent fixistic and cylindiristic tendencies in interpreting Tethyan palaeotectonics. We must not only use comparative tectonics, but we must be very careful in choosing our objects of comparison. I believe that such simplistic models of Tethyan evolution as ARGYRIADIS (1974) or DECOURT et al. (1986) are based on what I believe to be inappropriate comparisons between past environments and present-day tectonic settings. *An important goal of the IGCP Project No. 276 should be a detailed comparison of Palaeozoic environments with their present-day analogues.*
The conclusions of this paper have important implications for our understanding of the entire late Palaeozoic Tethyside evolution in particular, and for the study of orogenic belts in general. We saw in the foregoing sections that the common view of the 'passive margin' character of much of the late Palaeozoic northern margin of Gondwana-Land is wrong. Instead, much of that margin in Turkey (also farther W? see SENGOR & al., 1984), in the Transcaucasia, and in Iran, as in the Central Pamirs and in N Tibet and Thailand, appears to have been 'active' i.e. a magmatic arc orogen. It was the changing character of this semi-continuous arc that led to the birth and governed the early development of Neo-Tethys behind it. Enormous amounts of syn- and post-collisonal

strike-slip faulting, coupled with polyphase compressional deformation, hid large segments of this arc within fragments of its Gondwanian hinterland inside the Tethyside orogenic complex and obliterated others entirely.
The late Palaeozoic and Mesozoic evolution of the Middle Eastern Tethysides also has taught us that many of our reconstructions of former continental positions in the Tethyan realm may have errors exceeding 1500 km, mainly because of the abundance and the tremendous magnitudes of unrecognised strike-slip motion. Consequently a comprehensive evaluation of the tectonic history of a small area of a few 10^5 km^2 in size may require knowledge of an area nearly 10 times larger.

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I thank Aymon Baud and Gérard Stampfli for inviting me to Lausanne to read this paper which updates a similar one published in the Geological Society of London Special Publication 49 on the "Geology and Tectonics of the Oman Region". G. Stampfli also read a first draft and made numerous comments. Although I was unable to agree with all of them, his review did result in many changes and additions which I think improved the paper, although I must take the blame for any remaining errors. A. Baud, G. Stampfli and Jean Marcoux informed me of the new developments in the Omani geology during the Lausanne meeting. Jamshid Eftekhar-Nezhad kindly sent me a preprint of his important paper with A. Behroozi and has been responsible for keeping me up-to-date about Iranian geology. I am very grateful to Dr. D. Rabu for presenting me with a copy of his superb thesis and for sharing with me his knowledge of the 'autochthon' of the Oman Mountains. I am also indebted to Dr. F. Bechennec for discussional and some unpublished information. Dr. L.H. Royden of M.I.T. suggested the Ordos analogy for the formation of the Central Iranian Microcontinent. D. Papanikolau and E. Demirtasli encouraged me to take part in the activities of the IGCP Project No. 276. This work was supported by grants from a group of oil companies awarded to W.C. Pitman, III (Lamont-Doherty), J.F. Dewey (Oxford) and A.M.C. Sengör (I.T.Ü). The paper was written while I was a Royal Society Visiting Research Fellow in the Department of Earth Sciences at Oxford.

NOTES

1 - BELOV's (1981) Meso-Tethys denominates mostly oceans whose development may be considered a part of the Palaeo-Tethyan evolution. However, the origin of the Waser/Rushan-Pshart/Banggong Co-Nu Jiang ocean and its possible connection to Neo-Tethys via the Sistan Zone and the Hawasina basin remain unclear. Their future clarification may possibly justify BELOV's nomenclature.

2 - During the 15th-18th January 1990 "Symposium on ophiolite genesis and evolution of oceanic lithosphere" in Muscat, Oman, I had the opportunity to speak with M. RABU concerning the presence of late Palaeozoic shortening in Oman. M. RABU kindly informed me that he had not altered his opinion expressed in his thesis (RABU, 1988) in regarding the late Palaeozoic deformation as "gentle folding devoid of cleavage, possibly of externide type".

3 - ALAVI's new mapping of the Mashhad ophiolite has clearly shown the accretionary complex nature of this outcrop area. His results, including and excellent palaeogeographic and structural analysis, will soon be published.

4 - The Adzharia-Trialeti Zone has no outcrops older than the Cretaceous. The presence of an older basement, resembling the one in the Talesh Mountains in N Iran, is inferred (KHAIN, 1975).

5 - Although Ordovician deformation and felsic magmatism related to waning Pan-African orogenic events are known from the Gondwanian parts of Turkey (e.g. SENGOR et al., 1984a) and Iran (SENGOR, 1986), they all seem post-collisional. I am therefore unable to agree with a comment made by STAMPFLI in a review of this paper (March, 1990), that the crystallization ages of the ultramafics might be referred to Pan-African events.

6 - Ziegler (1990) has shown that the climatic infuences on vegetation in the present world serve as an excellent model for the Permian. He documented that the climatically defined biomes for the Permian (Ziegler 1990, figs. 3 and 4) display a symmetry about the equator and that the biome boundaries are essentially at the same latitudes as their modern equivalents. Especially Ziegler's map for the early Permian (Sakmarian, see his fig. 3) world showing the phytogeographic biome distribution makes it clear that climate, rather than separation by the Palaeo-Tethyan *cul de sac* determined plant provinciality. The same biomes existed both north and south of the Palaeo-Tethys in the Permian and I therefore see no reason to use palaeobotanical evidence only to construct paleotectonic assemblies. That is why the presence of a Euramerian flora on Gondwanian fragments (including Arabia see Ziegler's fig. 4) in the western Palaeo-Tethys does not appear troublesome to me, although this has been a constant source of objection to my interpretation of the Cimmeride tectonic evolution of the Middle East. This objection thus is the result of overinterpreting the palaeobotanical evidence.

7 - The age of the South Caspian oceanic hole is still a matter of dispute as there are still some who follow STÖCKLIN's (1974) original interpretation in considering it a remnant of Palaeo-Tethys. Even if none of the direct arguments by SHIKALIBEILI and GRIGORIANTS (1980) and BERBERIAN (1983) had been applicable, it would still have been difficult to view this deep depression as being so old. The neighbouring North Caspian Depression is a filled oceanic pocket, whose origin dates back to the late Devonian (cf. BURKE, 1977). It was filled up to become land by the Permian. It is very unlikely that the smaller South Carpian Depression could have accommodated all the debris shed from its surroundings during the growth of both the Cimmerides and the Alpides and yet still remained a basin retaining oceanic depths! No late Palaeozoic basins survive anywhere in the world with their oceanic bathymetry.

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TECTONIC FEATURES OF HINGGAN-MONGOLIAN OROGENE IN LATE PERIOD *

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Abstract

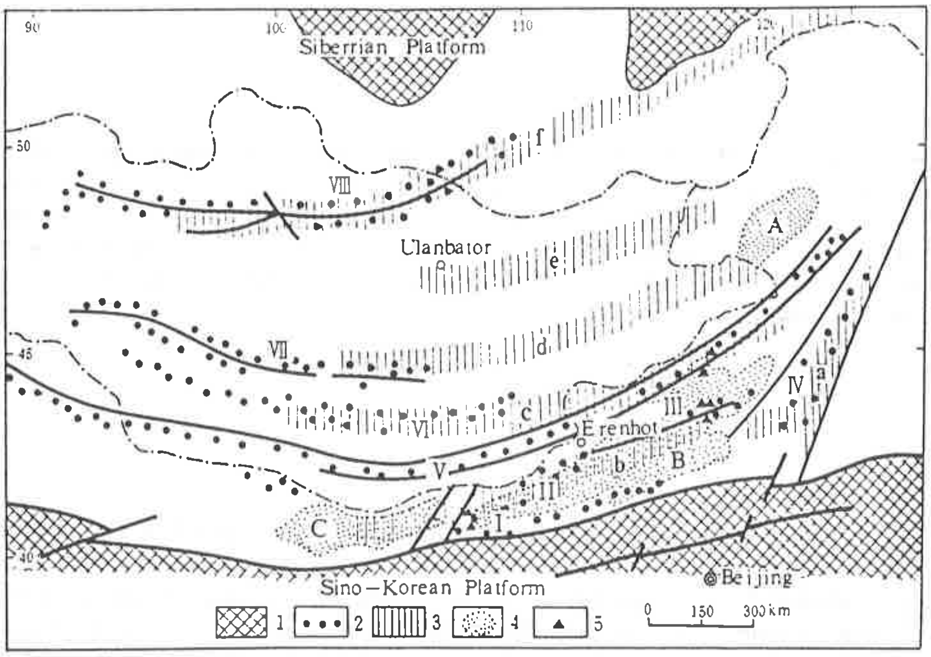
In researches of orogeny there has been a popular point of view: Strong nappes and thrusts always accompany an orogenic zone no matter what kind of movement it has experienced. The large scale A-type subduction occuring during the late period of the Alpine-Himalayan belt and the consequently formed huge nappe structures have been understood by some people as a common feature of the orogenic zones all over the world. This point will be examined in this paper on observation of the features of the Hinggan-Mongolian orogenic zone.

1. TECTONIC FEATURES OF THE OROGENE IN ITS LATE PERIOD

The Hinggan-Mongolian orogenic zone is a section of the Ural-Mongolia-Okhotsk orogene. It was formed in the middle Palaeozoic when the Ural-Mongolian ancient ocean was consumed and the ancient Siberian and Sino-Korean plates colliding to each other, locally preserving molasse formations of upper Devonian Famennian stage. This orogen is distinct from a typical Alpidic orogen in four respects:
(1) An even plateau without steep geomorphic features, and a crust thickness of 42-44km;
(2) Its formations since the late Palaeozoic mainly contains non- or slightly-metamorphic rocks;
(3) The molasse formation accompanying the orogeny is not very developed;
(4) Most importantly, instead of large scale nappe structures, strong folding and A-type subduction, several regional extension activities have occured in the accretionary crust since the late Palaeozoic, forming a series of nearly east-west oriented rift zones, which controls the bimodal volcanic rock belts, coal-collecting belts, and oil-gas-bearing sedimentary basins of various ages.
These four features of the Hinggan-Mongolian orogene should be related to the closure of the ancient ocean and repeated crustal deformations during collision of the plates, as well as to large scale extension activities in the late period of the orogeny.

2. HISTORY OF THE RIFT ZONES MOVEMENT

Fig. 1. Distribution of Late Paleozoic-Mesozoic rift zones in Inner Mongolia and adjacent area.
legend: 1. Platform, 2. Paleozoic rift zones, 3. Coal collecting basins, 4. Oil-gas bearing (oil potential) sedimentary basins, 5. Oil producing area.
I-VIII. The late Paleozoic rift zones: I. The Dahongshan zone, II. Bayan Obo-Zhenglan Qi zone, III. Xiujimgin zone IV. South Hinggan zone, V. Gobi-Tyanshan-Baolihe zone, VI. Main Mongolia zone, VII. Ih-bogd zone, VIII. North Khangai-Orchon-Selenge zone.
a--f: The Late Paleozoic-Mesozoic coal collecting belts:
a. South Hinggan belt, b. Yingen-Xiujimgin belt, c. South Gobi belt, d. Salair-Altai-Gobi belt, e. Kerulen R.-Ergun R. belt, f. Mongolia-Okhotsk belt.
A-B: The Late Paleozoic oil-gas bearing (or oil potential) basins : A. Hailar basin B. Eren basin C. Yingen basin.



Shown in the attached map (Fig. 1) is the distribution of the late Carboniferous-early Permian rift zones in the Hinggan-Mongolian orogene and its adjacent areas. Except the Gobi-Tianshan-Baolige and Main Mongolia rift zones, which are late Carboniferous, all the others are early Permian [1].
Most of them are developed along lines dividing folding zones of different ages or on structurally weaker zones, only the South Hinggan rift zone cuts accross several old structural zones. Between any two adjacent rift zones there is a continental denudation zone. In the rift zones exist very thick (10-20km) sedimentary formations of marine facies or marine-continental-alternating facies. The rapid changes of the stratigraphic thickness from inside to outside the rift zones shows that the accumulation is closely related to the fault activity. The important difference of the coarse clastic rocks in rift zones from the molasses in orogenes is in that the former are interlarded with volcanic complex. The volcanic rocks in some areas show bimodal features, consisting of subalkaline or calc-alkaline basalts and rhyolites (including basalts, trachybasalts, trachy-liparite, quartz-trachytes, dacites, etc.). For example, the differential index of the volcanic rocks in the South Hinggan zone shows a typical double-peak pattern, the

basalts belong to continental tholeiite, and the REE europium abnormality of the double-peak volcanic rocks also shows obvious magmatic differentiation. The high weight percentage of iron element (8.33-9.51 %) in the basalts and the low original ratio of strontium isotopes (Sr^{86}/Sr^{87} = 0.7042-0.705) both indicates a deep source of the magma.
The Dahongshan zone is located along the border between the North China ancient continent and the Proterozoic folding belt. There the volcanic rocks are alkali-basalts (igolite basalts, pseudoleucite basalts) accompanied by alkaline syenite, showing that its magma originates from the upper mantle. In the other volcanic rock zones there are also alkaline granites and syenites belonging to A-type and diorite-tonalite-granites and granodiorites belonging to I-type (2). The late Palaeozoic extension also covers the "Inner Mongolia earth's axis" in the North China ancient continent, resulting in a series of alkaline complex bodies (260-220 Ma) along active deep faults.
After the large upwarping from the end of the late Palaeozoic to the beginning of the Mesozoic, another extension from the early to middle Jurassic takes place, which continues until the early Cretaceous, forming the Mesozoic coal-collecting basins and oil-gas-bearing sedimentary basins. The Jurassic system consists of the lower black lacustrine basin, (bog facies, coal measure strata) and the upper continental facies (volcanic lava, tuffaceous rocks, sandstone and conglomerate rocks). Such environments characterize a continental rift with weak volcanic activities. In the Yingen-Baolige and Xi Ujimgin rift zones middle and lower Jurassic coal-collecting graben basins, together with the Mesozoic coal-collecting belts in Mongolia(3) six nearly parallel linear coal-collecting belts may be identified (see the map). In the three belts in Mongolia coal forming started since Palaeozoic and reached their significant coal-collecting stage in the Permian and early Cretaceous. Some prospective power coal deposits have been discovered there. The distribution of all the coal-collecting belts points out that besides the necessary ancient climate conditions an ancient geographic environment suitable for plant remains accumulation is also very important for coal collecting, a graben type setting, i.e. a rift condition would be ideal.
The distribution of the early Cretaceous oil-gas bearing (or oil potential) basins in the Chinese part(4) basically coincides with that of the aforementioned rift zones (see the map). Among the basins the Erlian basin is more thoroughly studied; it is 800 km long, 20-80 km wide, and 5-6 km deep, composed of 35 faulted basins. Those symmetric or non-symmetric faulted basins or steplike fault blocks are important oil-producing and oil-reservoir structures. The sediments in the faulted basins show relatively rapid accumulation rate and sharp change of thickness, and are interlarded with volcanic complexes. The geothermal gradient is rather large (4,9^{o}C/100m) near the faults. Those facts prove that those basins are of rift type instead of intermontane depression type. The latter often relates to orogenic depression. But in the late Cretaceous there is a large scale upwarping accompanied with strong intermediate acid magmatic intrusion in the area concerned.

3. COMPARISON OF FEATURES OF RIFTS IN DIFFERENT STAGES

As mentioned above, the spatial distributions of the rifts in different stages show an obvious characteristic of inheritance, but each of them also possesses its own features as shown in table 1. Studying their differences helps to understand the evolution of the tectonic environment.
The comparison shows that the rift activity in the early stage is stronger than that in the late stage. This phenomenon may be attributed to the fact that the young crust accreting in the middle Palaeozoic has become more consolidated with time, and that the continental crust has become thicker and more stable due to the magmatic activities of the rifts and the deformation-metamorphism of the strata.

4. DYNAMIC MECHANISM OF RIFT ACTIVITIES

It is worthwhile to notice that the break hiatus between the two stages of the rift activities, i.e. the large scale rising in the late Permian-Triassic and late Cretaceous, is not accompanied by any strong faulting or napping. Therefore, the extension regime, which controls the rifts, is the main geodynamic environment in the late Palaeozoic to Mesozoic; and the several rift zones spread between the two large plates, Siberian and Sino-Korean, and the rising denudation zones between them consist of the important structural framework of the area concerned. Such a large scale extension regime can not be explained simply with the late-orogenic stress relaxation or continuation of the residual ophiolite magma reservoirs. The effect of the newly opened late Palaeozoic oceans (ancient Pacific and ancient Tethyan oceans) must be taken into account, especially the transverse compression of the ancient Pacific-Ocean plate may be the main cause of extension activities in the originally structurally weak zones. But it should also be mentioned that such rift activities occuring in post-geosynclinal orogenic zones on the newly accreting young crusts are different from continental rifts (e.g. the East-African rift) occuring on old crystalline basements. Furthermore, the late-period structural features of the kind of orogene described herein are different from those of the kind which keeps a compression setup from beginning to end. If the latter is named Alpine-Himalayan type, the former may well be called North-Asian type. The essential difference between the two is in whether there exists transverse compression from adjacent plates.

Table 1. Comparison of Features of the Rifts in Different Stages

stages	Morphological features of the rifts	Sedimentary facies and thickness	Features of formation	Effective depth	Degree of crustal activity
Mesozoic (J_1-K_1)	composed of various symmetric, non-symmetric basins, forming relatively wide rift belts	continental facies accumulation, 5-6 km	characterized by coal and oil controlled by syntectonic sedimentary basins	depth of sedimentary basins is small	relatively calm
Late Palaeozoic (C_2-P_1)	distributing linearly along large faults, forming narrow and deep rift valley	marine-continent-alternating facies, 10-20 km	characterized by bimodal volcanic rocks and alkaline plutonic rocks controlled by rift activities	channels of magmatic movement reach lower crust and mante	relatively strong

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