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Kazimierz ŻYTKO

THE ATLANTIC, THE INDIAN OCEAN AND MAIN LINEAR FRACTURE ZONES OF THE POST-VARISCAN EUROPE

(3 Figs.)

Atlantyk, Ocean Indyjski i główne linearne pęknięcia powaryscyjskiej Europy

(3 fig.)

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Abstract: In the platform Europe area and in the Alpine zone there exists a Permian — Mesozoic linear rifting belt oriented NW—SE, extending from the North Sea to the Aegean Sea. It is related with the Zagros-Oman system through the deformed Taurides and the Bitlis zone. Its genesis may be connected with the initial rift of the Indian Ocean. Elements of a similar rifting oriented SW—NE in western Europe and Africa are connected with evolution of the Atlantic Ocean. Both rifting systems interfere with a geosynclinal Tethys system oriented W—E, found in the Eurasia — Gondwana suture zone. The transform faults line, running from North Anatolia, through the Southern Carpathians, the Alps, as far as Orlean, is related with the Gibbs fracture zone in the Atlantic. This line separates the Alpine — Mediterranean microplate belt from the rest of Europe. Important elements of the Tertiary rifting, i.e. the Red Sea rift, the Faeroe — Iceland ridge and, possibly, also the Rhine and Saône grabens, were formed on an extension of the Permian-Mesozoic rift elements of the adjoining plates.

Key words: plate tectonics, lineaments, Permian-Mesozoic rifting, Europe platform, Carpathians, Tethys, Atlantic Ocean, Indian Ocean

Kazimierz Żytko: Geological Institute Carpathian Branch, ul. Skrzatów 1, 31-560 Kraków.

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Treść: Na obszarze platformowej Europy i w strefie Alpidów istnieje pas permo-mezozoicznego linearnego ryftingu o kierunku NW—SE sięgający od Morza Północnego po morze Egejskie. Przez zdeformowane Taurydy i strefę Bitlis ma on związek z systemem Zagros—Oman. Jego geneza wiąże się przypuszczalnie z inicjalnym ryftem Oceanu Indyjskiego. Elementy podobnego ryftingu o kierunku SW—NE w zachodniej

Europie i Afryce wiążą się z ewolucją Atlantyku. Oba systemy ryftingu interferują z geosynklinalnymi elementami Tetydy występującymi w strefie szwu Eurazja—Gondwana. Linia transformujących uskoków biegnąca od północnej Anatolii przez Karpaty Południowe, Alpy aż po Orlean ma związek z rozłamem Gibbsa na Atlantyku. Oddziela ona pas alpejsko-medyterrańskich mikropłyt od pozostałej części Europy. Ważne elementy trzeciorzędowego ryftingu — ryft Morza Czerwonego, grzbiet Faeroe (Wyspy Owcze) — Islandia, a przypuszczalnie także grabeny Renu i Saony powstały na przedłużeniu elementów permomezozoicznego ryftu sąsiednich płyt.

INTRODUCTION

On the epi-Paleozoic platform of the western part of the Eurasian plate there occurred intensive, mainly vertical movements during the Alpine orogeny. Tectonic effects of these movements, occasionally termed Saxonian, point to stages of tension and compression within the plate. From the south the epi-Paleozoic platform adjoins an area of the Alpine folding. This area separates the Eurasian plate from the African-Arabian one. Between them a few microplates with a complicated history and pattern can be distinguished.

As regards methods aiming at determination of time, direction and extent of movements of plates and microplates, as well as their rotation, the analysis of transform faults and accreting plate margins appears to be the most useful; the consuming plate margins, due to difficulties in assessing the extent of subduction, appear to be unclear. For these reasons, especially in order to understand evolution of the Alpine — Mediterranean orogenic belt, young linear fracture zones of the crust, passing from the epi-Paleozoic platform into the Alpine system area, are of special importance.

The author discusses the course and time of formation of linear rift trans-European elements running from the North Sea through the Carpathians to the Aegean Sea. At the same time, he refers to Sonder's concept of rhegmatic Iceland — Red Sea line (1938). Elements of similar age and origin, directed SW-NE, can be found in western Europe.

The paper postulates the existence of a trans-European line of transform faults, running from the Charlie-Gibbs fracture zone on the Atlantic Ocean, between the Saône (Bresse) and Rhine grabens, up to Anatolia. North of that line, the Alpine orogens of Europe are represented only by the Eastern Alps and the Carpathians, as well as by the North Dobrogea (Dobrudja) — Crimea — Major Caucasus orogenic belt. South of the fault line there are main European Alpine chains.

A plate margin of a similar importance can be found on the eastern extension of the Azores — Gibraltar fracture zone in the Mediterranean Sea.

RIFT SYSTEM: THE NORTH SEA — THE CARPATHIANS — THE BALKAN PENINSULA — THE AEGEAN SEA

Sonder (1938), partly on the basis of Sieberg's seismic activity analysis, paid attention to the existence of a linear structure system ("zonale") of Iceland — the Red Sea, directed NW-SE, According to Sonder, this lineament which begins with the Red Sea graben can be found in structural features of the Aegean Sea and in the Balkan Peninsula. It runs between the Alps and the Carpathians towards the North Sea, then through Iceland and Greenland as far as the Baffin Bay.

North of the Aegean Sea, as far as the Belgrade area, the lineament is represented by the Serbo-Macedonian massif and its marginal elements: the Vardar and Kraishtide zones (Bončev, 1974a; Sikošek 1974), These linear elements are not recognized in the Pannonian basin region; therefore a few variants of a further course of the mega-lineament — Sonder's diagonal — can be considered.

LINEAMENT: EGERSUND BASIN — DANISH-POLISH TROUGH — EASTERN CARPATHIAN TROUGH — KRAISHTIDE TROUGH

The Norwegian—Danish—Polish segment. The Danish-Polish trough (furrow) is a distinct linear element of Europe's stable epi-Variscan platform (Fig. 1). It is a Permian—Mesozoic synsedimentary trough, directed NW-SE, which has a variable history of different segments as regards their details. The trough begins in the region of the meridional Viking — Central Graben system of the North Sea and is known as the Egersund Basin, localized between Ringkobing — Fyn High and Bergen High — the Fennoscandian Shield (Brennand 1975; Childs, Reed 1975; Ziegler 1978). The trough continues from northern Denmark (Danish Embayement) trough the Copenhagen area and the south-western Baltic Sea. Further on the trough crosses the Polish lowland. It reaches diagonally, under Neogene molassic deposits, the margin of the Carpathian orogenic belt (Fig. 2) between the Tarnów region (Poland) and the area situate SE of Stryj (the Soviet Union).

Senonian and Laramide inversional movements, as well as an Upper Cretaceous volcanism (Scania) are reported from the Danish-Polish trough (Ziegler 1978). The south-eastern, Polish part of that trough, referred to as the mid-Polish aulacogen (Pożaryski, Brochwicz-Lewiński 1979), underwent a visible inversion at the end of Cretaceous. The axial part of the furrow was transformed into a Laramide anticlinorium. In the south of Poland this axial zone of the Permian — Mesozoic trough, transformed into the anticlinorium uplifted by ca 3 km, is now represented by the Holy Cross Mts block and the Upper pre-Cambrian (Ry-

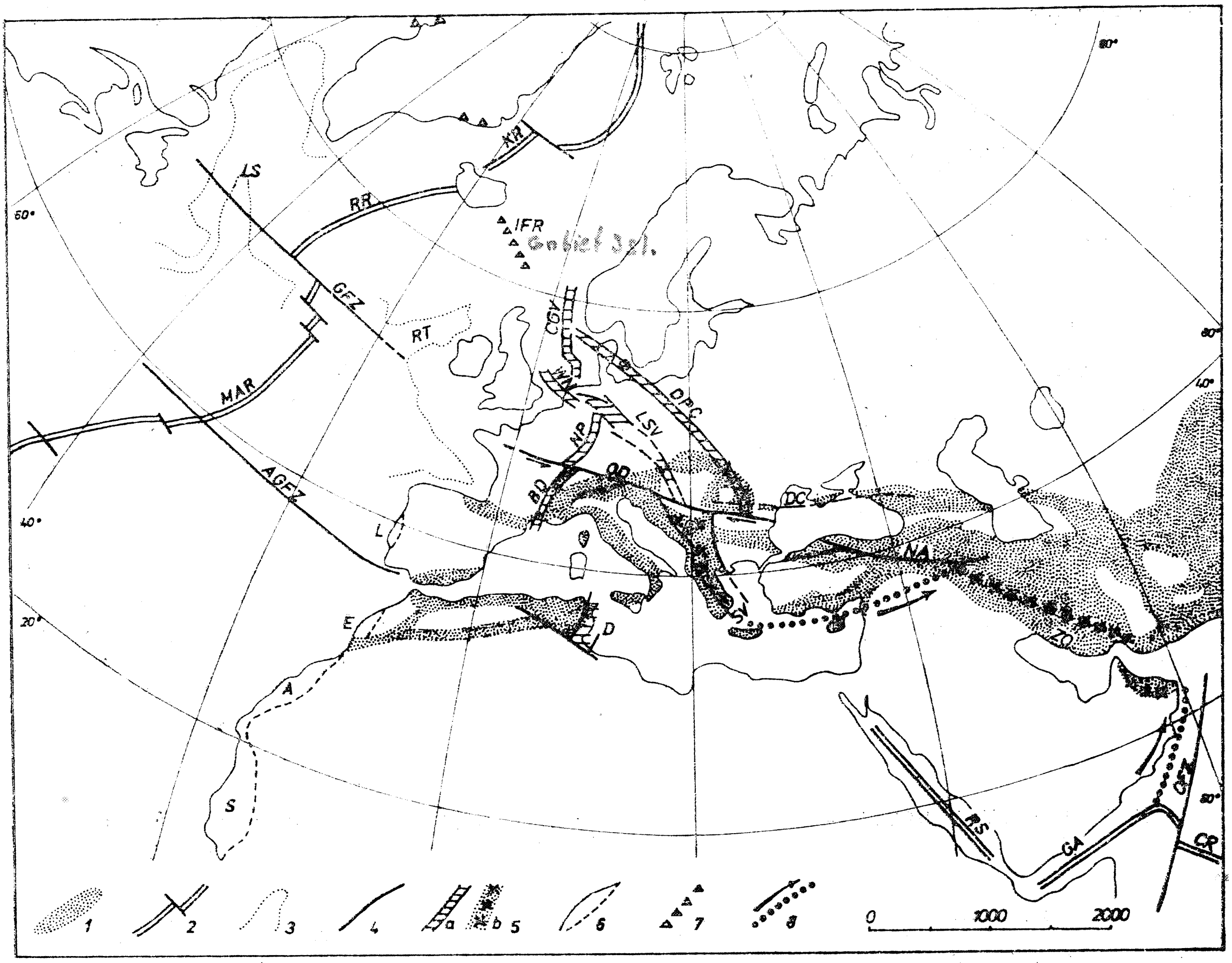

Fig. 1. A sketch of post-Variscan lineaments of Europe, presented in the text, and their relation towards the Atlantic and the Indian Ocean. 1 — Alpine fold chains; 2 — contemporary accreting plate margins (KR — Kolbeinsey ridge, RR — Reykjanes ridge, MAR — mid-Atlantic ridge, RS — Red Sea, GA — Gulf of Aden, CR — Carlsberg ridge); 3 — failed North Atlantic rifts (LS — Labrador Sea, RT — Rockall trough); 4 — transform plate margins (GFZ — Gibbs (Charlie) fracture zone, AGFZ — Azores—Gibraltar fracture zone, OD — Orlean—Zurich—South Dobrogea fault line, DC — northern margin of the Black Sea plate, NA — North Anatolia fracture zone, OFZ — Owen fracture zone); 5 — linear, Permian-Mesozoic rift zones: a) established on the grounds of an anomalously heavy subsidence (CGV — Central and Viking grabens, DPC — Danish-Polish trough and Eastern Carpathians trough, LSV — Lower Saxony-Vienna, Bakony Mts and Mecsek Mts basins, WN — Sole Pit-West Netherlands basin, NP — Nancy—Pirmasens—Hesse trough, BD — Burgundian trough-Dauphiné basin, D — Djeffara trough); b) established additionally on the grounds of presence of Upper Jurassic ophiolites (SV — Ophiolitic-Serbian, Vardar zones in the Dinarides-Hellenides) and Upper Cretaceous ones (ZO — Zagros—Oman zone); 6 — peri-oceanic Mesozoic basins (L — Lusitanian, E — Essaouira, A — Aaiun, S — Senegal); 7 — indications of Cenozoic volcanism on extension of NW—SE — oriented European rifting belt (IFR — Iceland—— Faeroe ridge); 8 — lines showing a post-Liassic dislocation of the Zagros—Oman segment belonging to the linear Indian Ocean rift system, under the influence of Atlantic Ocean opening

Fig. 1. Szkic referowanych w tekście powaryscyjskich lineamentów Europy i ich stosunek do Atlantyku i Oceanu Indyjskiego. 1 — alpejskie pasy fałdowe; 2 — współczesne akrecyjne granice płyt (KR — grzbiet Kolbeinsey, RR — grzbiet Reykjanes, MAR —

phean) San massif (Kutek, Głazek 1972). An Early Tertiary erosion removed Mesozoic sediments and a considerable part of Paleozoic sediments from the above mentioned areas of the mid-Polish anticlinorium. The San massif occurs now within that part of platform which had been reinverted and covered with Miocene molassic deposits of the Carpathian foredeep (Fig. 2).

Borehole data indicate that the San massif is bordered on either side by Mesozoic synclinoria and forms a symmetric bilateral tectogene. From the south-west it is the Miechów synclinorium, while from the north- east- the Lublin — Lvov synclinorium. The total thickness of Mesozoic sediments of both synclinoria diminishes outwards, from 5000—3000 m to a few hundred metres (Kutek, Głazek ibid., Figs. 5, 6). These sediments are developed in carbonate or clastic platformal facies. A zone of increased subsidence, forming the trough, is ca. 150 km wide in the Carpathian foreland, whereas the width of the Upper pre-Cambrian San massif (anticlinorium) is ca. 70 km. The internal zones of the above indicated synclinoria, situated close to the Upper pre-Cambrian axial massif, were folded at the end of Cretaceous (Jaroszewski 1972; Stupnicka 1972; Pożaryski 1977). The intensity of deformations gradually decreases outwards of the tectogene axis.

Characteristics of the Danish-Polish lineament in the Carpathian foreland have been presented, since it seems that a similar tectogene may exist farther southwards, on the other side of the Carpathians.

The Eastern Carpathian segment. Pożaryski and Żytko (1981) remarked that in contact zones between margins of the mid-Polish aulacogen and the Carpathian orogenic chain (Jasło — Krosno region in Poland, Nadvornaja — Kosmach region in the Soviet Union) the

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grzbiet śródatlantycki, RS — Morze Czerwone, GA — Zatoka Aden, CR — grzbiet Carlsberg); 3 — zamarłe ryfty północnego Atlantyku (LS — Morze Labradorskie, RT — rów Rockall); 4 — transformujące granice płyt (GFZ — rozłam Gibbsa-Charliego, AGFZ — rozłam Azorów-Gibraltaru, OD — linia uskoków Orlean—Zurych—Dobrudża Południowa, DC — północna granica płyty Czarnego Morza, NA — rozłam północnej Anatolii, OFZ — rozłam Owen); 5 — linearne, permomezozoiczne strefy ryftowe: a) wyznaczone anomalnie dużą subsydencją (CGV — grabeny Centralny i Viking, DPC — rów duńsko-polski i rów Wschodnich Karpat, LSV — baseny Dolnej Saksonii, Wiednia, gór Bakony i Mecsek, WN — basen Sole Pit — Zachodnia Holandia, NP — niecka Nancy--Pirmasens-Hesja, BD — rów burgundzki — niecka Dauphiné, D — rów Dżeffary); b) wyznaczone ponadto obecnością ofiolitów górnojurajskich (SV — strefy „ofiolitowa" — Serbii i Vardaru w Dynarydach oraz ich przedłużenie w Hellenidach) i górnokredowych (ZO — strefa Zagros-Oman); 6 — perioceaniczne baseny mezozoiczne (L — Lusitanian, E — Essaouira, A — Aaiun, S — Senegal); 7 — przejawy wulkanizmu kenozoicznego na przedłużeniu pasa riftingu NW — SE Europy (IFR — grzbiet Islandia — Wyspy Owcze; 8 — linie obrazujące poliasowe przemieszczenie segmentu Zagros-Oman, należącego do linearnego systemu ryftu Oceanu Indyjskiego, pod wpływem otwarcia Atlantyku

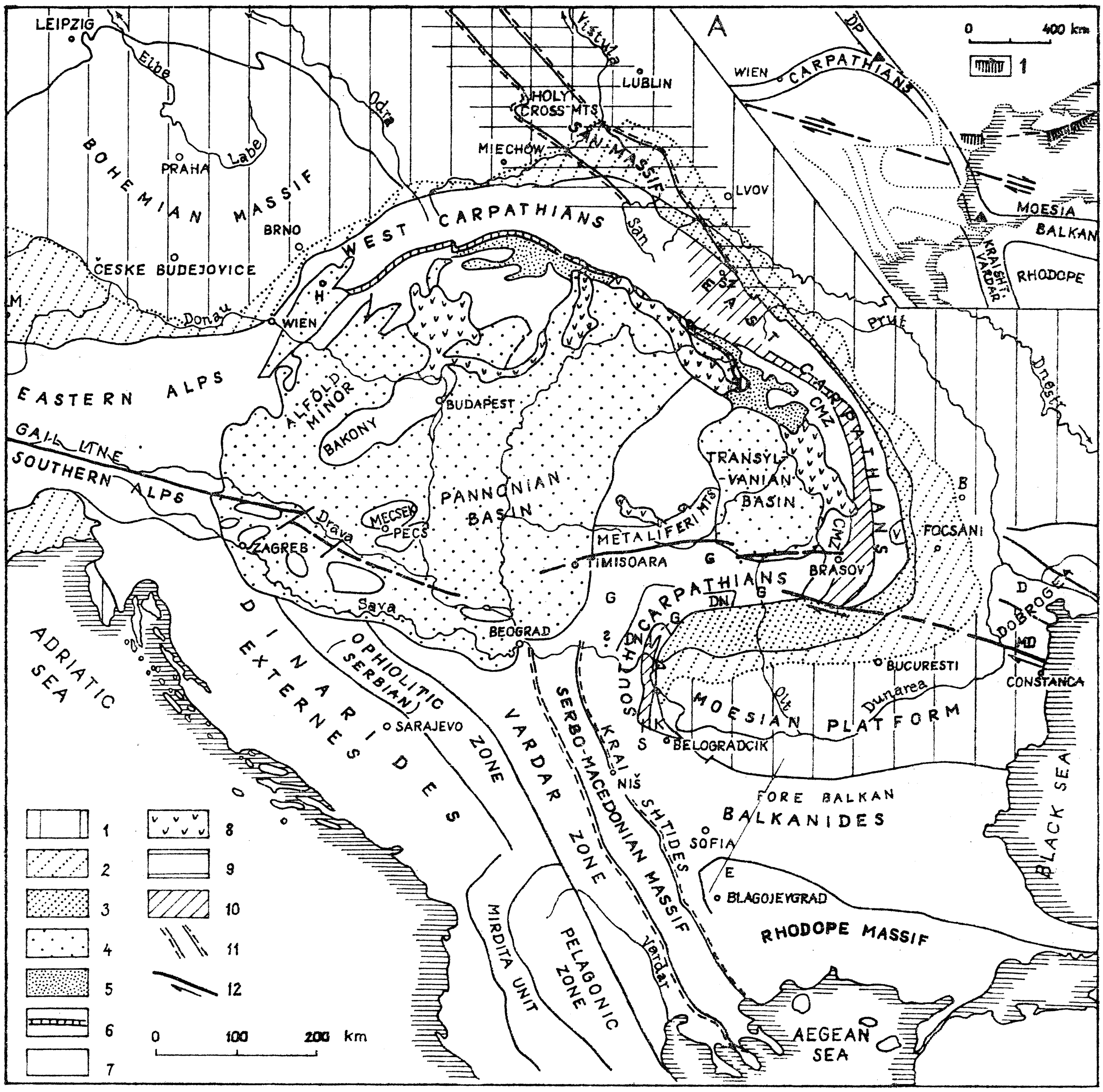

Fig. 2. A sketch of the Carpathians and adjoining areas (according to Mahel ed. 1973, 1974 and other papers cited in this text). 1 — Alpine foreland platforms; 2—4 — Neogene molasses (2 — external foredeep zone, 3 — internal foredeep zone, 4 — main intramontane basins); 5 — postorogenic flysch cover, mainly Paleogene; 6 — Pieniny Klippen Belt; 7 — Alpine folding area; 8 — Late Alpine (N—Q) volcanic rocks; 9 — area of anomalous Mesozoic subsidence of the Danish-Polish furrow; 10 — folded deposits of the tensional Eastern and Southern Carpathians trough (Shevchenkovo zone-Cretaceous; Sinaia-Palanca trough deposits-Tithonian-Lower Cretaceous); 11 — axial massifs with a similar geostructural position in relation to the Carpathian geosyncline; 12 — main fractures and the direction of dislocations. M — Munich, H — Hodonin, Sz — Shevchenkovo, B — Birlad, CMZ — crystalline-Mesozoic zone, V — Vrancea area, G — Getic unit, DN — Danubian autochthon, S — Severin-Kraina unit, K — Kula complex, MD — Central Dobrogea massif, D — North Dobrogea orogen, E — Entropolé line. Inset A. An outline of the Carpathian-Balkan sigmoide and its relation to the linear NW-SE — oriented rift before Upper Cretaceous deformation. 1 — Early Mesozoic North Dobrogea-Crimea orogenic belt. Triangles denote areas of triple conjunction-virgation. Dotted lines show present distribution of structural and geographic elements

course of the negative regional gravity anomaly of the Carpathians is interrupted. The Krosno — Kosmach section of the anomaly is shifted about 8—10 km to the north-east. This testifies to a distinct relation between the platformal linear element with a tensional origin and the structural pattern of the Carpathian basement. Therefore, the authors mentioned above have assumed that the Shevchenkovo zone of the anomalously thick Cretaceous flysch of the Skole — Tarcau unit in the external part of the Northern Carpathians and, farther southwards, the Tithonian — Lower Cretaceous Sinaia — Palanca trough, constitute an extension of the Danish — Polish trough (aulacogen) (Fig. 2). A basic magmatism is connected with the trough limits. Also the Middle Cretaceous, Senonian and Laramide movements were observed in that trough. The trough deposits gave rise to the following units: Rahov — Porkulec — Ceahlau — Teleajen of the Eastern Carpathians and Severin — Kraina unit of the western part of the Southern Carpathians (vide Mahel ed. 1973; Sandulescu 1975). The term "trough" is used here in a broader sense then that denoting the Dacide trough ("megasillon"), according to Sandulescu (1975). It comprises not only the area of Tithonian — Lover Cretaceous deposits of the Ceahlau unit but also a zone of rapid sedimentation of the Aptian — Albian — Vraconian "curbicortical" flysch (Palanca formation). The Teleajen unit, included by Sandulescu tc the Moldavides, is made up at this flysch.

In the Northern Carpathians, in the San-Prut interfluve, there occurred a Cretaceous virgation (Fig. 2 A). In the linear rift pattern, the platform Danish — Polish trough was an extension of the Sinaia — Palanca trough. In the geosynclinal pattern — such extension was represented by the flysch furrow of the Silesian and sub-Silesian units of the Western Carpathians.

Fig. 2. Szkic Karpat i obszarów przyległych (na podstawie Mahel ed. 1973, 1974 i innych prac cytowanych w tekście). 1 — platformy przedpola alpidów; 2—4 — neogeńskie molasy (2 — zewnętrzna strefa zapadliska, 3 — wewnętrzna strefa zapadliska, 4 — główne baseny śródgórskie); 5 — postorogeniczna pokrywa fliszowa, głównie paleogen; 6 — pieniński pas skałkowy; 7 — obszary fałdowań alpejskich; 8 — młodoalpejskie (N—Q) skały wulkaniczne; 9 — obszar anomalnej mezozoicznej subsydencji bruzdy duńsko-polskiej; 10 — sfałdowane osady tensyjnego rowu Karpat Wschodnich i Południowych (strefa Szewczenkowa — kreda; osady rowu Sinaia-Palanca — tyton-kreda dolna); 11 — osiowe masywy o podobnej pozycji geostrukturalnej w stosunku do geosynkliny Karpat; 12 — główne pęknięcia i kierunek przesunięć. M — Monachium, H — Hodonin, Sz — Szewczenkowo, B — Birlad, CMZ — strefa krystaliczno-mezozoiczna, V — rejon Vrancea, G — jednostka getycka, DN — autochton Dunaju, S — jednostka Severin-Kraina, K — flisz Kula, MD — masyw Centralnej Dobrudży, D — orogen Północnej Dobrudży, linia Entropolé. Fig. 2A. Zarys karpato-bałkańskiej sigmoidy i jej stosunek do linearnego ryftu NW-SE przed górnokredową deformacją. 1 — wczesnomezozoiczne pasmo Północna Dobrudża-Krym. Trójkąty wyznaczają obszary potrójnych węzłów-wirgacji. Linie kropkowane pokazują wczesny układ elementów strukturalnych i geograficznych

A conjunction of the Eastern Carpathian and Balkan — Aegean segments. The Danish — Polish trough and the Sinaia — Palanca one of the Eastern Carpathians determine a Mesozoic lineament extending for ca. 2000 km (Fig. 1). It ends abruptly in the Dimbovița river valley NW of Bucharest. However, a close connection between the Tithonian-Neocomian flysch of the Ceahlau unit (mainly the Sinaia formation) and of the Severin — Kraina unit mentioned above (Sandulescu 1975) seems doubtless, despite the fact that their main occurrences are ca. 200 km apart along the W-E line in the Southern Carpathian front (Fig. 2).

The main unit of the W-E oriented section of the Southern Carpathians is a Cretaceous Getic basement — shear nappe (Sandulescu, Nastaseanu, Kräutner 1974; Sandulescu, ibid.). In the north it is limited by deep fractures along the Timisoara-Brașov line, which separate it from the Metaliferi Mts zone and the basement of the Neogene Transilvanian basin. In the east, in the Brașov area, a pre-Albian or intra-Albian overthrust of the Getic nappe upon the Ceahlau flysch unit was found. West of the Olt river a southerly overthrust of the Getic unit is of an intra-Senonian or Laramide character. This unit is thrust upon the Danubian autochthon deformed by Alpine movements. The relation between this autochthon and the Moesian platform is not clear. In the west, in the Danube gate area, between the Getic nappe and Danubian autochthon, there occurs the Severin-Kraina cover decollement nappe mentioned above (Fig. 2).

The picture of an external boundary of the Southern Carpathians (Sandulescu, Nastaseanu, Kräutner, ibid.) points to the existence of a transcurrent, dextral fault between the Moesian platform and the W-E oriented part of the Carpathians (between the Danubian autochthon and Getic unit?). Various authors (Dewey et al. 1973; Birkenmajer 1976; Biju — Duval et al. 1977; Pożaryski, Żytko 1981) accepted such a fault as a result of a Cretaceous drift of the Moesian microplate, and assumed that it was of a transform fault character. The fault zone was deformed by thrusting movements in the Tertiary. The microplate drift split up the Sinaia — Palanca trough (lineament) and shifted the southern segment far westwards where part of its deposits have formed the Severin - Kraina unit (Fig. 2) .

North of Belogradcik, in the Danube gate area, outwards of the Tithonian — Neocomian Severin — Kraina flysch, there occurs a folded Upper Cretaceous flysch (the Kula complex); it is discordantly overlain by the Middle Eocene Staropatica formation (Cankov 1974). The Kula zone flysch is probably a relic of shifted and almost completely consumed external units of the Eastern Carpathians (the Moldavides). Therefore the moesian drift took place mainly at the end of Cretaceous and was very rapid. The latter phenomenon is also evidenced by the development of magmatism of Laramide banatites related, according to some authors, with the process of subduction (Radulescu, Sandulescu 1973). If a westerly inflection of the southern part of the Eastern Carpathian lineament is considered, the length of the drift should be estimated at more than 400 km (Fig. 2).

The pressure of the Moesian microplate, of the fore-Balkan area and of the present Balkanides flexibly related with it (vide Bončev 1974; Karagjuleva, Cankov 1974) resulted in a potent tectonic deformation and partial consumption of the transit area between the Carpathians and the Balkanides. In the internal zone of that area (the Banat Mts, East Serbian Mts) there can be distinguished narrow structural zones included with the Carpathian — Balkanide chain (Grubič 1974). At this point other authors report an independent, linear (NNW-SSE) orogenic belt of the Kraishtides which, in the west, comes into contact with the Serbo — Macedonian massif (Bončev 1974a; Karagjuleva et al. 1974).

The meridional Tithonian — Neocomian flysch belt of the Severin — Kraina unit (except for the Kula zone), ca. 150 km long, makes up an external, eastern zone of the tectonically deformed transit area mentioned above. This flysch ends NW of Belogradcik (Fig. 2). It remains an open question whether the latter flysch only is an extension of the Sinaia — Palanca trough deposits, or what formation makes a further extension of the rift trough southwards. Pożaryski and Żytko (1981) considered the Kraishtide flysch trough to be such an extension; however, this problem requires further elucidation.

In the geosynclinal pattern, going eastwards along the northern Alpine belt, the Tithonian — Neocomian flysch and flysch-like deposits are found in external parts of the Balkanides (fore-Balkan, Stara Planina), east of the Entropolé line (Bončev 1974; Karagjuleva, Cankov 1974). However, another flysch trough existed in that area in Tithonian — Valanginian time. It is assumed that the latter trough had a trend of 140—150°. Its deposits are found in the south-western part of the Kraishtides (Lužnica zone, Anina Ruj zone), south of Niš (Grubič 1974; Bončev 1974a; Karagjuleva et al. 1974). The Kraishtide Mesozoic deposits are preserved in the south only as far as the Blagojevgrad region (Fig. 2), but structural elements of that orogenic belt (Strimon line) continue up to the Aegean Sea. Therefore, in the regional Alpine pattern another virgation can be observed — a southerly-south-easterly divergence of the taphrogeosynclinal Kraishtide trough from the Carpathian — Balkan belt Bončev 1977). The latter virgation shows a great analogy to a virgation from the Northern Carpathians. Therefore, before the Moesian Laramide drift, the Carpathian — Balkan geosyncline had an outline of an inverted sigmoide (Fig. 2A), whereas after that drift it assumed the shape of the present arc.

South of the deformed sigmoide there is a Serbo-Macedonian massif, bordered from the west by a synclinorial Vardar zone with Upper Jurassic — Cretaceous flysch and ophiolites (Sikošek 1974), whereas from the east — by the Kraishtide zone mentioned above (Vardaride — Kraishtide lineament, Bončew 1974a). Upon the Serbo-Macedonian massif there are remnants of Upper Senonian shallow-water deposits (Dimitrievič 1974). In relation to the pre-Laramide sigmoidal section of the geosyncline the whole bilateral, symmetrical tectogene assumes the analogous position as the above-mentioned Laramide San — Holy Cross Mts tectogene in the north (Fig. 2). The width of main elements of both tectogenes, the direction of changes in the thickness of Mesozoic deposits in the Kraishtides and the Lublin — Lvov synclinorium, as well as their general structural features are similar. Paroxysmal tectonic movements in both tectogenes occurred at the end of Cretaceous time. Essential dissimilarities lie in the lack of Mesozoic magmatism and flysch facies in the San platformal tectogene.

The analogous geostructural position of both tectogenes, their similarities, as well as the fact that the Holy Cross Mts and the San massif can be regarded as an area with a considerable Mesozoic subsidence and Laramide inversion (Kutek, Głazek 1972), point to the possibility that the Tithonian-Neocomian flysch of the Vardar and Kraishtide zones was formed in the same NW-SE oriented basin; this flysch was later divided into two zones by a young pre-Campanian uplift of the Serbo-Macedonian axial massif and the removal of cover deposits from it. Similar type of evolution of Mesozoic marginal basins was observed in north-western Europe (Voigt 1963).

If the fore-going considerations are true, in the pre-Upper Senonian pattern the Bončev's Vardaride-Kraishtide lineament is an equivalent of the Danish — Polish — Eastern Carpathian lineament. Both lineaments are now separated from each other by the transform fault of the Southern Carpathian front. NW-SE oriented structural features of the Vardaride — Kraishtide lineament can be observed in the Aegean Sea as far as the Andros — Samos islands line (Kronberg, Günther 1978). However, the relation between the Severin — Kraina flysch and that of the Vardaride — Kraishtide trough remains unclear. The two flysch zones do not join on the surface. The existence of these two flysch zones is probably connected with the previously mentioned virgation of troughs, south of the transform fault (Fig. 2A).

The transform fault in question was an important structural boundary as early as in the Tithonian — Neocomian, and probably even much earlier. It is found in the contact zone of the W-E oriented part of the southern Carpathians and the Moesian platform. There exists an enormous shift of the southern area westwards. There are also significant tectonic deformations in the prolongation of the transform fault here discussed in the southern part of the Pannonian basin in the Sava-Drava interfluve (horst and graben zone, vide Mahel ed. 1973; Sikošek 1974). It is likely, therefore, that the fault traverses the meridional part of the Southern Carpathians. This rises again the question of the southword extension of the Getic nappe. On both sides of the transform fault here discussed, the paleogeographic patterns might have evolved in different manner. In the light of such a complex situation, the extension of the Sinaia — Palanca trough deposits of the Eastern Carpathians should be seen not only in the Severin — Kraina flysch (which is doubtless) but also in the flysch of the previously postulated Vardaride — Kraishtide trough or of its part. This trough was the first tensional trough west of the cratonic Rhodope massif.

The crystalline — Mesozoic zone of the Eastern Carpathians (Fig. 2) is not an extension of the San and Serbo-Macedonian tectogenes. A diversified pre-Middle Cretaceous cover of Mesozoic sediments indicates that that zone is not an axial zone of the geosynclinal trough; also the paroxysmal phase of deformation is in these zone of an earlier, Middle Cretaceous age.

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The lineament discussed above (Danish-Polish trough — Eastern Carpathian trough — Vardaride — Kraishtide trough) was a rift zone of an anomalously high subsidence; its activity changed in a time — and space. In Permian — Mesozoic time that zone extended from the North Sea to the Aegean Sea both in the regions of rigid cratons and in the geosyncline. At the turn of Cretaceous — Paleogene times the zone underwent paroxysmal tectonic movements. It was subject to on inversion in cratonic areas (except for the region NW of the Baltic Sea), and displacement along the Southern Carpathian front. The Senonian and post -Cretaceous movements were connected with the North Atlantic opening (sea-floor spreading of the Labrador Sea), and with the change of direction of Africa's motion in relation to Europe (Dewey et al. 1973). The fact that the inversion took place only in cratonic areas points to a "hydraulic" subcrustal uplift genesis of axial massifs; at the same time impulses came directly from the geosyncline area.

LINEAMENT: LOWER SAXONY BASIN — VIENNA BASIN — BAKONY MTS — MECSEK MTS

T h e C e n t r a l E u r o p e a n p l a t f o r m. Let us return to Sonder's primary idea concerning the northern extension of the Vardaride-Kraishtide lineament. In the Permian-Mesozoic North Sea rift system, the Viking Graben passes southwards into the Central Graben, oriented NW-SE (vide Childs, Reed 1975; Ziegler 1978). In the southern part, the latter graben is oriented submeridionally. An intracratonic Lower Saxony tectogene (Voigt 1963; Boigk 1968), oriented WNW-ESE, diverges from the North Sea rift system. The tectogene is limited by the Osning zone (Teutoburger Wald) in the south-west and by the Weser — Ems line and the Aller line in the north-east. The tectogene was formed as a result of an intra-Senonian — Laramide inversion and tectonization of the Permian — Mesozoic basin of anomalously high, variable subsidence. Jointly with the synclinoria bordering the axial Harz massif, it determines the NW-SE oriented Lower Saxony — Thuringia lineament in the Permian — Mesozoic pattern. This lineament resembles the Polish part of the eastern lineament described above. Subparallel axial zones of both troughs are ca. 400 km apart; however, they join in the north in the North Sea rift system (Fig. 1).

On the extension of the lineament "frames" (Thuringer Wald, Aller line, Flechtingen Scholle) there are young or rejuvenated Saxonian fractures of the stable Bohemian massif (grabens of the České Budějovice region, the Labe line, the Lusatian fault and others), oriented NW-SE (Zoubek, Malkowski 1974). In the region of Vienna and Brno these fractures reach the western Carpathian foredeep. However, the western part of the Carpathians joining the Alps is covered with Neogene molassic deposits of the internal Vienna basin (Beck-Mannagetta 1974; Prey 1974; Mahel and Vass 1974; Roth 1974). Over 2000 m thick autochthonous Jurassic deposits were found on the south-eastern slope of the Bohemian massif, under Neogene molasses of the foredeep between the Danube and Hodonin. Farther in the west, also Lower Cretaceous deposits were observed. The above-mentioned deposits, together with autochthonous Paleogene sediments filling the submarine canyons of Nesvacilka, Vranovice (Picha 1978), point to lability of that zone and its predisposition to subsidence.

The Pannonian basin. A labile zone, delineated by thick Neogene deposits of the internal Vienna basin, traverses the orogenic Alps — Carpathians belt and joins the Alföld Minor (Fig. 2). Hungarian geologists have gathered a number of data concerning the structure of pre-Neogene basement within the Carpathian arc in the Pannonian depression. A few belts of Mesozoic deposits, mainly SW-NE oriented have been observed (vide Balogh, Körösy 1974; Dank, Bodzay, Körösy in Lemoine ed. 1978, Figs. 9.42, 9.43). Following the direction of Sonder's lineament, elements elevated in the Neogene structural pattern are found to occur SE of the Vienna basin. They are represented by the Bakony Mts, the Mecsek Mts and the Villany area SE of Pecs (Fig. 2). The Mesozoic deposits in the Bakony Mts are 6000 m thick (in the Mecsek Mts — 4550 m).

There are data indicating that the previously mentioned elements were segments subject to an exceptional subsidence. This can be evidenced, e.g., by gradual passages from Triassic deposits into Liassic ones, and from Tithonian into Valanginian-Barremian ones in the Bakony Mts; at the same time discordances of that age were observed in the adjoining regions of Vertes and Gerecse (W of Budapest) belonging to the same belt. In the Upper Triassic, detrital delta and lacustrine sediments were deposited in the Mecsek bay; a passage into Jurassic deposits is found there, as well. In the Early Cretaceous, an alkalic — basic volcanism occurred in Mecsek; its traces can be also observed in the Villany area. Despite insufficient investigations of the pre-Neogene basement, it can be assumed that an extension of the Mesozoic rift lineament of the Lower Saxony — Vienna basin can be seen in the Pannonian depression as far as Mecsek — Villany interfering with elements of the Western Carpathians structural pattern oriented SW-NE. Thus the north-European segment of Sonder's lineament can also be traced in the western variant, as well.

Farther southwards, between the Drava and Sava rivers, there occurs a zone of horsts and grabens, mentioned elsewhere in this paper, which is an extension of the transform fault of the Southern Carpathian front. At present the Vardar zone lies on an extension of the Vienna basin — Mecsek line (Fig. 2). Since the area south of the fault was shifted westwards at the end of Cretaceous, it is likely that the main Ophiolitic Dinaride zone, or the Serbian zone, occurred in the Mesozoic pattern, opposite the rift element discussed elsewhere. The Ophiolitic zone joins the Hellenide sub-Pelagonic zone through the Mirdita zone. A great number of late Jurassic basic and ultrabasic rocks is characteristic of this eugeosynclinal belt (vide Mahel ed. 1973; Sikošek 1974; Karamata 1974).

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Subsident, Permian — Mesozoic Danish-Polish and Lower Saxony — Vienna zones frame a labile, NW-SE — oriented belt, 500—600 km wide and expanding southwards (Fig. 1). This belt belongs to the platformal Europe area, comprised within the rift-shear system (Illies 1974). In the north that belt also includes a similar zone of anomalous subsidence: Sole Pit basin — West Netherlands basin (vide Ziegler 1978). Faults and fractures of the belt continue in the Carpathians and in the Pannonian area. Their direction agrees with the prevailing direction of orogenic belts in the Alpine — Mediterranean microplate zone. The labile rift belt of Central Europe seems to be an extension of the Kraishtide and ophiolitic Dinaride zones (Vardar zone, Ophiolitic-Serbian zone) because of its long-lasting (until Middle or Late Cretaceous) tensional character. That southern area is considerably narrower — from 250 to 300 km (Fig. 1). The narrowing results from Late Alpine folding and thrusting movements.

THE TERTIARY EXPANSION OF THE RIFT ZONE

The lineament segments described above refer to the Permian-Mesozoic pattern. In some parts of these segments (e.g. the Carpatian foredeep, the internal Vienna basin) an anomalously high subsidence occurred in the Tertiary. In a far extension of the labile, NW-SE oriented rift belt of Central Europe in the Reykjanes ridge sea-floor spreading zone, there is volcanic Iceland area where igneous processes are still active (Fig. 1). As a result of this activity in the Tertiary an anomalous Iceland-Faeroe volcanic ridge was formed gradually at the bottom of the open ocean. Tertiary basalts found at Scoresby Sund in East Greenland and on coast of the Baffin Bay (Noe-Nygaard 1974) may be also connected with the NW-SE oriented fracture system.

On a small scale a similar phenomenon may be observed on the Moesian platform and in the Balkanides. There is a meridional zone (25°30'E) of Neogene basalt ocurrences connected with deep fractures and the volcanism of the Eastern Carpathians (Radulescu 1969). As it has been mentioned elsewhere in this paper, the Balkan Kraishtide-Vardaride lineament can be observed in the Aegean Sea as far as the Andros — Samos line. Farther in the south, already on the African-Arabian plate, a Tertiary Red Sea rift system makes an extension of that lineament.

Therefore, outside, on adjoining plates the linear rifting belt was supplemented in the Tertiary by a segment of young volcanics: Faeroe-Iceland-Baffin Bay in the north and the Red Sea rift system in the south.

TRANSFORM FAULTS LINE: THE ATLANTIC (GIBBS FRACTURE ZONE) — THE ALPS — THE SOUTHERN CARPATHIANS — NORTHERN ANATOLIA

The first part of the present paper stressed an important role of the discontinuity zone in the course of a linear structural element oriented NW-SE, determined by deposits of the Sinaia-Palanca trough in the Eastern Carpathians (Ceahlau unit) and in the Southern Carpathians (Severin-Kraina unit). This zone was regarded as a dextral transform fault of the Southern Carpathian front, making a northern border of the Moesian plate. In order to verify that concept, zones of a potential extension of the fault, both westwards and eastwards, were analysed. That procedure was inspired by a possible connection between the fault and an important European discontinuity — the Basel discontinuity — between two Tertiary troughs: the Rhine graben and the Saône (Bresse) graben. The analysis was undertaken despite some difficulties, mentioned elsewhere in the paper, in conducting the fault from the Carpathian foredeep through the Getic unit area east of Belgrade to the Pannonian depression.

The Drava line in the Pannonian depression — the Pusteria — Gail line in the Alps. In the Drava and Sava interfluve, between Zagreb and Novy Sad — Belgrade, there occurs the already mentioned, complicated zone of horsts and grabens. Its structural image is revealed, to a considerable degree, by the thickness distribution of Neogene deposits (Pletikapic in Lemoine ed. 1978, Fig. 9.46). A tectonic connection between the pre-Neogene pattern of this zone and the Dinarides has been stressed (Sikošek 1974). Since the pre-Neogene basement of the Pannonian depression north of the Drava river has entirely different predominating structural directions (SW-NE) than the Dinarides (NW-SE), an important structural boundary is represented by the Drava line. It is oriented WNW-ESE and occurs approximately on an extension of the northern border of the Moesian plate (Fig. 2).

The Drava structural line passes westwards into a big fracture, described as the Pusteria or Gail line. The fracture makes a tectonic boundary between the Southern and Eastern Alps. Differences between these segments of the Alps chain have been known and described several times (vide Rutten 1969; Lemoine ed. 1978). The genesis of the Gail line, also known as the "root zone" or the "peri-Adriatic line" has been given several interpretations. However, Van Bemmelen's opinion should be quoted here (fide Rutten 1969, p. 328) that the boundary between the Eastern and Southern Alps is a great dextral transcurrent fault.

The block south of this fault is shifted westward of about 80—100 km as it is seen from the present distribution of Pb-Zn ore deposits in the Triassic on both sides of the Gail-peri-Adriatic line (Brigo et al. 1977).

The Gail line is oriented WNW-ESE. North of Bolzano it joins a sinistral Judicaria fault, directed SW-NE. On the northern extension of the fault there is a western boundary of the Tauern window. It has been commonly accepted that the Gail line passes into the Judicaria fault, from which a structural Insubric line diverges; the latter line continues as far as the Po lowland, being the northern boundary of the Southern Aips.

Irrespective of this continuity which has been commonly accepted, it should be noted that on a rectilinear extension of the Gail line westwards there runs a boundary between the Eastern and Central Alps. Despite the continuity of tectonic units of both chain segments, the differences between them are very significant. External crystalline massifs, characteristic of the Western and Central Alps, are absent in the Eastern Alps. The Northern Calcareous Alps (Mesozoic of the Upper Austro-Alpine) end abruptly south of Bodensee.

Trümpy (1975) pointed to the presence of fault scarps, also including a transform fault (Dietrich's idea), in the sedimentation area of the Liassic, Dogger and Malm in the Pennine and Lower Austro-Alpine zones in the Graubünden region (the borderland between the Central and Eastern Alps). Later orogenic movements deformed the primary pattern of the geosyncline deposits. In the light of the quoted differences in structure and basing upon sedimentological observations it may be assumed that the boundary between the Central and Eastern Alps has the same structural predisposition as the Gail line. Here runs probably a transform fault zone (plate boundary) which goes beyond the Alpine geosyncline traversing its outer area. In the platform area the northern extent of the Jura chain may be connected with the structural line discussed (Fig. 1).

The Montbard — Zurich line in the Tertiary rift zone. The area between the northern end of the Saône (Bresse) graben and the Jura chain, and the southern extremity of the Rhine graben belongs to the most tectonized and well-investigated regions of Europe. A comprehensive geological image of that area was presented in Report 8 ICG (Illies, Fuchs eds 1974). This enormous zone of structural discontinuities was interpreted in the above-mentioned report by Illies (1974, Fig. 7) as a transform fault zone oriented WSW-ENE. That fault compensates, without a big shift, the effects of a Tertiary rift of the Limagne, Saône (Bresse) and Rhine grabens. The plate boundary runs from the Swabian lineament (Freudenstadt-Tubingen-Aalen), through the Freiburg area (Kaiserstuhl), as far as Montluçon in France. The role of that line in the Tertiary pattern should not be questioned; however, the Mesozoic pattern is more important for our search for an extension of the Drava-Gail line.

Boigk and Schöneich (1974) compared data on the thickness of the Permian, Triassic, and Lower Jurassic in the area between the Ardennes--Rhenic massif and the Mediterranean Sea. Figure 3 presents in a synthetic way the main results of the above-mentioned authors using selected isopachous lines which show the general paleogeography and zones of the maximum subsidence. The above comparison indicates that from the Lower Permian up to the Liassic there existed two independent main areas of a heavy subsidence. In the south it was the Burgundian trough oriented SW-NE, situated, generally speaking, between the area east of Lyon and Basel. Its southern extension was distinctly marked not until the Liassic (Dauphiné basin), although an earlier conjunction with the Rhône — Provence basin was also possible.

The other distinct subsidence area is known from the Permian — Triassic as an elongated Nancy — Pirmasens depression. It is also oriented SW-NE, and continues northwards (Werra trough, Hessen Senke). On the crossing of the Rhine graben with the Mesozoic subsidence zone a significant Tertiary volcanism could be observed (Vogelsberg).

Apart from the main subsidence areas, zones of a less significant subsidence were found; especially the "Kraichgau Senke", adjoining the

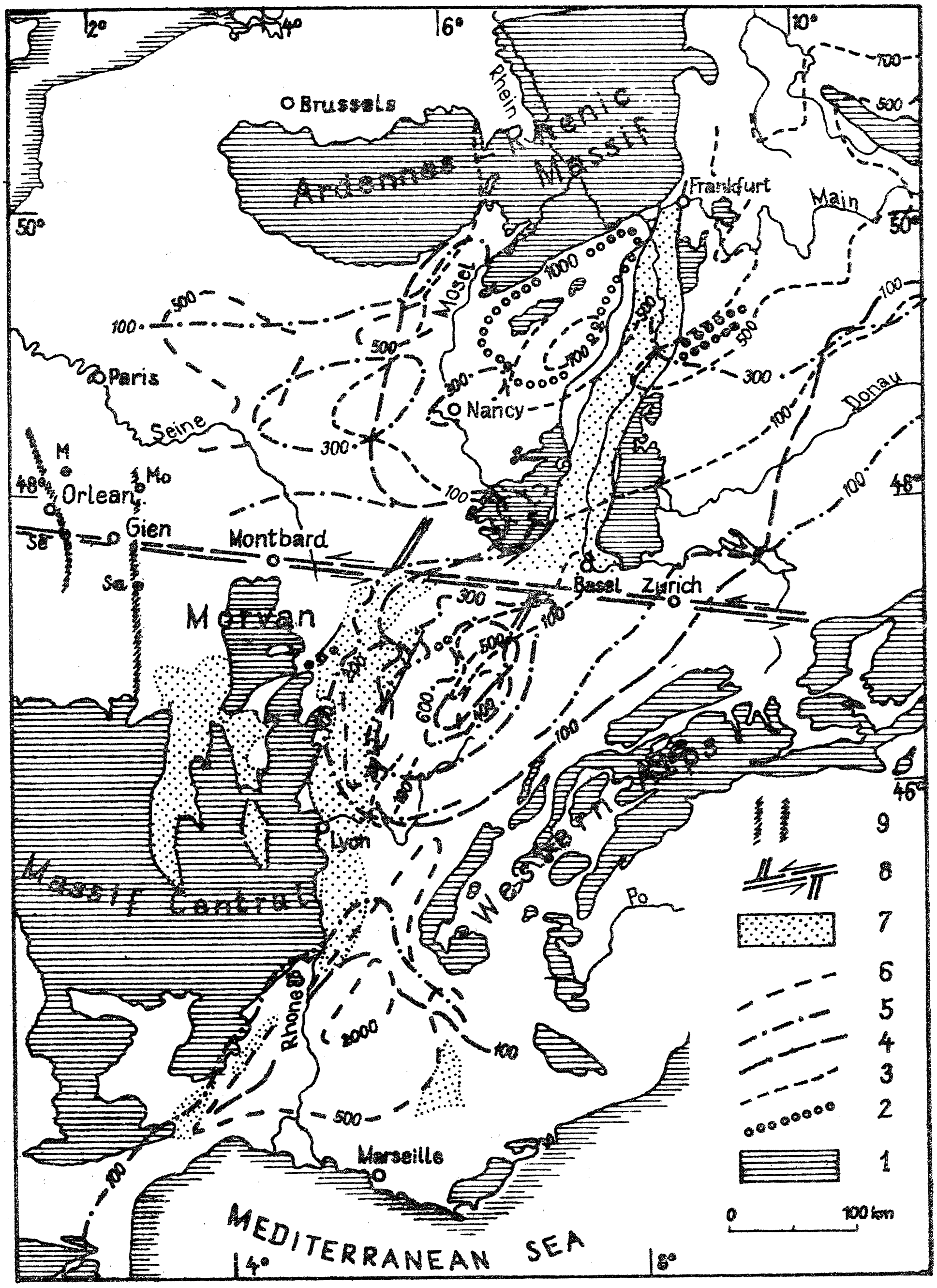

Fig. 3. Permian and Lower Mesozoic thickness distribution in the western Europe Tertiary rift zone. 1 — pre-Permian outcrops; 2—6 selected isopachous lines, characteristic of maximum subsidence zones (according to Boigk and Schöneich 1974); numbers denote thickness in metres (2 — Rotliegendes, 3 — Buntersandstein, 4 — Muschelkalk, 5 — Keuper, 6 — Liassic); 7 — Tertiary grabens and depressions; 8 — line of hypothetical Orlean-Zurich fault; 9 — Orlean area faults (according to Pomerol et al. 1980). M — Montvilliers, Se — Senelly, Mo — Montargis, Sa — Sancerre

Fig. 3. Rozkład miąższości permu i dolnego mezozoiku w strefie trzeciorzędowego ryftu zachodniej Europy. 1 — wychodnie prepermu; 2—6 — wybrane linie miąższości charakteryzujące strefy maksymalnej subsydencji (na podstawie Boigk i Schöneich 1974), liczby oznaczają miąższość w metrach (2 — czerwony spągowiec, 3 — pstry piaskowiec, 4 — wapień muszlowy, 5 — kajper, 6 — lias); 7 — grabeny i depresje trzeciorzędowe, 8 — linia hipotetycznego uskoku Orlean—Zurich, 9 — uskoki rejonu Orleanu (na podstawie Pomerol et al. 1980). M — Montvilliers, Se — Senelly, Mo — Montargis, Sa — Sancerre

Nancy depression from the east should be mentioned here. The intensity of subsidence moved southwards. In the Permian and Lower Triassic it was very conspicuous in the Nancy and Kraichgau depressions, in the Middle-Upper Triassic it increased in the Burgundian trough, and in the Liassic — in the Dauphiné basin. It may be assumed, therefore, that the two above indicated zones of heavy subsidence, primarily unconnected but lying on the same line, were separated by a sinistral transform fault (Fig. 3). These zones comprised the Nancy and Kraichgau depressions on one side and the Burgundian trough and its extension on the other. The transform fault, situated among the Morvan, Vosges and Schwarzwald massifs, may occur on the Montbard-Montbelard-Zurich line. A relative dislocation of plates, resulting from a shift of elongated depression axes, exceeds 100 km.

The time of formation and the activity of the fault are difficult to determine. Boigk and Schöneich (1974) observed that in the Liassic a separate subsidence zone (ca. 200 m thick, Fig. 3) was formed west of the Burgundian trough (the thickness of Liassic — up to 400 m) in the zone of a future Tertiary Saône (Bresse) graben. This may indicate that the dislocation of plates in the zone of a hypothetical Montbard-Zurich fault occurred at the end of Triassic or in the Early Jurassic, and that the new Saône depression began to form on an extension of the labille Nancy — Pirmasens depression on the other side of the fault.

A similar mechanism also operated later on. In the Middle Eocene a Rhine graben began to develop within the Vosges — Schwarzwald block: afterwards, its zone of maximum deposition shifted northwards (Illies, 1974). That graben was formed exactly opposite the labile (rift?) Burgundian trough whereas, at the same time, the Saône and Limagne grabens developed on a far extension of the labile, elongated Nancy — Pirmasens depression (Fig. 3). According to the above formulation, Tertiary grabens should be genetically connected with a propagation of movements from an earlier rift zone of the adjoining plate onto a still stable plate. A similar, Eocene age of initiation of the Saône and Rhine grabens should be stressed here. The difference in extent of the Tertiary grabens in relation to directions of the Mesozoic pattern may be due to a change in the regional stress field. A direct cause of the graben formation was a compression in the plate pattern along the N-S direction which occurred in the Lower Eocene — 53 m.y. (Dewey et al. 1973). It accounts for a subcrustal interrelation between grabens and the Alpine subduction accepted by Illies (1975).

In the light of the presented concept of the sinistral transform Montbard-Zurich fault, the folding of part of the Jura chain sedimentary cover and the lack of such a folding north of the fault seems evident. It is the result of an independent, but common with the Alps, evolution of the southern plate also in the Tertiary. Moreover, the difference in age and the change in character of the fault which is an extension of the Gail line, from dextral in the Southern Carpathians and the Eastern Alps to sinistral in the Basel area, should be noted (Fig. 1).

The Paris basin and the Armorican massif. The results of an analysis of the Basel area, presented in this paper, support the idea of a trans-European fracture zone oriented WNW-ESE, active in the Mesozoic. A problem of its western extension arises. The Montbard-Zurich line is not precise enough yet it indicates that the hypothetical lineament goes through the Paris basin and the Armorican massif. There has been no information concerning any significant Mesozoic tectonization of those areas. However, some regional geological features of the Paris basin support the idea of lineament. The following analysis makes use of a brief geological outline of the basin, recently worked out by a group of authors and edited by Pomerol (1979).

A map of the magnetic residual field, included in this outline, shows an intensive positive anomaly oriented NNW-SSE, crossing the Paris basin from the English Channel in the Rouen area up to the Central Massif. A belt of basic rocks, probably of Upper Proterozoic age, responsible for the anomaly, determines the boundary of geologically dissimilar blocks of the pre-Mesozoic basement of the Armorican type in the west and of the Morvan type in the east. In the Orlean — Gien region (Fig. 3) there is a shift of the southern part of the uninterrupted belt of basement basic rocks by ca. 40 km eastwards from the Montivilliers — Sennely line to the Gien — Sancerre line. This Loire bend can be also observed in the course of a crystalline rock belt found on the basis of gravity anomalies and borehole data.

There exists a concept (vide Pomerol ed. ibid.), based upon the leucogranite distribution, that the Morvan block was shifted by ca. 70 km towards NNW along the Sancerre fault at the end of Carboniferous. This movement appears to be connected with the above-mentioned shift of a block boundary between the submeridional Montivilliers — Sennely and Montargis — Sancerre faults (Fig. 3). It seems, however, that irrespective of the Morvan block movement towards NNW, an alternative simpler explanation of that shift can be found. A sinistral shift of the basic rock belt in the Orlean — Gien area may be connected with a western extension of the hypothetical Montbard — Zurich fault, oriented WNW—ESE. The extent of the Orlean — Gien shift is smaller than that resulting from separation of the Burgundian trough and the Nancy — Pirmasens depression.

It is possible that the extension of the Montbard — Zurich fault reaches as far as the Armorican massif in the west; in the Alençon area there exists a zone of tectonic deformations of Jurassic and Cretaceous deposits, while between Angers — La Fleche, Upper Cretaceous deposits overlie transgressively the older Mesozoic and the massif. In the Triassic and Liassic, the La Fleche — Orlean line delineated the south — western section of the Paris basin with an individual subsidence centre. In addition this line is noticeable in the distribution of Lower Miocene deposits. However, there are no data allowing a precise determination of an extension of the investigated lineament west of Orlean, or specification of time of the main tectonic activity along this line.

There are also no direct arguments, either, supporting the existence of an extension of the discussed Early Mesozoic fracture in the Armorican massif region. Theoretically, it is possible that some mylonitization zones, including the South Armorican Shear Zone considered Hercynian (Avedik 1975; Cogné 1978), are younger and are connected with the phenomena along the investigated lineament.

T h e G i b b s (C h a r l i e) f r a c t u r e z o n e. Despite the lack of any significant Mesozoic tectonization of the Paris basin and the Armorican massif, the results of an analysis of the Basel and Orlean areas, presented in this paper, calls for reference to the situation in the Atlantic Ocean.

Going from the Southern Carpathians as far as the Basel area, the NNW-SSE lineament under investigation is shifted, on an average, by ca. 1° northwards per each 7° of longitude (azimuth ca 280°). The direction thus determined indicates that in the zone of accreting Atlantic plates margin, the Gibbs fracture zone may constitute the extension of our lineament (Fig. 1); moreover, the latter zone has a segment of a similar azimuth (vide Le Pichon et al. 1973, Fig. 21, 34). In the fracture zone, the Reykjanes ridge was shifted by ca. 350 km in relation to the mid-Atlantic ridge. This fracture zone is one of those which separate sections of the Atlantic showing different evolution or, more precisely, parts of the ocean differing from one another with respect to the time of the first phase of active sea-floor spreading (La Pichon 1968; Pitman, Talwani 1972; Dewey et al. 1973; Nairn, Stehli 1974).

Except for the Late Triassic — Early Jurassic symptoms of rifting initiation (191 ± 6 m.y. to 202 ± 10 m.y. — Dewey et al. ibid.), an active Jurassic sea-floor spreading (180—148 m.y.) was observed only between North America and Africa, thus in the ocean section which is situated between the Romanche fracture zone and the Azores — Gibraltar fracture zone. In all those fracture zones and on their extension, evoked by the active sea-floor spreading and separation of continents were compensated. An eastern extension of the Azores — Gibraltar fracture zone compensated a relative motion of the African — Arabian plate towards ESE in relation to the Alpine — Mediterranean belt of microplates. Insignificant movements with a strike-slip component along the northern margin of the belt are connected with the Jurassic sea-floor spreading; this phenomenon manifested itself by a sinistral displacement in the Orlean — Zurich area or, maybe, also farther eastwards.

An intensive opening in the following, mainly — Cretaceous phase (123—80 m.y. — Devey et al., ibid.) already included the Atlantic Ocean as far as the Gibbs (Charlie) fracture zone in the north. During the Cretaceous the sea floor spreading occurred also north of the Gibbs fracture zone in subsequently failed rift zones of the Rockall trough (110—80 m.y. — Roberts 1975) and the Labrador Sea (80—63 m.y. — Dewey et al. 1973). However, the map of magnetic anomalies (Pittman, Talvani 1972) shows that the Gibbs fracture zone is a distinct discontinuity zone and, at the same time, a southern boundary of the part of the ocean which was intensively opened by the sea-floor spreading of the Reykjanes ridge, just beginning with Paleocene (63 m.y.).

The Gibbs and Azores fracture zones delimit the Atlantic region located in prolongation of the zone of Alpine Europe microplates. These latter include the Iberian microplate. At this point it is easy to understand the direction of the Bay of Biscay opening line (NW—SE), the extension of which is the Labrador Sea rift on the other side of the Gibbs fracture zone. This direction corresponds to the Dinaric strike, prevailing in the mobile microplate zone between the European plate and the African-Arabian one.

The Gibbs fracture zone reaches the Newfoundland fault system in the west (Havorth, Keen, 1979), and the south extremity of the Rockall trough in the east (Le Pichon et al. 1973, Fig. 34; Roberts 1975). Possibly, the southern boundary of the Porcupine Ridge is connected with this fracture zone. The Rockall trough does not compensate the effects of an intensive Cretaceous opening of the ocean south of the Gibbs fracture zone. In this situation the lack of distinct Cretaceous strike-slip movements on the eastern extension of the Gibbs fracture zone in the Armorican massif and farther on, is surprising. The compensation may be seen in the strike-slip component of the Pyrenees orogen, as well as along the initial opening line of the Bay of Biscay and the Labrador Sea (Dewey et al. 1973; Biju-Duval et al. 1977). This line traverses the Gibbs discontinuity zone.

In the light of the fore-going discussion and the above indicated central and north Atlantic opening, we should consider the existence of a genetic relation between the Gibbs fracture zone and the Orlean — — Montbard — Zurich — Southern Carpathians hypothetical faulting zone. This fracture line of the compensation delimits the microplate zone of the Alpine — Mediterranean belt from the north, and has Early Mesozoic foundations.

D o b r o g e a a n d N o r t h A n a t o l i a. In consequence of accepting the great drift of the Moesian platform westwards along the

Southern Carpathian front, we need to analyse the eastern extension of the transform fault. Airinei (1977) distinguished a Black Sea microplate in south-eastern Rumania, having determined its boundaries by means of gravity anomalies and seismic activity lines.

Airinei (ibid.) localized the northern margin of the Black Sea microplate north of the Danube delta, on the Focsani parallel, in the pre-Dobrogea Birlad depression. It should be noted that this margin is found on an extension of the important Timisoara — Brasov line, mentioned elsewhere, delimiting the Southern Carpathians from the north (Fig. 2). This Black Sea microplate boundary is connected with the Alpine orogenic belt of the southern Crimea and the Major Caucasus. According to the concept of linear plate margins, this orogenic belt does not bend northwards along the East European platform boundary but runs perpendicularly into the Eastern Carpathian chain. An extension of the Black Sea plate boundaries implies of a structurally individual character of the Southern Carpathians Getic nappe. It should be stressed that the folded Mesozoic belts of the Pannonian depression basement, mentioned elsewhere (Dank, Bodzay, Körösy in Lemoine ed. 1978, Figs 9.42, 9.43), as well as the Metaliferi Mts (vide Fig. 2) approximate the Dobrogea — Crimea strike.

In the south, the Black Sea microplate adjoins, according to Airinei, the Moesian microplate in the Constanca region, in the southern boundary zone (Capidava-Canara fault) of the Upper pre-Cambrian central Dobrogea massif. This boundary is an extension of the Southern Carpathian front fault. At the same time, the northern boundary of the massif, or the Pecineaga — Camena fault, accepted by Dewey et al. (1973) as the northern boundary of the Moesian plate, is found within the Black Sea microplate. It is likely that the southern boundary of the microplate in Dobrogea is connected with the North Anatolian strike--slip fault with a potent neotectonic activity (Fig. 1). A geological image of this fault was analysed by Pavoni (1961). The main idea of his interpretation has been supported by some geologists (vide Letouzey et al. 1977, Fig. 1).

On the extension of the Moesian and Black Sea microplates boundary in Turkey there is an eastern segment of the North Anatolian fault oriented WNW-ESE (azimuth ca. 110°), extending as far as the Wan lake. South of Sinop (Ladik region) an abrupt change in the fault direction into WSW-ENE is observed. The western segment continues as far as the Aegean Sea, becoming less distinct and spliting up. A rectilinear extension of the eastern segment towards WNW reaches the Black Sea on the Inebolu region. At the same time, this extension is a zone of great structural changes. Therefore we may accept that the northern fault boundary of the Moesian platform in Rumania is connected with the Inebolu — Ladik fault and with the eastern segment of the North Anatolian fault. However, the problem is more complicated.

The analysis of geological elements separated by the fault, carried out by Pavoni (ibid.), points out that it is a Laramide dextral fault whose dislocation amounts to 400 km. In the model proposed by the latter author (Pavoni, ibid., Figs 4,5) the main strike — slip displacement took place along the WNW-ESE line; a deflection of the western fault segment results from deformation of the southern plate shifted westwards (Turkish plate — Dewey et al. 1973). An Alpine tectonics of Anatolia, especially the Taurides broken arc and the Aegean Sea arc are connected with this deformation.

It should be noted that the North Anatolian fault, according to Pavoni, and the fault of the Moesian platform northern boundary in Rumania, described above, correspond with each other in time of formation, extent of dislocation and direction. Pavoni's model may be included with the concept of the Black Sea opening between two boundaries of the Black Sea microplate; at the same time, an important role must have been played by fractures in the western part of that basin and between the Balkanides and the Pontides.

However, joining the fault of the Moesian platform northern boundary and the North Anatolian fault, with regard to Pavoni's interpretation, is not a simple matter. It is not clear in which zone the western segment of the fault went beyond the Pontic geosyncline area and passed onto the platform. The problem of the eastern extent of the Moesian platform and the Black Sea microplate remains open to discussion. In this situation, the similarity between the analysed structural line systems in the Pontides and in the Alps should be considered. The eastern segment of the North Anatolian fault may be compared with the Pusteria — Gail line, while the western one — with the Judicaria fault and the Insubric line. The Inebolu — Ladik fault and its hypothetical extension in the Black Sea play a similar role as the Montbard--Zurich fault. Irrespective of the question whether a model of deformation of the old, rectilinear plate boundary, or that of virgation is accepted, the role of Inebolu-Ladik fault, running towards Constanca, remains important.

It seems that the eastern extension of the North Anatolian fault is not connected with the Zagros system, as proposed by Pavoni, but continues rectilinearly towards ESE in the Tabriz area and in the Elburs Mts, as accepted by Dewey et al. (1973). The orogenic Zagros belt has a prevailing direction of NW-SE, the same as the Alpine — Mediterranean microplate belt (Fig. 1).

RECAPITULATION AND CONCLUSIONS

1. West of the Fennoscandian shield, the East European platform, the Moesian platform, and the Rhodope massif there existed in the Permian-Mesozoic a wide, labile NW-SE oriented belt with rift zones and an anomalously heavy subsidence in certain periods. Elements of that belt can be observed both in the epi-Paleozoic platform area and in the geosynclinal region. They intersect other elements of the geosynclinal structural pattern (the Western Carpathians, the outer zone of the North-Eastern Carpathians, the Pannonian basin) or are congruent with them (the innerflysch Eastern Carpathians, the Kraishtides, the inner Dinarides and the eastern section of the Hellenides). The Lower Saxony — Vienna and Danish — Polish lineaments, which differ in the general picture but are similar as regards some of their segments, formed marginal zones of that belt on the platform. In the north they joined the meridional system of the North Sea rift. South of the Drava-Danube line, among eugeosynclinal deposits of the belt (the Vardar and Ophiolitic (Serbian) zones, the sub-Pelagonian zone), Upper Jurassic ophiolites are present. At the end of Cretaceous, inversion and paroxysmal tectonic movements were observed in linear subsidence zones of that labile belt. Those phenomena in the geosyncline area were contained within the pattern of early and late movements of the Alpine orogeny.

2. Special attention should be paid to the linear NW-SE system of fractures of the labile belt, the intensification of rifting southwards, the presence of sediments of the oceanic floor type in the inner Dinarides and Hellenides, the lack of symptoms of the Mesozoic rift on the present extension of that belt on the African-Arabian plate, as well as to the fact that the initial rifting in that belt started in the Permian. Therefore, we should refer to movements of Africa and Arabia in relation to Eurasia, as well as to the reconstructed accreting plate margins (Dewey et al. 1973). This reference indicates that in the initial phase the rift situated between the African-Arabian plate on one side and the Turkish and Iranian plates on the other (Tethys 1 — Dewey et al. ibid.) was a linear extension of the NW-SE oriented rifting belt of the platformal and Alpine Europe. At present that rift zone is found in the deformed Taurides, and in the linear Zagros system and its foreland (Stöcklin 1978). The rift started in the Permian (Oman spilites — vide Dewey et al. ibid.). Until the beginning of Jurassic, up to the opening of the Atlantic section between the Romanche and Azores — Gibraltar fractures it formed a linear system together with the Dinaride — Hellenide rift and their northern extension, intersecting diagonally the protogeosynclinal, W-E oriented Tethys belts: the Carpathians—Dobrogea — the Major Caucasus and Carpathians — the Balkanides — the Pontides. The conjunction of the Hellenides with the Zagros rift system in the Late

Triassic and Liassic was already accepted by Dewey et al. (1973). The Jurassic drifting of Africa eastwards in relation to Europe and the microplate belt, followed by the Cretaceous sinistral rotation of Africa, separated the Dinaride — Hellenide lineament from the Zagros one. These movements deformed the connecting segment, or Cyprus — the Taurides and the Bitlis zone. The direction of the Dinaride — Zagros rift remained dominant in the Alpine — Mediterranean microplate belt.

The linear rift concept brings about a problem of a south — eastern extension of the Permian — Mesozoic Zagros — Oman system. The Carlsberg ridge (Fig. 1) resembles that system in its NW-SE strike. It joins the Tertiary Red Sea rift through a sigmoidal bend of the Gulf of Aden. If the Jurassic — Cretaceous drifting eastwards and the sinistral rotation of the African-Arabian plate are considered, it may be assumed that in the Permian — Early Mesozoic the Zagros — Oman rift had linear extension in the initial fracture system of the future Indian Ocean. The ocean-floor spreading developed from those fractures only in the Upper Cretaceous (Mc Kenzie, Sclater 1971), and has been continuing up to the present in the Carlsberg ridge/rift. The effects of drifting and differences in rift intensity of the Red Sea and the Zagros-Oman system in relation to the Carlsberg ridge/rift were compensated in the Owen fracture zone (Fig. 1). Therefore, the rift of the Zagros-Oman system and of its extension in Europe as far as the North Sea may be connected with the initial Indian Ocean rift. The effect of the latter rift was reflected in the geological picture of Europe and of the Alpine-Mediterranean microplate belt. The NW-SE oriented rifting belt intersected the geosynclinal Tethys belt, directed W-E, or the former suture between Eurasia and Gondwana, and interfered with it.

3. In western Europe, in the Permian — Mesozoic pattern, there occur similar linear zones of heavy subsidence. They are represented by the Nancy-Pirmasens depression and the Hessen depression in the north, and by the Burgundian trough and the Dauphiné basin in the south. These zones extend SW-NE or SSW-NNE. It is likely that in the Permian and Triassic they were situated on the same line forming a lineament with a rift genesis. Anomalously thick Permian — Triassic platformal sediments are found in Djeffara in Libya and in southern and eastern Tunisia (Burollet et al. 1978). In the pre-Liassic pattern of the main plates (vide Dewey et al. 1973) these sediments may represent a farther segment of the lineament. Perioceanic Mesozoic basins of western Africa (the Essaouira basin, the Aaiun basin — Dillon, Sougy 1974) and Portugal also belong to the Atlantic system.

4. In the Tertiary, on the extension of Permian — Mesozoic linear zones of rifting and anomalous subsidence, new rifts, i.e. the Red Sea rift, the Rhine and Saône grabens, were formed in the adjoining plates. On the northern extension of axis of the whole, NW-SE oriented rif-

ting belt, the anomalous, volcanic Faeroe — Iceland ridge is of a similar origin. It was formed as the opening of the northern part of the North Atlantic proceeded in the Tertiary. Traces of young volcanism of eastern Greenland and the Baffin Bay may be also connected with that ridge. The Faeroe — Iceland ridge volcanism determines the anomalous zone on the background of the North Atlantic floor spreading. This zone occurs on an extension of the Indian Ocean opening axis. Together with the Red Sea rift and the NW-SE oriented rifting belt of Europe, they delimit the Sonder lineament. It is a line of post-Variscan activity symptoms of a stable, linear mantle expansion zone over which the African-Arabian plate was dislocated.

5. The line of WNW-ESE oriented transform faults, described in this paper, played an important role in evolution of the post-Variscan Europe. In the western part, between Orlean and Zurich, it is a sinistral fault which was presumably active mainly in the Early Jurassic. That activity was an echo of the eastern drifting of Africa in relation to Europe's microplate belt (Dewey et al. 1973). In the eastern part, from the Judicaria fault in the Alps up to Anatolia, it is a dextral fault, active mainly at the end of Cretaceous. It was formed between the Santonian and Paleocene, or 80—63 m.y. ago (Dewey et al. 1973) as a result of a sinistral rotation and drifting of the African-Arabian plate together with the microplate belt westwards with respect to Europe. It is very likely that the faults described above are related to the Gibbs (Charlie) fracture zone in the North Atlantic. Both the fracture zone and the fault line make a boundary between the European plate and the zone of mobile Alpine-Mediterranean microplates, described by Illies (1975) as „macro--mylonite" for the Cenozoic model of plate arrangement. In the Alpine zone a tensional and compressional orogenic activity which followed the N-S direction was temporarily connected with that fault line.

Beginning with the Carpathians (the Metaliferi Mts, the Timisoara — Brasov line) and northern Dobrogea, a boundary of a similar type was represented by the northern margin of the Black Sea plate.

6. The Black Sea and Caspian Sea opening, unlike the Mediterranean Sea one, is connected with the northern boundary of the microplate belt and with the northern boundary of the Black Sea-Caspian Sea plate situated outwards, east of the Carpathians (Dewey et al. 1973).

7. Linear, Permian — Mesozoic rift characteristics of the "Indian Ocean system" (NW-SE) and the "Atlantic system" (SW-NE) can be seen both in the platformal and Alpine Europe plan. These two systems partly penetrate each other and interfere with the W-E — oriented Tethys geosyncline system. The delimitation line of the prevailing Atlantic and Indian Ocean influence is distinct on the platform. It is demarcated by the North Sea central rift. Farther south, this line runs

through the Munich area into the Alps (where it is delimited by the passage of the Judicaria fault into the Gail line) and along the Adriatic.

The linear rifting systems are intersected by a parallel belt of the Alpine-Mediterranean microplates the northern boundary of which runs from the Gibbs fracture zone to the Dobrogea region and, farther on, along the northern margin of the Black Sea-Caspian Sea plate. The southern boundary runs from the Azores-Gibraltar fracture zone, through the Mediterranean Sea, to the Bitlis zone. Both rifting systems refer to the north — western and north — eastern boundaries of the African--Arabian plate. Identification of Early Mesozoic, pre-Liassic lineaments of both systems diverging southwards on the European and African--Arabian plates, mobile with respect to each other, should allow an estimation of the degree of approximation of both continents in the Mediterranean belt of subductive boundaries, inclusive of the one in the Azores-Gibraltar ridge.

The outer boundaries of the Mediterranean microplate belt of Europe are connected with phases of the Atlantic Ocean development. They were established in the approximation zone of both rift systems, in a labile zone of European and North African Paleozoides, north of the African-Arabian craton.

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STRESZCZENIE

Dla określenia czasu, kierunku i wielkości ruchów płyt i mikropłyt przydatna jest analiza transformujących uskoków i akrecyjnych granic płyt, natomiast strefy subdukcji kryją wiele niejasności. W Europie ważne są w związku z tym młode, linearne pęknięcia skorupy przechodzące z epipaleozoicznej platformy na obszar alpidów. Autor nawiązuje do idei regmatycznej linii Islandia-Morze Czerwone (Sonder 1938).

W permomezozoicznym planie epiwaryscyjskiej platformy i alpidów Europy zaznacza się system pęknięć o kierunku NW-SE sięgających od Morza Północnego przez Karpaty i Półwysep Bałkański do Morza Egejskiego. Wyróżniono pas objęty tym systemem, obrzeżony linearnymi strefami (lineamentami) o zwiększonej subsydencji w permie i mezozoiku (fig. 1).

W s c h o d n i l i n e a m e n t zaczyna się w strefie ryftu Morza Północnego i kontynuuje się na platformie w postaci synsedymentacyjnego rowu duńsko-polskiego. Jego przedłużeniem w geosynklinie była strefa jednostki skolskiej o anomalnie dużej miąższości kredy (strefa Szewczenkowa), a dalej ku SE, w Karpatach Wschodnich, tytononeokomski rów Sinaia-Palanca. Z osadów tego rowu uformowały się jednostki Rachov, Porkulec, Ceahlau, Teleajen, ciągnące się aż po Dimbovicę na NW od Bukaresztu. Osady rowu Sinaia-Palanca znajdują się też dalej na zachodzie, w Karpatach Południowych, w postaci jednostki Severin-Kraina. Zostały one przesunięte górnokredowym dryfem platformy mezyjskiej, z którą była związana luźno strefa przedbałkańska i obszar obecnych bałkanidów. Transformujący uskok oddzielający Karpaty Południowe od mikropłyty mezyjskiej odegrał ważną rolę przypuszczalnie jeszcze przed dryfem Mezji. Z linearnym przedłużeniem rowu Sinaia-Palanca na południe od uskoku był związany nie tylko obszar sedymentacji osadów jednostki Severin-Kraina, ale również rów Krajsztydów. Rysy strukturalne Krajsztydów odchodzą ku południowi od obszaru karpato-bałkańskiej geosynkliny i widoczne są aż po Morze Egejskie.

W planie przedsenońskim, a zwłaszcza dolnokredowym, karpacka geosynklina miała kształt odwróconej sigmoidy (fig. 2A). Istniały w niej dwie strefy wirgacji. Na północy od strefy linearnego ryftu wyznaczonego rowem duńsko-polskim i rowem Sinaia-Palanca Karpat Wschodnich odchodziła ku zachodowi geosynklinalna rynna piaszczystego fliszu jednostek śląskiej i podśląskiej Karpat Zachodnich. Na południu od linearnej ryftowej strefy Karpat Wschodnich — Krajsztydów odchodziła ku wschodowi rynna geosynkliny bałkańsko-pontyjskiej. W sąsiedztwie obu wirgacji istnieją wykazujące dużo cech wspólnych masywy Sanu i serbo-macedoński (fig. 2). Jest możliwe, że również ten ostatni wypiętrzył się dopiero w górnej kredzie w efekcie ruchów przedkampańskich i rozdzielił obszar tytonokredowych utworów, głównie fliszowych, utwo-

rzonych pierwotnie we wspólnym basenie na odrębne strefy Vardaru i Krajsztydów. Tak więc przedłużeniem lineamentu duńsko-polsko-wschodniokarpackiego na południe od transformującego uskoku frontu Karpat Południowych może być potrójny, rozdzielony masywem osiowym lineament Vardaridy-Krajsztydy Bonczewa lub jego wschodnia część. Rysy strukturalne o kierunku NW-SE widoczne są w Morzu Egejskim aż po linię wysp Andros-Samos.

Zachodni lineament obrzeżający pas pęknięć o kierunku NW-SE zaczyna się w sąsiedztwie ryftu Morza Północnego. Reprezentują go basen Dolnej Saksonii o anomalnie dużej miąższości osadów permu i mezozoiku, synklinorialne strefy obrzeżające masyw Harcu oraz pęknięcia o kierunku NW-SE na obszarze masywu czeskiego. Ważną strefę anomalnej subsydencji wyznaczają osady jury na skłonie masywu czeskiego w podłożu molas zapadliska przedkarpackiego między Dunajem a Hodoninem. Neogeński wewnętrzny basen wiedeński przecinający łuk Karpat potwierdza labilność tej strefy i skłonność do subsydencji również w trzeciorzędzie.

Na przedłużeniu dolnosaksońsko-wiedeńskiego odcinka lineamentu, na tle pasów osadów mezozoicznych o rozciągłości SW-NE rozpoznanych w podłożu neogeńskich osadów depresji panońskiej zaznaczają się odcinki o anomalnej skłonności do subsydencji w mezozoiku. Są to obszary gór Bakony i Mecsek podniesione w neogeńskim planie depresji. Dalej na południe, między Drawą i Sawą znajduje się zaburzona strefa horstów i grabenów stanowiąca przypuszczalnie przedłużenie strefy transformującego uskoku rozdzielającego platformę mezyjską od Karpat Południowych. Na przedłużeniu linii basen wiedeński — Mecsek znajduje się strefa Vardaru wewnętrznych Dynarydów (fig. 2). Ponieważ obszar na południe od uskoku został przesunięty z końcem kredy ku zachodowi, jest możliwe, że w planie mezozoicznym naprzeciw omawianego zachodniego lineamentu znajdowała się strefa „ofiolitowa" (strefa Serbii), główna ofiolitowa strefa Dynarydów. Poprzez strefę Mirdita łączy się ona ze strefą subpelagońską Hellenidów. Duża ilość bazytów i ultrabazytów późnej jury jest charakterystyczna dla tego eugeosynklinalnego pasa. Labilny pas pęknięć Europy platformowej miał więc przedłużenie aż po środkową lub późną kredę w Krajsztydach i ofiolitowych strefach Dynarydów-Hellenidów. Z końcem kredy lub z początkiem trzeciorzędu w pasie tym zaznaczyły się paroksyzmalne ruchy tektoniczne częściowo o inwersyjnym charakterze.

Lineament wyznaczony przez pas ryftingu został uzupełniony w trzeciorzędzie o odcinek młodych wulkanitów grzbietu Wyspy Owcze-Islandia na północy oraz o ryft Morza Czerwonego na południu (fig. 1).

Przeprowadzono analizę stref znajdujących się na przedłużeniu uskoku frontu Karpat Południowych, stanowiącego północną granicę płyty mezyjskiej (fig. 1, 2). Analizę podjęto mimo trudności w przeprowadzeniu uskoku przez obszar jednostki getyckiej w rejonie przełomu Dunaju. Możliwy jest związek tego uskoku z linią Drawy rozdzielającą obszar Dynarydów od podłoża depresji panońskiej o odmiennym planie strukturalnym. Przedłużeniem tej ważnej granicy jest przypuszczalnie linia Pusteria-Gail, rozdzielająca Alpy Południowe od Wschodnich. Według Van Bemmelena linia ta jest dekstralnym uskokiem przesuwczym. Potwierdza to analiza złóż rud Pb-Zn zlokalizowanych po obu stronach tej linii.

Przedłużeniem linii Gail jest uskok Judicaria, a następnie linia Insubric kontynuująca się do równiny Padu. Natomiast na prostolinijnym przedłużeniu linii Gail ku zachodowi znajduje się granica Alp Centralnych i Wschodnich. Jest możliwe, że przesuwczy uskok linii Gail w tym właśnie rejonie (Graubünden) opuszcza geosynklinę alpejską i zaznacza się na platformie ograniczając od północy łańcuch gór Jura.

Autor przeprowadził analizę permomezozoicznego planu strefy trzeciorzędowego ryftu między północnym zakończeniem grabenu Saony (Bresse) i południowym zakończeniem grabenu Renu w oparciu o materiały Boigka i Schöneicha (1974). W planie tym zaznaczają się strefy o anomalnie dużej subsydencji, o rozciągłości SW-NE. Należą do nich niecka Nancy-Pirmasens-Kraichgau na północy oraz rów burgundzki i jego przedłużenie na południu. Autor wysunął koncepcję, że strefy osiowe tych dwóch depresji znajdowały się pierwotnie na jednej linii. Zostały one rozsunięte z końcem triasu lub w liasie sinistralnym uskokiem transformującym Montbard-Zurych (fig. 3) mającym związek z linią Gail. Trzeciorzędowe grabeny mogą być w tym ujęciu związane z przenoszeniem ruchów z labilnej strefy mezozoicznego ryftu sąsiedniej płyty na płytę jeszcze stabilną.

Jest możliwe, że wczesnomezozoiczne ruchy zaznaczyły się na przedłużeniu hipotetycznego uskoku Montbard-Zurych w podłożu basenu paryskiego koło Orleanu, a być może także w masywie armorykańskim.

Na dalekim przedłużeniu linii uskoków transformujących Orlean — Karpaty Południowe znajduje się ważny dla ewolucji Atlantyku rozłam Gibbsa. Oddziela on część oceanu, która przeszła fazę intensywnego otwarcia w kredzie od części północnej, której intensywne otwarcie zaczęło się w paleocenie, w ryfcie grzbietu Reykjanes. Omówiona linia pęknięć o rozciągłości WNW-ESE ogranicza od północy strefę mikropłyt alpejsko-medyterrańskiego pasa i ma założenia wczesnomezozoiczne. Południowe ograniczenie tego pasa stanowi rozłam Azory—Gibraltar i jego wschodnie przedłużenie.

Jest prawdopodobne, że uskok frontu Karpat Południowych przedłuża się ku wschodowi i ma związek z uskokiem Inebolu-Ladik oraz wschodnim segmentem północnoanatolijskiego rozłamu. W ujęciu Pavoniego (1961) rozłam ten jest dekstralnym laramijskim uskokiem przesuwczym.

WNIOSKI

1. Szeroki pas objęty w permie-mezozoiku systemem pęknięć o kierunku NW-SE znajduje się na zachód od kratogenicznych obszarów tarczy fennoskandynawskiej, platform wschodnioeuropejskiej i mezyjskiej oraz masywu Rodopów. Pas ten obrzeżony jest na platformie lineamentami dolnosaksońsko-wiedeńskim i duńsko-polskim. Na północy łączy się z meridionalnym systemem ryftu Morza Północnego, na południu wchodzi na obszar geosynklinalny.

2. Podkreślić trzeba linearny układ NW-SE pęknięć tego labilnego pasa, intensyfikację ryftingu ku południowi, obecność utworów typu oceanicznego dna w wewnętrznych Dynarydach i Hellenidach, brak przejawów mezozoicznego ryftu na obecnym przedłużeniu tego pasa na płycie afrykańsko-arabskiej, a także fakt, że inicjalny ryfting w tym pasie zaczął się w permie.

Autor nawiązał do wyników analizy ruchów Afryki i Arabii względem Eurazji oraz do zrekonstruowanych granic akrecyjnych (Dewey et al. 1973). W fazie inicjalnej, począwszy od permu, przedłużeniem omówionego pasa ryftingu NW-SE Europy był ryft umiejscowiony między płytami afrykano-arabską z jednej a turecką i irańską z drugiej strony. Obecnie strefa tego rytfu mieści się w zdeformowanych Taurydach, strefie Bitlis i linearnym systemie Zagros-Oman. Aż do początku jury, do otwarcia sektora Atlantyku między rozłamami równikowymi (Romanche i sąsiednie) i rozłamem Azory—Gibraltar tworzył on linearny system z ryftem Dynarydów—Hellenidów i ich północnego przedłużenia, a przypuszczalnie także z inicjalnym ryftem przyszłego Oceanu Indyjskiego. Ten linearny system przecinał skośnie równoleżnikowe, protogeosynklinalne pasy Tetydy: Karpaty — Dobrudża — Krym — Kaukaz i Karpaty — Bałkanidy — Pontydy.

Jurajski dryf płyty afrykano-arabskiej ku wschodowi względem Europy i pasa mikropłyt, a następnie kredowa sinistralna rotacja Afryki rozsunęły lineamenty Dynarydów-Hellenidów i Zagrosu oraz Zagrosu i centralnego ryftu Oceanu Indyjskiego (fig. 1). Wpływ inicjalnego ryftu tego oceanu został utrwalony w geologicznym obrazie platformowej Europy i alpejsko-medyterrańskiego pasa mikropłyt.

3. W planie permomezozoicznym zachodniej Europy istnieje linearna strefa anomalnej subsydencji o kierunku SW-NE lub SSW-NNE obejmująca nieckę Nancy-Pirmasens i depresję Hesji na północy a rów bur-

gundzki i nieckę Dauphiné na południu. Podobny kierunek mają perioceaniczne baseny mezozoiczne północno-zachodniej Afryki i Portugalii. Mają one związek z ewolucją Atlantyku, reprezentują atlantycki system ryftingu.

4. W trzeciorzędzie na przedłużeniu permomezozoicznych linearnych stref ryftingu i anomalnej subsydencji tworzyły się w sąsiednich płytach nowe ryfty — grabeny Renu i Saony (Bresse), ryft Morza Czerwonego. Ten ostatni a także anomalny wulkaniczny grzbiet Wysp Owczych — Islandii wraz z omówionym pasem NW-SE pęknięć Europy wyznaczają lineament Sondera. Znajduje się on na przedłużeniu osi otwarcia Oceanu Indyjskiego. Lineament jest linią przejawów aktywności stabilnej, linearnej strefy ekspansji płaszcza, nad którą przemieszczała się afrykańsko-arabska płyta.

5. Ważną rolę w ewolucji powaryscyjskiej Europy pełni opisana linia transformujących uskoków o rozciągłości WNW-ESE. W części zachodniej, między Orleanem a Zurichem jest to sinistralny uskok, który był aktywny przypuszczalnie głównie we wczesnej jurze. Aktywność ta była oddźwiękiem wschodniego dryfu Afryki względem pasa mikropłyt Europy (Dewey et al. 1973). W części wschodniej od uskoku Judicaria w Alpach aż po Anatolię jest to uskok dekstralny, aktywny głównie z końcem kredy. Powstał on w efekcie sinistralnej rotacji i względnego dryfu afrykano-arabskiej płyty wraz z pasem mikropłyt ku zachodowi w stosunku do Europy między santonem a paleocenem, czyli 80—63 mln lat temu (Dewey et al. 1973). Związek opisanych uskoków z rozłamem Gibbsa (Charliego) na Północnym Atlantyku jest bardzo prawdopodobny. Tak rozłam, jak i linia uskoków tworzą granicę między stabilną płytą europejską a strefą mobilnych mikropłyt alpejsko-medyterrańskich. W strefie alpejskiej z linią tą związana była okresowo tensyjna lub kompresyjna aktywność orogeniczna wzdłuż kierunku N-S.

Począwszy od Karpat (Metaliferi Mts, linia Timisoara—Brasov) i północnej Dobrudży granicę podobnego typu reprezentuje północny brzeg płyty czarnomorskiej.

6. Otwarcie Morza Czarnego i Kaspijskiego w przeciwieństwie do Śródziemnego związane jest z północną granicą pasa mikropłyt i północną granicą znajdującej się na zewnątrz, na wschód od Karpat płyty czarnomorskiej (Airinei 1977) — kaspijskiej (Dewey et al. 1973).

7. Linearne, permomezozoiczne rysy ryftu systemu „Ocean Indyjski" (NW-SE) i systemu „Atlantyk" (SW-NE) widoczne są tak w planie platformowej, jak i alpejskiej Europy. Częściowo się przenikają. Interferują z systemem W-E geosynkliny Tetydy. Linia rozgraniczenia dominującego wpływu Atlantyku i Oceanu Indyjskiego jest wyraźna na platformie. Wyznacza ją centralny ryft Morza Północnego. Dalej ku południowi linia ta biegnie przez rejon Monachium w Alpy (gdzie wyznacza ją przejście uskoku Judicaria w linię Gail) i dalej wzdłuż Adriatyku.

Systemy linearnego ryftingu przecięte są równoleżnikowym pasem alpejsko-medyterrańskich mikropłyt, którego północna granica biegnie od rozłamu Gibbsa do rejonu Dobrudży i dalej północnym brzegiem płyty czarnomorsko-kaspijskiej. Granica południowa biegnie od rozłamu Azory — Gibraltar przez Morze Śródziemne do strefy Bitlis. Oba systemy ryftingu nawiązują do północno-zachodniej i północno-wschodniej granicy płyty afrykano-arabskiej. Identyfikacja wczesnomezozoicznych przedliasowych lineamentów obu rozchodzących się ku południowi systemów na mobilnych względem siebie płytach europejskiej i afrykano-arabskiej pozwoli ocenić wielkość zbliżenia obu kontynentów w medyterrańskim pasie subdukcyjnych granic, w tym także w grzbiecie Azory — Gibraltar.

Zewnętrzne granice pasa medyterrańskich mikropłyt Europy związane są z etapami rozwoju Atlantyku. Powstały one w strefie zbliżenia obu systemów ryftu, w labilnej strefie paleozoidów Europy i północnej Afryki, na północ od afrykano-arabskiego kratonu.