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THE FORMATION OF THE INTRA-CARPATHIAN BASINS AS DETERMINED FROM SUBSIDENCE DATA

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The Carpathian arc is the result of continental collision during subduction of the European plate beneath a Pannonian continental block. In the Early/Middle Miocene, during and after the last stages of thrusting in the Outer Carpathians, several "back-arc" basins started to form within the Carpathian loop. These basins are of two types: (1) those lying in the peripheral regions of the intra-Carpathian lowlands (Vienna, West Danube, Transcarpathian and Transylvanian), and (2) those lying in the central intra-Carpathian region (East Danube, Little Hungarian and Great Hungarian (Pannonian)).

Though both groups of basins have thin crust, the subsidence history and the present heat flow are different. The peripheral basins exhibit a rapid initial subsidence followed by a much slower general increase in depth. Their heat flow is close to the average for continental areas. In contrast the central basins have no initial subsidence but do show a fast linear increase in depth which has continued until the present. The heat flow is nearly twice the average for continents.

We believe that the basins are thermal in origin and are the direct result of the continental collision which formed the Carpathian arc. The peripheral basins appear to be the result of uniform stretching of the lithosphere by about a factor of two. The rapid initial subsidence is an immediate isostatic adjustment to the stretching, the slower linear subsidence is due to conductive cooling of the thinned lithosphere. In the central basins, uniform stretching by about a factor of 3 could explain the thermal subsidence and the high heat flow. Unfortunately such a simple explanation is not supported by either the geology or the absence of a clearly defined initial subsidence. Alternative explanation involve crustal stretching with additional subcrustal thinning or, alternatively, attenuation of the whole subcrustal lithosphere and part of the crust by melting and erosion. Both explanations create a very thin lithosphere, reduce the initial subsidence to a minimum but still give a rapid thermal subsidence and high heat flow.

The subsidence history gives quantitative information concerning the evolution of the inter-Carpathian basins. In other areas, it may place equally important constraints on the development of intercontinental basins and continental shelves.

1. Introduction

One of the fundamental concepts of plate tectonics is that sea floor is created by the intrusion of hot molten material along mid-ocean ridges. As this material moves away from the spreading centers, it cools and contracts. This intrusion process creates an initial thermal anomaly in the lithosphere, and the subsequent cooling of the lithosphere toward thermal equilibrium can explain the observed increase in depth of the ocean floor and the decrease in heat flow with age [1]. Continental shelves and intracontinental basins also appear to subside with age in an analogous fashion. Like the oceans gravity anomalies are small and hence the subsidence is compensated. Although the early subsidence of these regions

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may be complex, 20–30 m.y. after formation the increase in depth is observed to be exponential and to have a time constant of 50–60 m.y. [2–4]. Because this is similar to the sbusidence observed for normal ocean floor, it is strong evidence that the long-term subsidence of continental shelves and basins also results from the decay of a thermal anomaly in the lithosphere. Except for a study of the Gulf of Lyon [5] little effort has been devoted to analyzing the early subsidence of these regions. Hence here is little quantitative data on which to base a reasonable reconstruction of the specific thermal and mechanical processes which are associated with the creation of continental shelves and inland basins.

A variety of models have been suggested to account for these regions. However, most early explanations involved only crustal processes [6–8] and are unable to account for long-term subsidence. Other models have suggested that these regions have been created by a thermal plume which causes the lithosphere to expand and results in uplift and erosion of the crust [2]. As the lithosphere cools a basin is formed because the upper surface returns to a depth below its original position. However, the amount of sediment deposited will correspond roughly to the amount of crust removed by erosion. Thus thick sedimentary deposits on continental shelves and in inland basins present a major space problem for this model. Based on work in the Aegean, McKenzie [9,10] has recently suggested simple extension as an explanation for basin formation. Royden et al. [11] have suggested similar extensional explanations of these areas based on an analysis of Atlantic-type continental margins. In this paper three simple mechanical models will be dealt with in detail. Two of these mechanisms, stretching and dike intrusion, involve significant extension of both crust and lithosphere. Since these processes have been discussed in detail in previous publications [9,11,12], they will only be briefly described. The third involves attenuation of the subcrustal lithosphere and can leave the crust relatively undeformed. Each of these mechanisms result in two distinct phases of subsidence or uplift. The first is an immediate isostatic response to density changes associated with changes in the structure of the lithosphere and can result in either subsidence or uplift. The second is a long-term subsidence controlled by decay of the thermal anomaly associated with that change in structure. As will be seen this is an important distinction.

Subsidence and sedimentation began in the intra-Carpathian basins in the early Miocene, during and after the last stages of compression and thrusting in the Outer Carpathians. Subsidence has continued until the present. We show that in these basins both an initial, isostatic subsidence, and the long-term thermal subsidence are observed. Our objective in this paper is to examine the subsidence of these basins in a regional geologic and tectonic framework and to show that calculations of sedimentation rate and total sediment thickness give valuable insight into the mechanical processes which formed these basins.

We start by analyzing the data within the framework of the crustal stretching model that has been successful in simple extensional settings [10,12]. However, we are forced by the observations in the Pannonian Basin to consider a modified version of the original model.

2. Three thermal models

2.1. Stretching

Rapid extension of the entire lithosphere results in pronounced attenuation in both the crust and the underlying material (Fig. 1a). When extension occurs, the upper portion of the lithosphere exhibits brittle failure, whereas the lower lithosphere extends and attenuates by ductile flow and is replaced by passive upwelling of hot asthenosphere. Graben formation, block and listric faulting are manifestations of brittle failure in the crust. This attenuation of lighter crustal rocks results in an initial isostatic subsidence when crustal thickness is originally greater than about 20 km (Fig. 2a). After the initial isostatic subsidence, the early ($t \leqslant 20$ m.y.) thermal subsidence is approximately linear (Fig. 2b). In this model it is not necessary that the extension be uniform, i.e. that the crustal thinning exactly match the attenuation of the lithosphere. Flow in the ductile layer can attenuate the lithosphere over a broader area than that observed in the crust (Fig. 1c).

2.2. Dike intrusion

Rapid extension leads to cracking of the lithosphere and large-scale intrusion of vertical dikes (Fig. 1b)

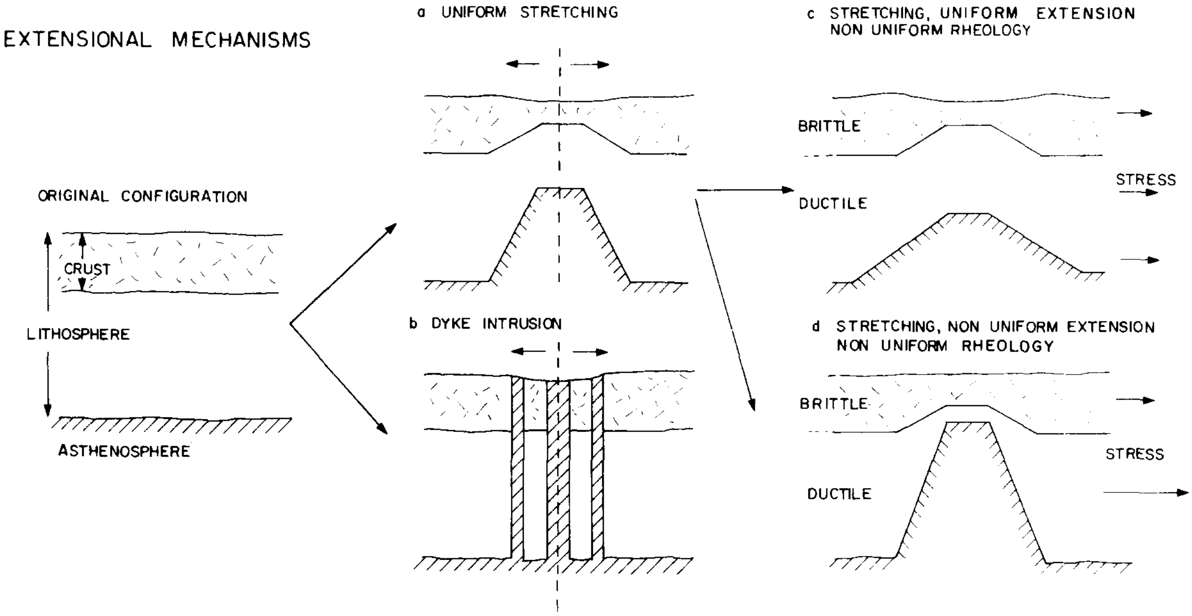


Fig. 1. Various extensional models for the creation of the initial thermal anomaly which can account for the subsidence and heat flow of intracontinental basins. (a) and (b) are uniform stretching [10] and dike intrusion [11]. (c) and (d) are modifications of the stretching model for a non-uniform rheology. In model (c), the ductile lithosphere thins over a wider zone than the rigid crust, although net extension is the same. In model (d), net extension is greater in the lithosphere than in the crust.

[11,13]. Replacement of light crustal rocks by denser ultrabasic or basaltic material also results in initial isostatic subsidence for crustal thickness greater than 20 km (Fig. 2a). It is unlikely that this mechanism can operate at depth where the lithosphere is thought to deform by ductile flow and we do not believe that this model has widespread geological applicability [12]. On the other hand, dyke intrusion is observed in some of the smaller basins within the Carpathians. As we believe these dikes have a significant effect on the local heat flow and subsidence, we included the model in this section and work out the simple theory in the Appendix.

2.3. *Subcrustal attenuation*

Attenuation of subcrustal lithosphere without significant extension or deformation of the crust results in initial isostatic uplift. If a sufficient thickness of the crust is involved in the attenuation process, initial isostatic subsidence rather than uplift may occur (Fig. 2a). This attenuation could be the result of either subcrustal extension (Fig. 1d) or of subcrustal erosion and melting of the lithosphere. Such a mechanism may be supported by observations in the Basin and Range in the Western United States where 30 km crust is underlain by a thin subcrustal lithosphere [14,15]. Further evidence comes from drill holes and seismic refraction information across the Labrador Shelf. Royden and Keen [16] have shown that the thermal subsidence can only be explained by stretching significantly greater than that consistent with the observed decrease in crustal thickness.

The general equations governing initial subsidence or uplift, and the thermal subsidence and heat flow after the initial deformation are presented in the Appendix. In each model the extension or thermal event is assumed to occur instantaneously. Jarvis and McKenzie [17] have shown that only extensional events of long duration are likely to have a significant effect on these simple calculations. In the region of this study the extensional and other events are of short duration ($\leqslant 5$ m.y.). For each of the three models we have examined the initial and thermal subsidence and the heat flow as a function of time for two different amounts of heat input (Fig. 2a, b and c). Note that the initial subsidence in the cases of uniform stretching and dike intrusion is the same for equivalent heat input. However, except when a large volume of crustal material is involved in the deformation, the subcrustal extension (or erosion) model (2″ and 4″, Fig. 2c) more often results in an initial uplift rather than in an initial subsidence. Uniform stretching and subcrustal thinning give roughly the same thermal subsidence curves for the same amount of heat input. The dike intrusion model yields a much faster early subsidence because heat is preferentially distributed near the surface. These differences are clearly illustrated by the early heat flow decay: the dike intrusion model shows a very high initial heat flow followed by a rapid decay, the uniform stretching model shows a lower initial heat flow followed by a less rapid decay, and the subcrustal thinning shows an early increase, a point of maximum heat flow, and then a decay which closely parallels the uniform stretching model. About 15 Ma after the extensional or thermal events have ceased all three models show approximately the same heat flow. Thus, for basins older than 10 Ma, it is not possible to use heat flow to distinguish among these models for basin formation.

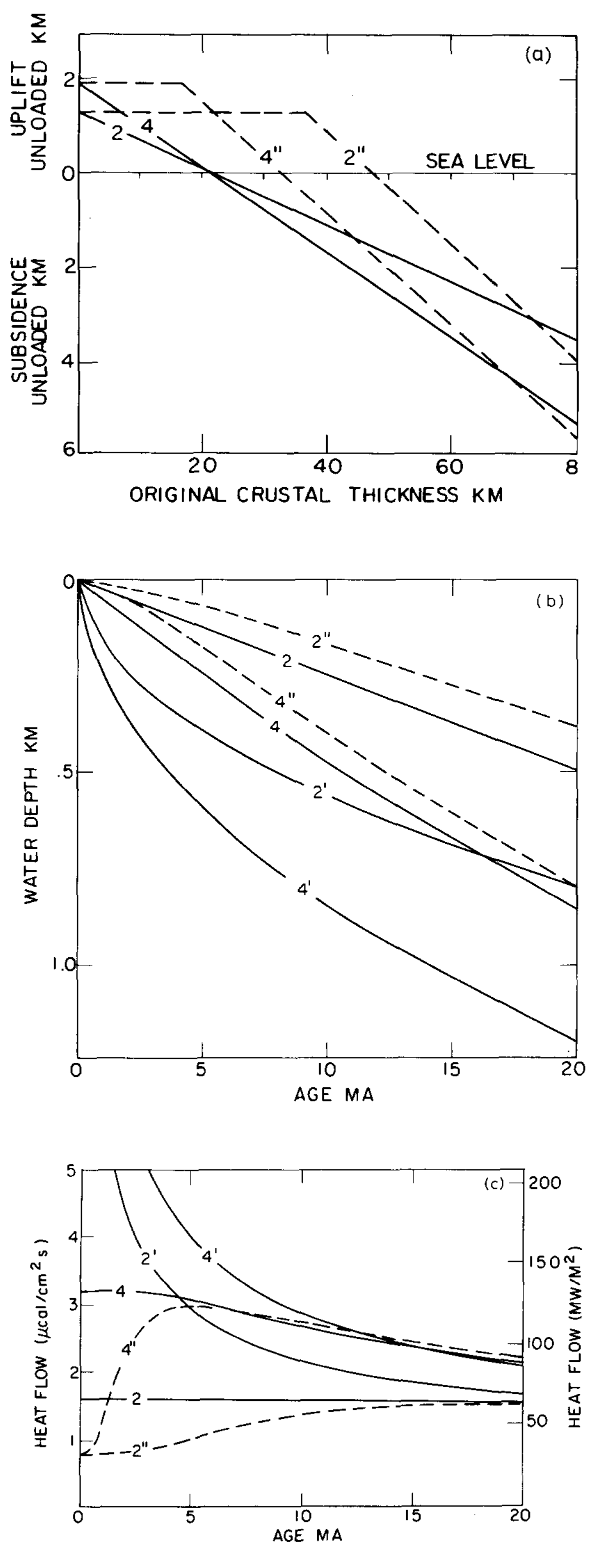


Fig. 2. (a) Initial isostatic subsidence (unloaded) as a function of crustal thickness for various models of extension. The solid lines represent both the stretching and dike instrusion models, the dashed lines represent the subcrustal attenuation model. For the uniform stretching and dike intrusion models the subsidence curves represent extension by a factor of 2 and 4, respectively. For the subcrustal attenuation, the subsidence curves correspond to the same initial heat input as simple stretching by a factor of 2 or 4. For the purposes of calculation we have assumed that subcrustal attenuation is the result of eroding the base of the lithosphere. $\beta = 2''$ and $4''$ correspond to eroding to depth of 36 and 17 km, respectively. (b) Calculated thermal subsidence/water loaded as a function of time corresponding to the models considered in (a). The solid curves are uniform stretching (2 and 4) and dike intrusion (2′ and 4′) and the dashed curves are subcrustal attenuation (2″ and 4″). (c) Calculated heat flow as a function of time, corresponding to the models in (a) and (b). The solid lines are simple stretching and dike intrusion, the dashed lines are subcrustal attenuation.

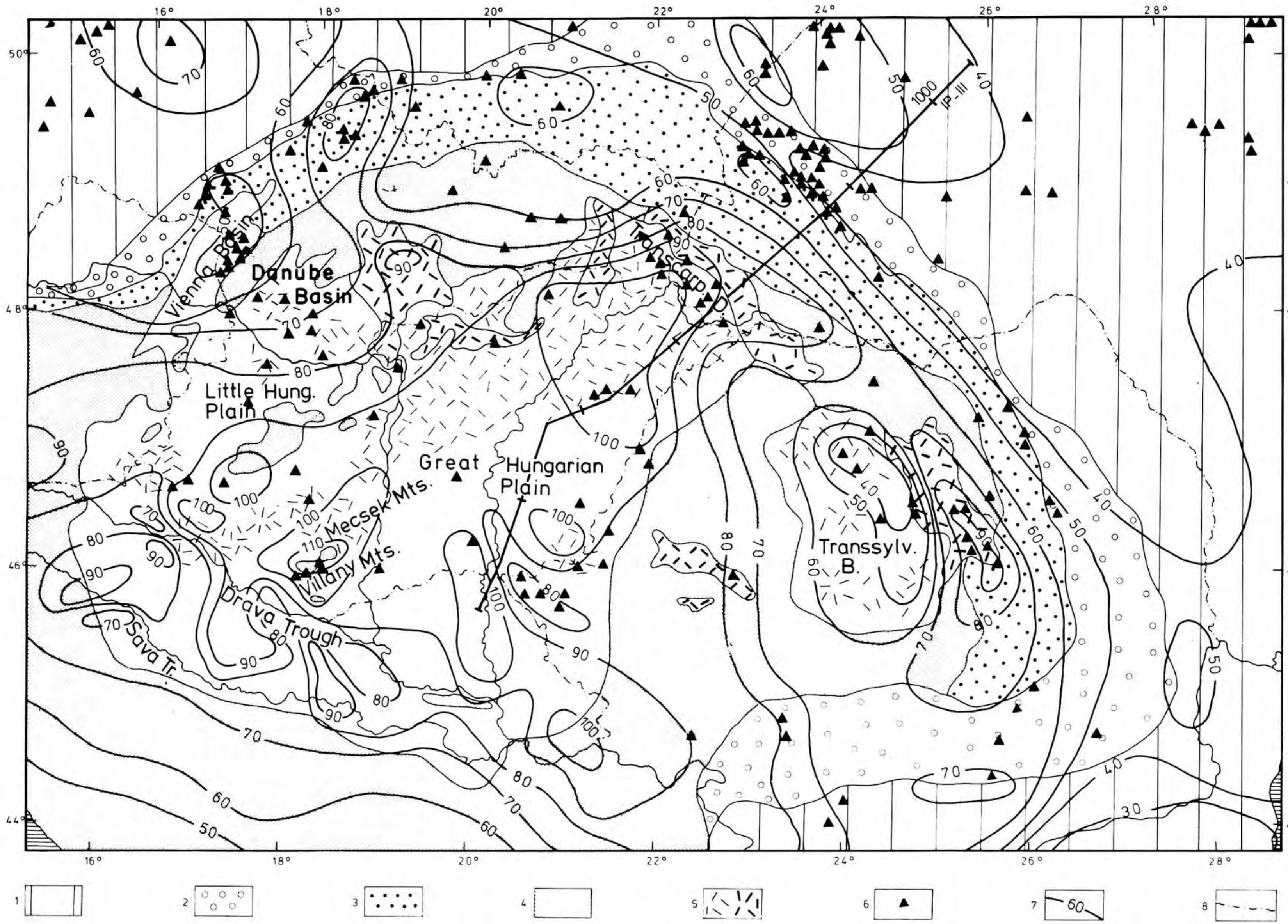


Fig. 3. Tectonic sketch and heat flow map of the Pannonian Basin and the surrounding Carpathian arc. The position of the heat flow stations and the seismic refraction profile discussed in the text are shown. Key: 1 = European foreland; 2 = Oligocene to Pliocene molasse foredeep; 3 = Upper Cretaceous, and Paleogene Flysch Nappes; 4 = Late Mesozoic (mainly Upper Cretaceous) nappe system of the eastern Alps, inner Carpathians, Apuseni Mountains and Dinarides; 5 = Mio-Pliocene calc-alkaline volcanics (covered and outcropping respectively); 6 = Neogene through Quaternary basins with heat flow stations; 7 = heat flow isolines, in mW/m^2; 8 =

3. Geology, tectonic setting and basin development

The Carpathian arc can be divided into three units of different origin: the Western, Eastern and Southern Carpathians. The Western and Eastern Carpathians can be further subdivided into two major belts: an outer and inner belt. The outer, morphologically continuous belt is a pile of large thrust sheets composed of Cretaceous/Paleogene flysch which overrode the foredeep molasse during the Early and Middle Miocene (Fig. 3). The inner belt of the Carpathian arc, comprised mainly of Mesozoic and older rocks, is discontinuous: large areas are subsided, others are covered by Neogene volcanics. The Inner Western Carpathians are composed of north vergent nappes which came into existence during mid-Cretaceous tectonic events. The Inner East Carpathians were deformed into northeast and east vergent nappes during the mid-Cretaceous and latest Cretaceous/Paleocene tectonic phases. The sourth vergent nappe structure of the Southern Carpathians was completed during the latest Cretaceous/Paleocene events. The basin surrounded by the folded arc is not uniform. The Neogene/Quaternary subsidence only affected certain areas, leaving some ranges emergent and/or uplifted. These ranges divide the back-arc area into several sub-basins (Fig. 4). Of these basins, we shall call the Vienna Basin, Transcarpathian Depression, West Danube Basin and the Transylvanian Basin *peripheral* basins as they are situated at the margin of the *central* Pannonian Basin.

The basement of the peripheral basins, Danube

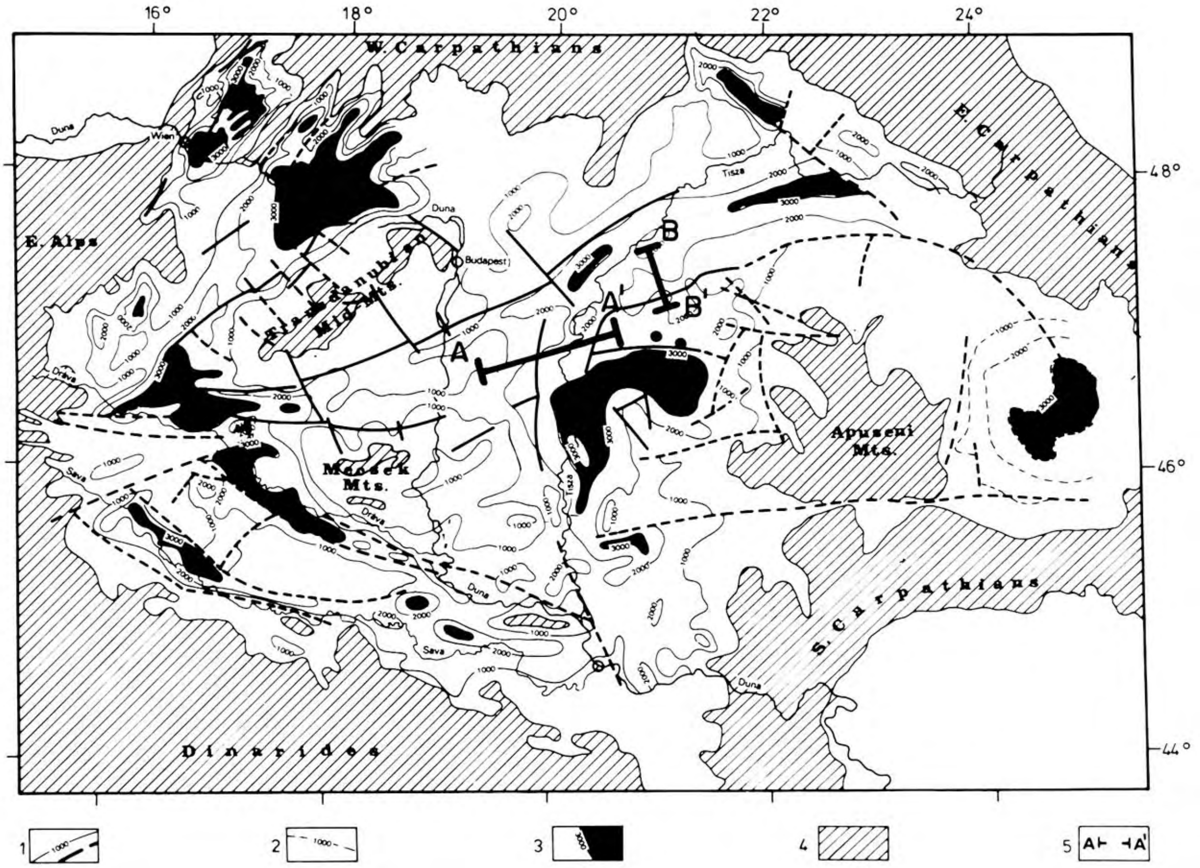


Fig. 4. Neogene-Quaternary sedimentary thickness of the intra-Carpathian basins (modified after Mahel [43]). Key: 1 = isoline of the thickness of Neogene and Quaternary deposits and main faults; 2 = isoline of the thickness of the post early Miocene deposits in the Transylvanian Basin; 3 = deep depressions; 4 = outcropping pre-Neogene rocks; 5 = profiles of wells discussed in Fig. 6a, b. Solid dots show location of magnetic dating of sediments in boreholes.

Lowland and Little Hungarian Plain, is made up from the subsided blocks of the mountain arc, and their nappe structure is confirmed by deep borings and geophysical data [18–21]. The structure of the pre-Neogene bedrock of the Great Hungarian Plain is still under discussion. Though no extensive thrust sheets have been demonstrated by drilling, the pre-Neogene basement is known to be composed of belts which can be correlated with nappe units of the Apuseni Mountains and Inner Eastern Carpathians [22,23]. Thus we believe that much of the pre-Neogene basement of the intra-Carpathian depressions was part of a Late Cretaceous/Paleogene orogenic belt before it became involved in general extension and subsidence during the Neogene. The present block-like structure is the result of this neotectonic development.

The existence and significance of tectonic boundaries within the intra-Carpathian region is not generally agreed upon. Facies and faunal data, and structural connections suggest that the Inner West Carpathians and the western part of the Pannonian basement are continuous with the Austroalpine and South Alpine units to the west. While it has been suggested that the Inner East Carpathians, Apuseni Mountains and the southeastern part of the Pannonian basin have European affinities [24–26], the derivation of these regions and their relation to western terranes is controversial and uncertain. Anyway, the supposed large-scale movements of these regions occurred during the Late Cretaceous/Paleogene, and the beginning of Miocene saw a single, but certainly not immobile, Pannonian unit.

Sedimentological analysis has shown that the Carpathian flysch was deposited on continental rise and partly on deep-sea plains [27], and the deepest parts of the flysch basin were probably floored by oceanic crust [28–30]. The subduction of the oceanic crust went on, possibly with intermission, during the latest Cretaceous/Paleogene and was directed toward the Pannonian center [26,29]. Thrusting of the flysch nappes onto each other and their transport, on the foredeep molasse during the Early and Middle Miocene, indicate continental collision. The European foreland was overriden at least 35–60 km by the Outer Carpathian nappes [31] and the onset of extension and subsidence of the intra-Carpathian basins are synchronous with the continental-continental collision in the arc.

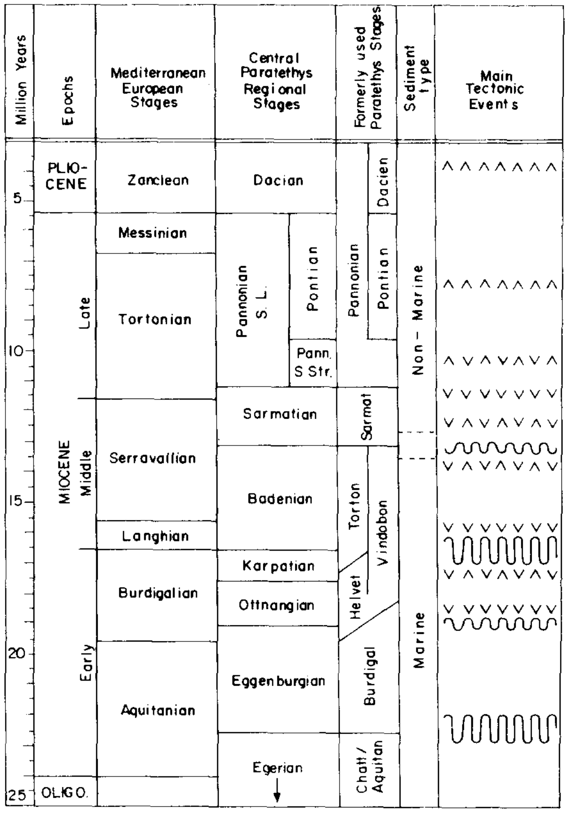


Fig. 5. Comparison of world-wide and regional biostratigraphical time scales. Also shown are the sediment type and the major episodes of compression (~~~~) and rhyolitic (V) and andesitic (Λ) volcanism within the Carpathian system.

Because of the isolation of the Paratethys from the Mediterranean during the late Middle Miocene, there is some difficulty in correlating the dating on regional stages in the Carpathians with that of the better known Mediterranean stages. For simplicity and consistency in our analysis we have adjusted all ages and dates to the time scale developed by Rögl et al. [32], as modified by F.F. Steininger (personal communication) (Fig. 5).

During the Late Oligocene, the central Paratethys covered southeastern Europe and parts of the slightly deformed flysch basin. At this time a large part of the central Pannonian region was emergent and subject to erosion, while the surrounding areas remained below sea level [33]. Beginning in the Eggenburgian folding and thrusting of the flysch complex was associated with intense vertical movements. Initiation of rapid subsidence in the periphery of the Pannonian region led to marine sedimentation in the Vienna Basin and Transcarpathian Depression [20,34], while a large part of the central Pannonian region remained uplifted until the Badenian. The Neogene calc-alkaline volcanism began in this central region during the Ottnangian and continued sporadically until the Pannonian (s.s.) [35–37]. Here block faulting, graben formation and marine sedimentation began in the Karpatian, and rhyolites were erupted at several places including the Danube Basin and Transcarpathian Depression (Fig. 5). The tectonic activity reached its climax during the Badenian. Continuing compression gave rise to the overthrusting of the flysch nappes onto the foredeep molasse. The fill of the Vienna Basin was also deformed. In the early Badenian subsidence started which affected nearly the whole Carpatho-Pannonian area. Synsedimentary faulting was very active in the Vienna Basin, Danube Basin and Transcarpathian Depression. Andesitic lavas were erupted in northern Hungary, central Slovakia, and in the Apuseni Mountains. Rhyolites and their pyroclastics constitute beds of regional extent in the Badenian sediments of the Danube Basin, Transcarpathian Depression and Transylvanian Basin. This general period of extension in the back-arc area was interrupted by periods of slight compression, shown locally by gentle folding of the basin fill [38]. Reverse faulting of this age is evidenced to the south of the Transdanubian Mid-Mountains [39].

During the Sarmatian the whole Carpathian "fire-belt" except the easternmost range, was active [40]. Andesites, dacites and rhyolites were erupted. At this time, the very fast subsidence of the peripheral basin finished. The central, Pannonian area was characterized by rapid subsidence which was accompanied by the eruption of rhyolites on the northeastern part of the basin. The Transdanubian Mid-Mountains and Mecsek Mountains started to emerge and elsewhere new areas subsided below sea level.

In the central region the transgression became general during the Pannonian. At about the same time the rate of subsidence in the peripheral depression drastically decreased and the uplift of the present-day Carpathian arc began.

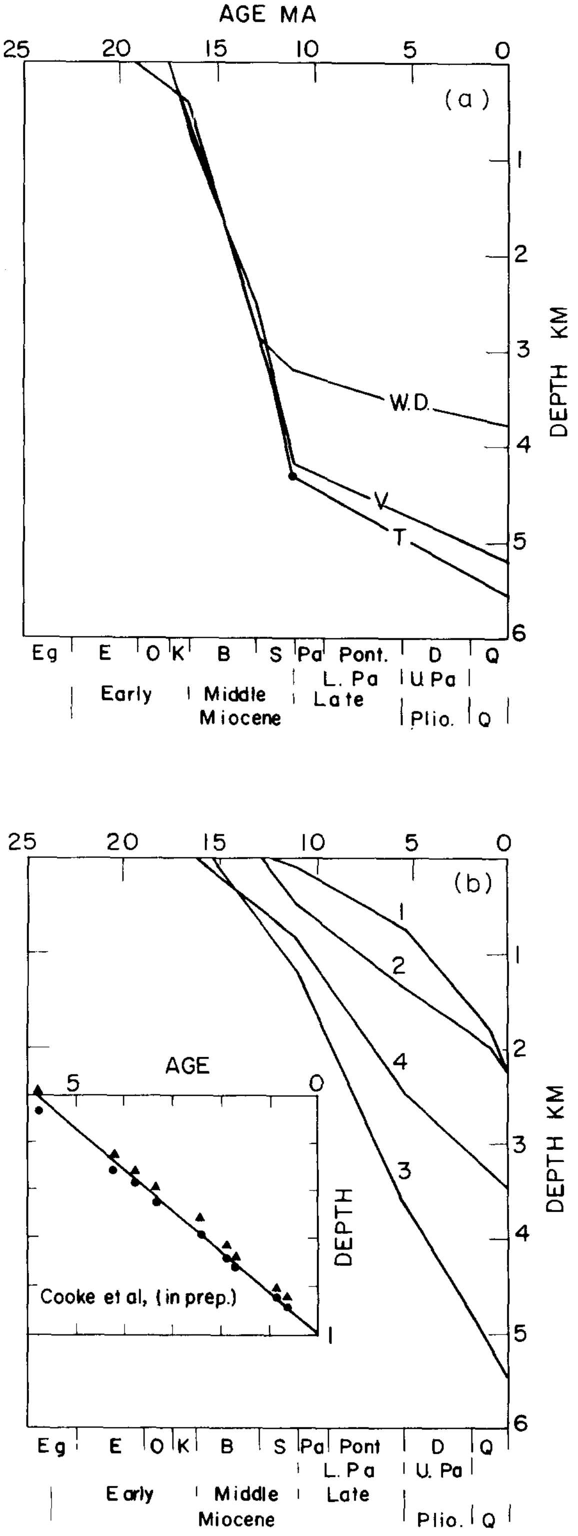
Subsidence and lacustrine-fluvial sedimentation has continued during the Pliocene and Quaternary in the central part of the Little Hungarian Plain and over a large part of the Great Hungarian Plain [41]. The Carpathian arc, Apuseni Mountains and the Hungarian Mid-Mountains have remained emergent.

4. Basin subsidence

In this preliminary analysis of the Carpathian region we concentrate on the subsidence of the Neogene basins (Fig. 4). We begin with a discussion of subsidence in the peripheral basins, whose development is in marked contrast to that of the central basins. Although we shall only discuss the subsidence history of three of these, the Vienna and West Danube Basins and the Transcarpathian Depression, the general pattern of subsidence in the Transylvania Basin is similar.

From the Egerian to Ottnangian increased tectonic mobility, subsidence and uplift can be observed in the three basins. After this the subsidence in these basins may be divided into two distinct phases. The first of these begins in the Ottnangian (Vienna Basin) or the Karpatian (West Danube Basin and Transcarpathian Depression) and continues until the end of the Sarmatian. This phase is characterized by rapid subsidence. Initially deep-water deposits dominated but by the end of the Sarmatian the whole depositional environment became shallow water with a total sediment accumulation of more than 4 km. During the second phase of subsidence, from the end of the Sarmatian to present, the subsidence rate is considerably reduced and the sedimentary environment is generally shallow water. This second phase of subsidence appears to be linear and remarkably uniform from basin to basin (Fig. 6a).

In contrast, the subsidence of the Little and Great Hungarian Basins consists of a single phase after an early period of tectonic unrest (Ottnangian/Karpatian). In these basins, the major onset of sedimentation did not begin until the late Badenian or early Sarmatian. This single phase has continued until the present, and the sedimentary environment has been generally shallow water (Fig. 6b). Although the amount of subsidence varies across the basin, the rate of subsidence is remarkably consistent. The mean subsidence for the Great Hungarian Basin since the Sarmatian is approximately 1700 m. This is almost double that in the peripheral basins during the same time period.



Recent paleomagnetic analysis of sediments recovered from two wells in the center of the Great Hungarian Plain, show a linear sedimentation/subsidence history for the last 6 Ma (Fig. 6b) (H.B.S. Cooke, personal communication). These data consist of approximately six hundred measurements for each well. The calculated subsidence rates are in good agreement with subsidence rates determined from paleontological dating and those estimated from fluvial terraces [41]. Furthermore, these data suggest that the cyclic nature of sedimentation is primarily the result of climatic variations and/or paleocurrents, rather than major changes in overall subsidence rate. Recent geodetic leveling measurements in Hungary also yield comparable subsidence rates.

The rate and amplitude of basement subsidence is easily measured and is directly related to the manner in which the basin was created. It is possible to use these observations to distinguish between the various explanations of basin formation. For this purpose we examined in detail the three wells from the peripheral basins discussed in the previous section (Fig. 6a) and twelve wells from the Great Hungarian Plain, the largest of the central Carpathian basins.

Our choice of wells in the peripheral basins is restricted since, apart from the Vienna Basin, there are few lithologic logs or complete cross sections available. Fortunately there are many drill holes in the central Carpathian basins from which subsidence data can be gathered and maps showing the sites and the overall thickness of sediments have been published in the Hungarian literature [44]. However, few profiles of wells with sufficient detail have been presented. Thus even in these basins, our choice of wells is limited. We have selected two profiles presented by Körössy [45] from the center of the Great Hungarian

Fig. 6. (a) Subsidence of basement as inferred from a well in the center of the Vienna Basin (*V*) [42], the Transcarpathian Depression (*T*) [34] and West Danube Basin (*W.D.*) [43]. The symbols represent the regional biostratigraphic zonation. (b) Subsidence as inferred from three wells in the great Hungarian Basin (*1*–3) and one bore hole in the Little Hungarian Plain (*4*). The symbols are the same as for (a). The inset shows on a proportionally exaggerated scale ages from paleomagnetic dating plotted against depth in the bore holes at Devavanya and Vesztö from H.B.S. Cooke (personal communication).

Plain for our analysis. These holes have been selected because all the stratigraphic boundaries, except the Quaternary on the lower profile, are documented. The position of the holes is known and it is possible to determine the thickness of the Quaternary from another map [46]. We selected twelve wells, six from the eastern profile and six from the north-south profile in Fig. 4.

In order to compute the depth of the basement through time it is necessary to allow for compaction, remove the sediment load, adjust for water depth at the time of deposition and add eustatic sea level changes [3]. All the wells were corrected for compaction using an exponential porosity/depth relation:

$$f = f_0 e^{-cz}$$

where f_0 and c were 0.42 and 0.31×10^{-5} cm, respectively. These figures were calculated from the density versus depth plots of Stegena [47] assuming a sediment grain density of 2.85 g/cm^3. The basic principles and the method used are discussed at length in Sclater and Christie [12]. For this analysis we ignored the effect of eustatic sea level and have assumed that the depth of burial is close to sea level. Apart from the early Badenian in the Vienna Basin where deep-water marine fauna have been reported this is a reasonable assumption. Endemic molluscs are found from the Sarmatian onwards in the Carpathian basins and are a good indicator of a shallow-water environment. Except for the Transylvanian Basin where there has been Quaternary uplift, all the basins are currently within 200 m of sea level.

When corrected for sediment load (Table 1) the three peripheral basins still retain the remarkably sharp initial drop in basement depth followed by a slow linear increase shown by the uncorrected plots of sediment thickness versus time (compare Figs. 6a and 7). The large amount of early subsidence in the Transcarpathian Depression may not be significant. For this basin we followed the cross section given in Mahel [43, p. 123] and assumed basement as base Karpatian. This could well be in error as below the Karpatian there is an absence of sediment (Ottnangian uplift) followed by several hundred meters of Eggenburgian strata [34]. If the compaction of this were considered the initial subsidence would be much

TABLE 1

Subsidence analysis of three peripheral basins

Basin	Horizon	Age (Ma)	Depth (m)	Sediment thickness above basement (m)		ρ (g/cm^3)	Unloaded depth (m)	
				observed	corrected		1	2
Vienna	surface	0	0	7200	7200	2.53	2482	1302
	Pannonian	11	1000	6200	6544	2.52	2319	1139
	Badenian	16.5	4800	2400	3107	2.34	1333	153
	Ottnangian	19	5200	2000	2667	2.31	1180	–
	basement	–	7200	–	–	–	–	–
Transcarpathian	Upper Pannonian	6	0	5514	5514	2.47	2052	
	Pannonian	11	1103	4411	4755	2.44	1843	
	Sarmatian	13	2014	2500	3071	2.34	1321	
	Badenian	16.5	4749	765	1111	2.18	557	
	Karpatian	17.5	5514	0	0	–	–	
West Danube	surface	0	0	3782	3782	2.39	1553	
	Pontian	6	232	3550	3627	2.38	1504	
	Pannonian	11	582	3200	3377	2.36	1423	
	Sarmatian	13	873	2909	3152	2.34	1349	
	Badenian	16.5	3782	–	–	–	–	

1 = basement subsidence, 2 = basement subsidence since Ottnangian.

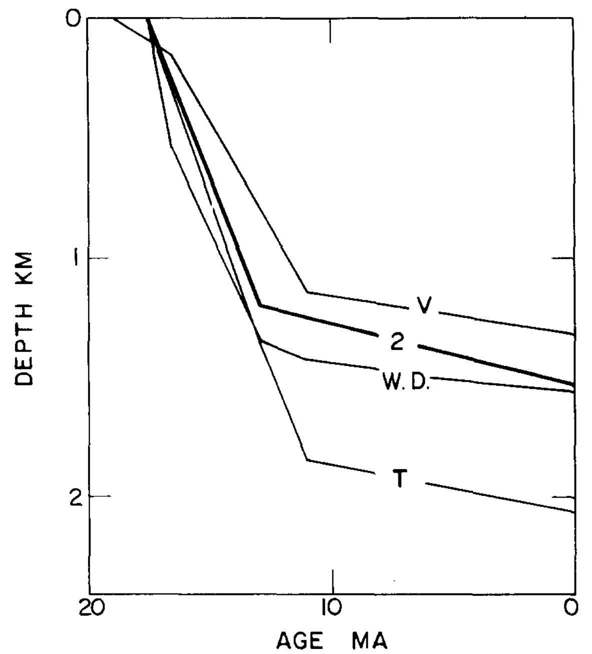


Fig. 7. Comparison of the inferred water loaded basement subsidence for the three peripheral basins from Table 1 with stretching by a factor of 2 (β = 2) of 40 km thick continental crust. The stretching was assumed to occur in the Badenian (16.5–13 Ma).

reduced. Even given these differences the structure of the profiles from each of the three basins show exactly the same initial sharp drop of about 1.5 km followed by a much slower linear subsidence totaling 200–300 m from the Sarmatian to present (Fig. 7).

The subsidence information for the 12 wells on the Great Hungarian Plain is presented in Table 2. Though the total depth to base Sarmatian (13 Ma) of the wells varies from about 1100 to 2400 m the subsidence is fairly linear. When corrected for compaction and loading the basement subsidence remains linear but reduces to between 600 and 1050 m. Because of the porosity of the sediments changes only slowly with depth, compaction has little effect on the thickness of the sedimentary layers when basement is assumed to be base Neogene. However, there is sediment below the base Neogene in parts of the basin and as it also compacts on burial the effect must be considered. To estimate the range of this effect we chose basement at 2 km below the base Neogene as the maximum realistic contributon to the subsidence from compaction of the underlying sediments. When we recomputed each profile with the basement at this depth, we found the overall basement subsidence to be reduced to between 300 and 600 m.

To simplify the data anlysis and to permit easy comparison with theoretical models we averaged the subsidence data and calculated basement depth as a function of age assuming, first, no sediment, and then, 2 km of sediment below the base Sarmatian (Table 3). The resulting profiles cover the likely range of the mean subsidence of the twelve wells. To generalize these results we compared these two profiles with the averages depths to the base Quater-

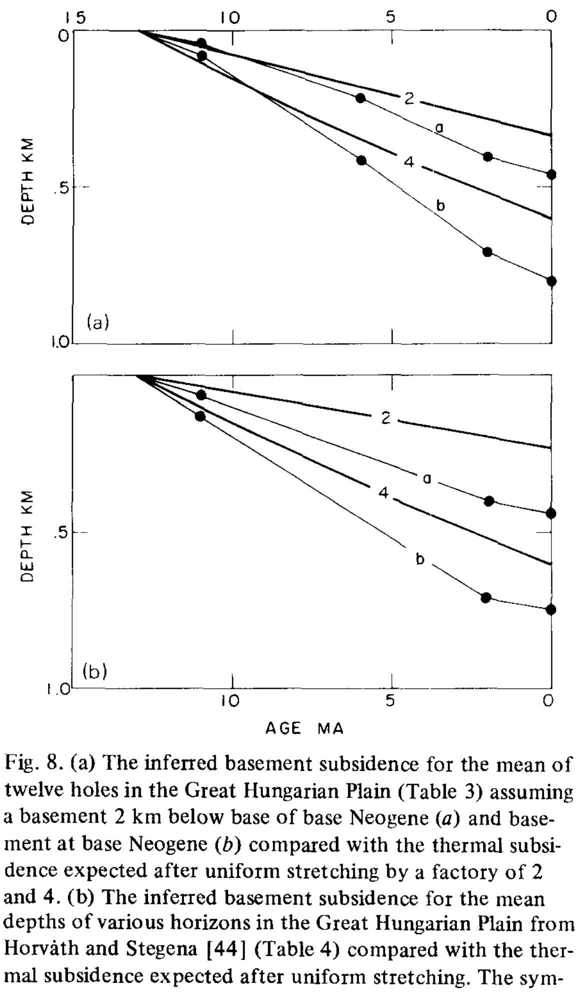


Fig. 8. (a) The inferred basement subsidence for the mean of twelve holes in the Great Hungarian Plain (Table 3) assuming a basement 2 km below base of base Neogene (*a*) and basement at base Neogene (*b*) compared with the thermal subsidence expected after uniform stretching by a factory of 2 and 4. (b) The inferred basement subsidence for the mean depths of various horizons in the Great Hungarian Plain from Horvàth and Stegena [44] (Table 4) compared with the thermal subsidence expected after uniform stretching. The symbols are the same as in (a).

TABLE 2

Subsidence data from wells along profiles AA' and BB' (Fig. 4)

Well	Horizon	Age (Ma)	Depth (m)	Sediment thickness above basement (m)				ρ (g/cm³)		Unloaded depth (m)		
				observed		corrected		1	2	1	2	3
				1	2	1	2					
Subsidence data from six wells along profile AA'												
Kecskemét W.1	sediment surface	0	–	2316	4316	2316	4316	2.29	2.43	1052	1715	618
	Pleistocene	2	263	2053	4053	2122	4144	2.26	2.41	980	1664	547
	Upper Pannonian	6	972	1344	3344	1522	3626	2.21	2.37	738	1504	407
	Lower Pannonian	11	1916	400	2400	522	2819	2.11	2.32	276	1234	137
	Sarmatian	13	2316	–	2000	–	2434	–	2.29	–	1097	–
	basement	–	4316	–	–	–	–	–	–	–	–	–
Kecskemét 1	surface	0	–	1156	3156	1156	3156	2.18	2.35	578	1352	321
	Pleistocene	2	270	886	2886	927	2925	2.16	2.33	472	1286	255
	Upper Pannonian	6	815	341	2341	395	2550	2.11	2.30	211	1139	108
	Lower Pannonian	11	1116	40	2040	49	2294	2.05	2.28	27	1043	12
	Sarmatian	13	1156	–	2000	–	2259	–	2.28	–	1031	–
	basement	–	3156	–	–	–	–	–	–	–	–	–
Kecskemét 3	surface	0	0	1291	3291	1291	3291	2.19	2.35	637	1395	355
	Pleistocene	2	272	1019	3019	1066	3104	2.17	2.34	537	1333	293
	Upper Pannonian	6	842	449	2449	519	2668	2.11	2.31	275	1182	142
	Lower Pannonian	11	1091	200	2200	243	2459	2.08	2.29	132	1106	66
	Sarmatian	13	1291	–	2000	–	2282	–	2.28	–	1040	–
	basement	–	3291	–	–	–	–	–	–	–	–	–
Jászkarajenö-1	surface	0	0	1496	3496	1496	3496	2.21	2.36	726	1462	409
	Pleistocene	2	360	1136	3136	1201	3248	2.19	2.35	547	1381	328
	Upper Pannonian	6	1110	386	2386	464	2659	2.11	2.31	247	1179	126
	Lower Pannonian	11	1420	76	2076	97	2387	2.06	2.29	54	1079	26
	Sarmatian	13	1496	–	2000	–	2317	–	2.28	–	1053	–
	basement	–	3496	–	–	–	–	–	–	–	–	–
Rakoczifalva-1	surface	–	–	1483	3483	1483	3483	2.21	2.37	720	1458	406
	Pleistocene	2	250	1233	3233	1282	3314	2.19	2.36	633	1403	351
	Upper Pannonian	6	920	563	2563	654	2804	2.13	2.32	341	1230	178
	Lower Pannonian	11	1425	58	2058	74	2368	2.06	2.28	41	1072	20
	Sarmatian	13	1483	–	2000	–	2315	–	2.28	–	1052	–
	basement	–	3483	–	–	–	–	–	–	–	–	–

Turkeve-1	surface	–	–	2218	4218	2218	4218	2.27	2.41	1016	1686	594
	Pleistocene	2	300	1918	3918	1994	4021	2.26	2.39	929	1627	535
	Upper Pannonian	6	1349	869	2869	1041	3215	2.17	2.35	525	1369	277
	Lower Pannonian	11	2142	76	2076	105	2496	2.06	2.30	57	1120	28
	Sarmatian	13	2218	–	2000	–	2422	–	2.29	–	1092	–
	basement	–	4218	–	–	–	–	–	–	–	–	–
Subsidence data from six wells along BB'												
Biharnagybajom-27	surface	–	0	1100	3100	1100	3100	2.18	2.34	552	1331	304
	Pleistocene	2	200	900	2900	931	2962	2.16	2.33	474	1284	257
	Upper Pannonian	6	650	450	2450	505	2623	2.11	2.31	268	1165	122
	Lower Pannonian	11	1050	50	2050	61	2292	2.06	2.28	34	1043	16
	Sarmatian	13	1100	–	2000	–	2248	–	2.28	–	1027	–
	basement	–	3100	–	–	–	–	–	–	–	–	–
Püspökladány-1	surface	0	0	1939	3939	1939	3939	2.25	2.39	910	1602	524
	Pleistocene	2	300	1639	3639	1708	3739	2.23	2.38	815	1539	461
	Upper Pannonian	6	1030	909	2909	1053	3190	2.17	2.35	531	1362	284
	Lower Pannonian	11	1839	100	2100	134	2480	2.07	2.30	73	1114	36
	Sarmatian	13	1939	–	2000	–	2388	–	2.29	–	1078	–
	basement	–	3939	–	–	–	–	–	–	–	–	–
Nádudvar-6	surface	0	0	1715	3715	1715	3715	2.23	2.38	817	1532	466
	Pleistocene	2	200	1515	3515	1560	3582	2.22	2.37	754	1490	424
	Upper Pannonian	6	985	730	2730	847	2992	2.15	2.34	435	1294	228
	Lower Pannonian	11	1688	27	2027	36	2377	2.05	2.29	20	1076	10
	Sarmatian	13	1715	–	2000	–	2352	–	2.29	–	1066	–
	basement	–	3715	–	–	–	–	–	–	–	–	–
Nádudvar-11	surface	0	0	1696	3696	1696	3696	2.23	2.38	809	1526	461
	Pleistocene	2	200	1496	3496	1541	3563	2.22	2.37	745	1484	419
	Upper Pannonian	6	912	784	2784	890	3024	2.16	2.34	460	1296	231
	Lower Pannonian	11	1609	87	2087	114	2430	2.06	2.29	63	1096	31
	Sarmatian	13	1696	–	2000	–	2349	–	2.28	–	1065	–
	basement	–	3696	–	–	–	–	–	–	–	–	–
Nádudvar-7	surface	0	0	1931	3931	1931	3931	2.25	2.39	904	1599	521
	Pleistocene	2	200	1731	3731	1780	3780	2.24	2.39	844	1559	481
	Upper Pannonian	6	985	946	2946	1088	3217	2.17	2.35	547	1370	292
	Lower Pannonian	11	1778	153	2153	202	2528	2.07	2.30	110	1131	53
	Sarmatian	13	1931	–	2000	–	2383	–	2.29	–	1078	–
	basement	–	3931	–	–	–	–	–	–	–	–	–
Nagyiván 1	surface	–	0	1812	3812	1812	3812	2.24	2.39	857	1562	490
	Pleistocene	2	300	1512	3512	1579	3612	2.22	2.37	762	1500	428
	Upper Pannonian	6	1247	565	2565	682	2876	2.13	2.33	356	1255	148
	Lower Pannonian	11	1712	100	2100	132	2460	2.07	2.30	72	1107	35
	Sarmatian	13	1812	–	2000	–	2366	–	2.29	–	1072	–
	basement	–	3812	–	–	–	–	–	–	–	–	–

1 = Subsidence assuming basement at base Sarmatian, 2 = Subsidence assuming basement 2 km below base Sarmation, 3 = Subsidence since the Badenian assuming basement 2 km below base Sarmatian.

TABLE 3

Mean subsidence data for twelve wells in the Great Hungarian Plain

Horizon	Age (Ma)	Depth (m)	Sediment thickness above Sarmatian (m)	Unloaded depth (m)	
				1	2
Surface	0	0	1679 (1100–2316)	798 (552–1052)	456 (304–618)
Pleistocene	2	259 (200– 360)	1426 (900–2053)	712 (472– 980)	400 (255–567)
Upper Pannonian	6	960 (650–1149)	695 (341–1344)	410 (211– 238)	212 (108–407)
Lower Pannonian	11	1566 (1050–2142)	114 (27– 400)	80 (20– 276)	39 (10–157)
Sarmatian	13	1679 (1100–2316)	–	–	–

Numbers between brackets indicate range of values. 1 = depth computed assuming basement at base Sarmatian, 2 = depth computed assuming basement 2 km below base Sarmatian.

nary, Pannonian and Sarmatian from Horvàth and Stegena [44] when corrected for compaction and sediment load (Table 4). The mean of the twelve wells and the average data from all wells in the central basins are remarkably similar showing a steady basement subsidence of between 400 and 700 m since the Sarmatian (Fig. 8a, b). This steady subsidence of the central basins contrasts strongly with the very rapid initial and then slow later subsidence of the peripheral basins (Fig. 7).

5. Other geophysical data

Any viable thermal model for the formation and evolution of the intra-Carpathian basins must be consistent with the geophysical observations. In this region these consist of heat flow, electrical conductivity, crustal seismics, magnetics, and gravity. Horvàth et al. [48] have recently completed a re-analysis of the heat flow measurements in Hungary, and these have been added to the heat flow data for the surrounding regions (Fig. 3). There is a significant contrast between the heat flow through the peripheral basins, excluding the Transcarpathian Depression, and that through the central Great Hungarian Plain. The former is low, 1.2 μcal/cm^2 s (50 mW/m^2) while the latter is high, about 2.4 μcal/cm^2 s (100 mW/m^2). The Danube Basin and Little Hungarian Plain are transitional and are characterized by nearly normal values of 1.7 μcal/cm^2 s (70 mW/m^2). Based on the most reliable heat flow measurements in the Great

TABLE 4

Mean subsidence data for the Great Hungarian Plain from Horvàth et al. [26]

Horizon	Age (Ma)	Depth (m)	Sediment thickness above basement (m)				ρ(g/cm^3)		Unloaded depth (m)		
			observed		corrected		1	2	1	2	3
			1	2	1	2					
Surface	0	0	1617	3617	1617	3617	2.23	2.38	756	1500	439
Quaternary	2	183	1434	3434	1473	3494	2.12	2.37	716	1461	400
Lower Pannonian	11	1423	194	2194	245	2513	2.07	2.30	133	1126	65
Sarmatian	13	1617	–	200	–	2336	–	2.29	–	1061	–
Basement	–	3617	–	–	–	–	–	–	–	–	–

1 = assuming basement at base Sarmatian, 2 = assuming basement 2 km below base Sarmatian, 3 = unloaded depth in the case of 2.

Hungarian Plain, we have computed a mean heat flow of 2.5 μcal/cm^2 s (110 mW/m^2), considerably higher than the average heat flow for the continents of 1.5 μcal/cm^2 s (63 mW/cm^2). We think it likely that the slightly high heat flow observed in the Carpathians and other areas not affected by subsidence, 1.7–2.0 μcal/cm^2 s (70–80 mW/m^2), is due to erosion. This value is average for recent orogenic belts. In contrast, the heat flow in the basins, which have clearly not been subjected to erosion, must be due to increased temperatures within the lithosphere.

Many seismic refraction lines have been shot in the vicinity of the Carpathians. These include individual profiles in localized areas such as the Vienna and Danube Basins and long composite lines crossing both basins, mountains and shields. The individual refraction profiles in the Vienna, West Danube and Transcarpathian Basins have revealed Moho depths of 26–32 km [49]. The most extensive of the composite profiles is the IP-III line which has been combined with the geological observations to produce a cross section of the crust from the Ukrainian Platform to the Great Hungarian Plain (Fig. 9). Note that while the depth of the crust under the Ukrainian Platform is normal, 40 km, that under the Outer Carpathians is deeper than normal, 50–65 km. Under the Great Hungarian Plain the Moho is quite shallow, 25–30 km. Other seismic data from the Great Hungarian Plain clearly show the same shallow Moho, and demonstrate that the IP-III profile is representative for the basin. Density estimates for the subcrustal lithosphere, obtained from inversion of gravity data along this profile [44] indicate that the upper mantle under the Plain is less dense than that under the Ukrainian Shield. This discrepancy may reflect temperature differences in the two regions.

Adam et al. [51] have made a comparison of seismic velocities and electrical conductivity measurements from beneath the Great Hungarian Plain. They have shown that both the P-wave velocity and the specific resistivity decreases markedly at a depth of 60 km. Extending the present high heat flow to depth, they interpret these decreases as the onset of partial melting at the base of the lithosphere. The depth of this feature lies between 50 and 70 km. This is very much shallower than elsewhere under the continents, except for areas like the Basin and Range (Fig. 9). Positive travel time residuals in this region (0.7–2.6 seconds [52]) also suggest that lithosphere under the Pannonian Basin is hot compared to normal continental lithosphere.

The geophysical measurements from the central

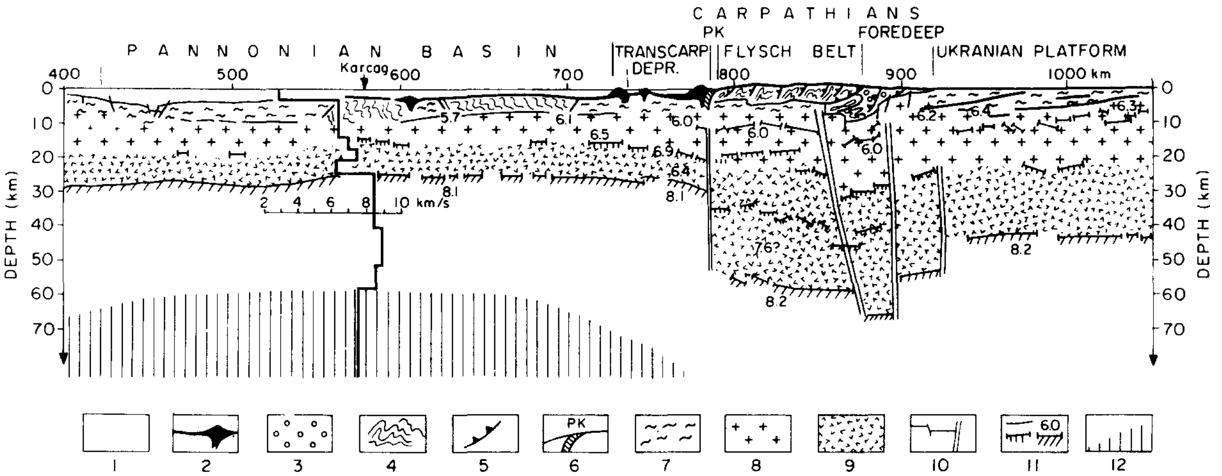


Fig. 9. Crustal structure of the Pannonian basin, Carpathians, and Ukrainian platform along the IP-III seismic line (see Fig. 3) (modified after Sollogub et al. [49]. The velocity vs. depth function was determined by Posgay [50] in the vicinity of Karcag. Key: 1 = Neogene-Quaternary sediments; 2 = Mio-Pliocene calc-alkaline volcanoes and volcanoclastics; 3 = molasse of the Carpathian foredeep; 4 = Upper Cretaceous Paleogene flysch; 5 = major thrust plane; 6 = Pieniny Klippen belt; 7 = Paleozoic and earlier basement, locally including Mesozoic rocks; 8 = upper crust; 9 = lower crust; 10 = local and major faults, respectively; 11 = seismic reflector, Conrad and Moho discontinuity, and seismic velocity, in km/s; 12 = subcrustal lithosphere and asthenosphere, respectively.

TABLE 5

Comparison of observations and predictions

	Observations	Predictions	
Peripheral basins			
Basement subsidence (water filled)		$\beta = 2$	
initial (km)	1.4	1.2	
thermal (m)	200	300	
Crustal thickness (km)	25	20 *	
Heat flow (μcal/cm^2 s)	1.4	1.7 **	
Pannonian Basin			
Basement subsidence (water filled)		$\beta = 4$,	β *** = 4
initial (km)	0	1.6,	0
thermal (m)	500–800	600,	500
Crustal thickness (km)	20–25	10 *,	20 *
Heat flow (μcal/cm^2 s)	2.5	2.8 **,	2.8 **

* Assuming 40 km thickness for pre-stretched crust.
** Assuming 0.6 μcal/cm^2 s heat flow from pre-stretched crust.
*** In this model the lithosphere was attenuated to 40 km and then stretched by a factor of 2.

basins are remarkably consistent. They imply a high heat flow, shallow lithosphere, and an attenuated crust.

6. Simple extensional explanations

Three idealized thermal models have been proposed to account for the geological development of continental shelves and intracontinental basins. These mechanisms are not exclusive and combinations of some or all could be operational in any basin. In this section we examine the compatibility of these proposed thermal models with the Neogene subsidence of the intra-Carpathian basins and the regional geological and geophysical information.

The development of the peripheral basins and that of the central basins are quite distinct. The peripheral basins are characterized by a subsidence history with two phases; the first with rapid sedimentation and a precipitous drop in basement depth, the second with slower, linear subsidence which continues to the Present (Fig. 6a). We propose that this first phase is an initial isostatic adjustment to deformation or alteration of the lithosphere, and that the second results from conductive decay of a thermal anomaly. In the central basins, there is only a linear subsidence phase which we propose is thermally controlled. For reasons which we shall discuss later, the initial, isostatic subsidence is not observed.

In considering mechanisms which are compatible with observations for the peripheral basins, we can rule out both dike intrusion and subcrustal attenuation as the dominant mechanism. The strikingly linear subsidence during the second (thermal) phase is not in good agreement with the $\sqrt{t}$ relationship predicted by the dike intrusion model. Furthermore, in the Vienna Basin there is little or no evidence of igneous activity. Although both the stretching model and subcrustal attenuation predict linear subsidence compatible with the observed subsidence, only one of these, stretching, can satisfactorily explain the initial isostatic subsidence. In order for the subcrustal attenuation model to produce initial isostatic subsidence, a considerable fraction of the lower crust would have to be melted, presumably resulting in huge amounts of calc-alkaline volcanism. This is not observed in the Vienna Basin. There is considerable volcanism in the other peripheral basins. Moreover, the heat flow average of the Transcarpathian Depression is nearly the double that of the Vienna Basin, 95 mW/m^2, as opposed to 50 mW/m^2 (Fig. 3). It shows that stretching alone cannot explain the formation of the depression. However, the observed features can be explained in terms of stretching by a factor of two combined with 30% dike intrusion which occurred some time later (cf. Appendix for computational procedure).

The present depth of Moho under the Vienna Basin and Transcarpathian Depression is 27–32 km and 26–28 km, respectively. By subtracting the 4- to 6-km sedimentary thickness 20–26 km is obtained, which is the end-product of the suggested stretching by a factor of two. This is reasonable since the crust under the Outer Carpathians is now about 50–55 km thick. Initial isostatic adjustment to stretching by a factor of two results in 2 km of unloaded subsidence. If 1 km of subsidence occurred above sea level, this would result in the observed 4 km of sediment fill. Thereafter, the water-filled subsidence would be about 300 m in 15 Ma (Fig. 2a, b and Appendix). This simple history is remarkably close to what is observed in the center of the Vienna Basin and Transcarpathian Depression. By reducing the amount of stretching it can be made to explain the West Danube Basin (Fig. 7).

The timining of these subsidence events is consistent with the geological history of the region. Extensional tectonics associated with compressional events started in the Eggenburgian, and continued until the earliest Badenian when the last major phase of compression terminated. It is during the latter part of this period that the rapid, isostatic subsidence occurred. Heat flow predicted for stretching by a factor of two gives 1.56 μcal/cm^2 s (65 mW/m^2) for the heat flow at present, i.e. after 15 Ma of conductive cooling. If the cooling effect of fast sedimentation is taken into consideration a slightly lower value is obtained (1.3–1.4 μcal/cm^2 s). Stretching by a factor of two and 30% dike intrusion at 10 Ma ago gives 2.4 μcal/cm^2 s for the heat flow at present, which is decreased by the sedimentation to 2.1–2.2 μcal/cm^2 s in good agreement with the observed values in the Transcarpathian Depression.

Although application of the stretching model gives good results in the peripheral basins, there are difficulties when this model is applied to the central basins. If we assume that the Pannonian was initially high, and subsided to sea level during the isostatic adjustment phase, then stretching by a factor of between 3 and 4 could account for the rapid basement subsidence since the Sarmatian. The present crustal thickness is 20–25 km. Stretching by the amount suggested implies an original crustal thickness of at least 60–80 km and an initial unloaded isostatic subsidence of between 3 and 5 km (Fig. 2a). Though not totally unreasonable it is excessive. Furthermore, three- to four-fold stretching requires about 250–300 km of extension, a figure which is considered extreme by all geologists working in the area. It is these geological constraints coupled with the absence of any observed initial subsidence that imply that simple uniform stretching is not the total explanation of the observed subsidence.

The crust under the Pannonian Basin appears to have been attenuated by about a factor of two. If we assume a stretching by a factor of two it gives a total extension of 150–200 km. This amount of extension cannot be excluded because extreme estimates of overthrusting of the Western Carpathian flysch complex onto the foredeep molasse are of this magnitude [54]. However, stretching by this amount cannot explain the observed subsidence and high heat flow and therefore we have to consider the thermal role of some other mechanisms.

We have calculated subsidence for a model where extension occurs by 50% stretching and 50% dike intrusion. This gives a good fit to the observed subsidence, but is not entirely consistent with other observations. First, an intrusional event of this magnitude would result in variable Moho depth which is not seen in these basins. Second, in the deeper portions of the basins, where maximum dike intrusion would be expected, a $\sqrt{t}$ contribution to the depth-age relation should be observable. On the contrary, the subsidence in these regions is still linear, as shown by the deepest profile in the basin (Fig. 6b).

As a consequence of these problems it is clear that there is no simple explanation of the peripheral basins that will also account for the central basins. In essence the subsidence and heat flow observations from the central basins require the rapid shallowing of the isotherms in the Middle Miocene without any initial subsidence or stretching by more than a factor of two.

7. Modified stretching model

In order to explain the observations it is necessary to introduce extra heat into the lithosphere during the initial phase of basin formation. The simplest way of doing this is to assume that the initial conditions were wrong and that the lithosphere was hotter than the linear equilibrium condition assumed before stretching. This is not unreasonable as the Pannonian basin was part of the Pannonian plate that lay over the downgoing European plate in the Early to Mid-Miocene [53]. Though it is unlikely that this extra heat was carried upwards by conduction because of the time constants involved there are other mechanisms by which the subcrustal lithosphere could be heated above equilibrium conditions. One possibility is that this entire heat results from tectonic interaction of smaller units within the Pannonian continental fragment which coincided chronologically with the termination of subduction along the arc. An alternative possibility is that the extra heat resulted directly from the termination of subduction. For example, assuming that the lithosphere consists of a brittle layer possibly as thick as the crust, overlying a more ductile layer then during subduction drag from the downgoing plate could remove some of the ductile layer from the bottom of the overlying plate. In this model the subcrustal attenuation could raise the depth of the lithosphere from 120 to roughly 40 km, without affecting the crust. This could have been followed by uniform stretching of the thinned lithosphere by a factor of two. If the initial crustal thickness were 40 km then the crust would be thinned to 20 km, and the asthenosphere raised to 20 km. These processes did not necessarily happen in the order given, and may have been synchronous. As a result of combining the two effects there is little or no initial isostatic subsidence and the temperature profile is close to that of simple stretching by a factor of 4. The consequent subsidence due to cooling is linear and matches that which is observed. In addition, the calculated heat flow is close to the average heat flow 2.5 $\mu cal/cm^2$ s (110 mW/m^2) in the basin. A simple schematic representation of how the initial temperature structure would be changed by this model is presented as Fig. 10.

There are spatial problems with this explanation. Extension of the lower lithosphere in the overriding plate cannot be greater or more rapid than the length or rate of subduction of the downgoing slab. It is unclear whether or not the amount of Miocene subduction is sufficient to produce the requisite thinning under the Great Hungarian Plain. On the other hand, we know little about the behavior of subducted plates

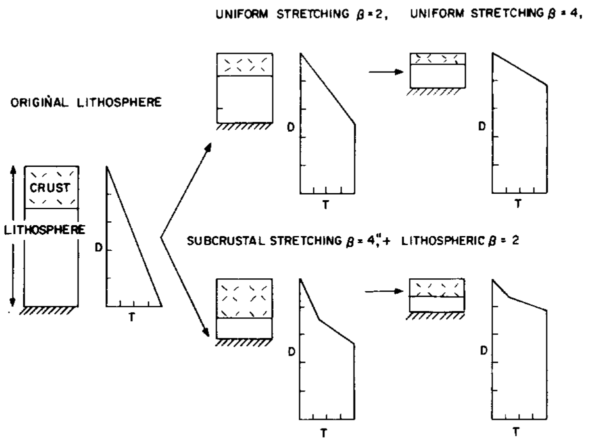


Fig. 10. A schematic diagram outlining how subcrustal stretching followed by whole lithosphere stretching can result in close to the same temperature profile as extensive uniform stretching. This enables high-temperature material to be brought close to the surface without excessive crustal thinning thus removing or reducing the initial subsidence. The same effect could also be created by lithospheric erosion.

at the time subduction ceases. Thus, it may be possible to create additional extension in the overriding plate by some sort of downward rotation or hinging of the subducted plate (Fig. 11b). In Roumania, where there is still some folding and crustal shortening in a small region in the outer portion of the arc, focal mechanisms for deep seismic activity suggests that the subducted plate is now descending vertically [55].

8. Erosional model

An alternative explanation for the extra heat necessary to produce the observed thermal subsidence not involving significant extension is subcrustal thinning and crustal erosion. If the downgoing plate drags with it aesthenospheric or even lithospheric material there would necessarily be flow in the aesthenosphere. Andrews and Sleep [56] have suggested that under these conditions erosion and melting of the lithosphere is rapid and results in the replacement of part of the lithosphere by asthenosphere. In this model the lithosphere under the central basin is fully attenuated by melting, erosion and downward

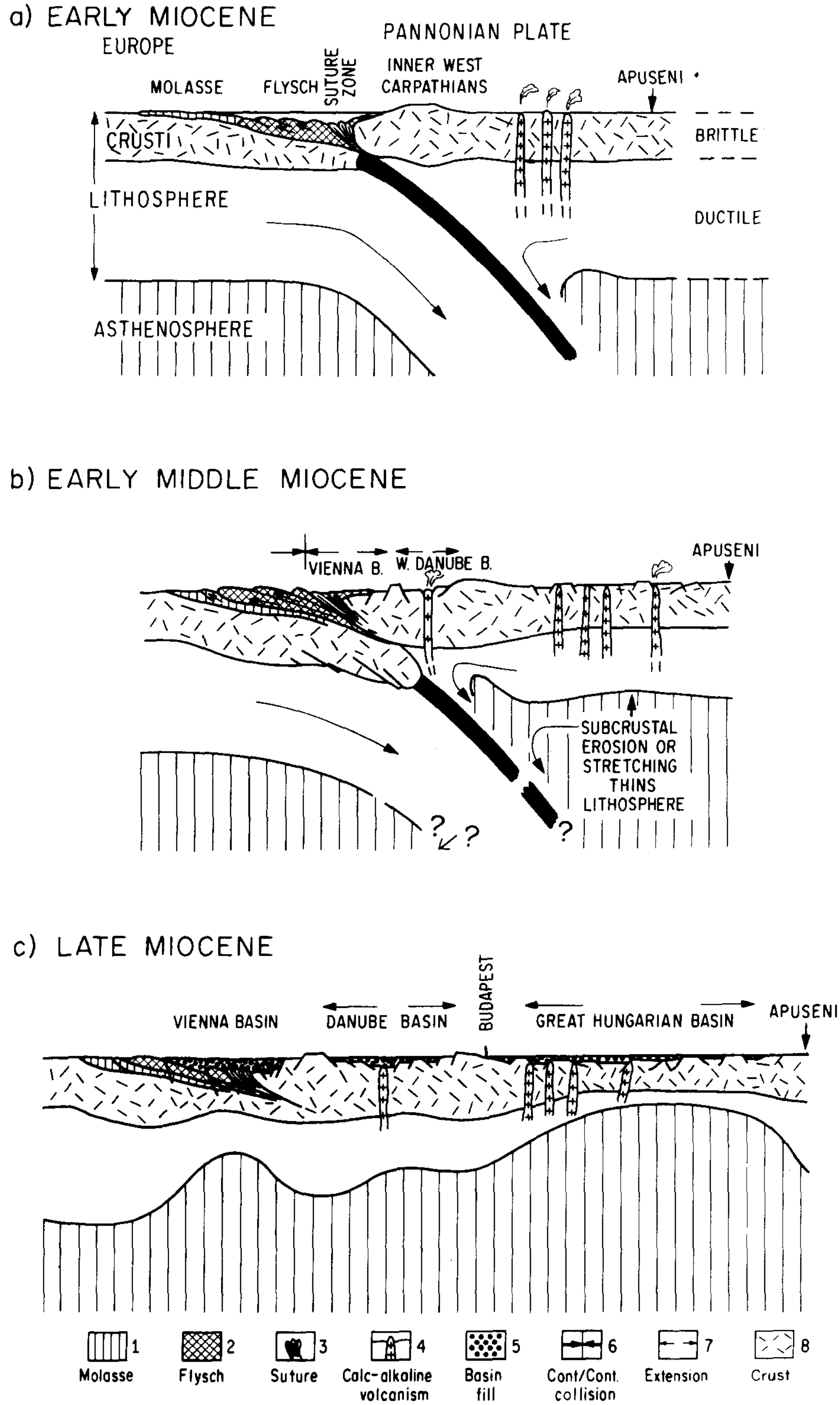


Fig. 11. (a) Early Miocene position of the European and Pannonian plates showing the development of the molasse and flysch deposits, rhyolitic volcanism and oceanic subduction. (b) Early Middle Miocene collision of the two continental portions of the plates. The Outer Carpathians are formed by the collision. Stresses, associated with the subduction of the ductile portion of the Pannonian plate and asthenospheric processes, thin the lithosphere. (c) At the end of Middle Miocene collision terminates and the Vienna and West Danube Basins are formed by the uniform stretching of a cold thick crust and lithosphere. In contrast the Little and the Great Hungarian Basin are formed by the cooling of a thin hot lithosphere. Key: 1 = molasse; 2 = flysch; 3 = suture zone; 4 = calc-alkaline volcanism; 5 = basin fill; 6 = continent-continent collision; 7 = extension; 8 = continental crust.

drag caused by the undergoing slab. Further about 10 km from the originally 35 km thick crust is also melted and replaced by hot asthenosphere. This process also does not necessarily happen in the order given, and may be synchronous. After these combined events there is a slight initial isostatic subsidence (0.1 km) and temperature profile is also close to that of a simple stretching by a factor of 4. The subsequent subsidence history and heat flow matches that which is observed. This model is similar to that presented by Haxby et al. [57] for the Michigan Basin.

The melting and erosion mechanism has the advantage of requiring only one process to account for the elevated temperatures. As a consequence for this case the amount of extension of the Pannonian Basin was most probably about 50 km and certainly not more than 100 km. The subcrustal melting model is compatible with the geographic distribution of regional volcanism. The largest volume of the calc-alkaline volcanics in the Carpatho-Pannonian system can be found in the Pannonian Basin, where they are mostly rhyolite ignimbrites. It is widely accepted that ignimbrites originate by widespread crustal fusion [37,58,59]. However, this model has a major space problem: if 10 km of the lower crust has been melted then where has all this igneous material gone? There is no evidence for 30–40% dyke intrusion in the crust or for a 10-km igneous layer at the base of the sedimentary layer.

Four of the authors (J.G.S., L.R., B.C.B. and S.S.) prefer the modified stretching and the other two the erosional explanation. The differences of interpretation are based on different approaches to the physics of the problem. The four who prefer to modify the stretching model view thermal anomalies as the passive reaction of the lithosphere to extensional or mechanical forces. The two who favor lithospheric erosion are satisfied with this concept and prefer it because it is the simplest mechanically.

Though we disagree on the interpretation of the high heat flow and subsidence of the central basins it is important to note that we agree on the observations and the interpretation of the peripheral basins. Furthermore, all of us are convinced that the uniform subsidence in the central basins results from the cooling of a very thin lithosphere created some time in the Badenian (13 Ma).

9. Tectonic model for Carpathian basin development

By correlating the timing of events in the Carpathian arc with the onset and rates of subsidence in the intra-arc basins, we can suggest a tentative outline for the Neogene development of the Carpatho/Pannonian region. We will use a profile from the Vienna Basin, extending through Budapest, and terminating against the Apuseni Mountains. During the latest Oligocene and earliest Miocene the European foreland was underthrusting the "Pannonian plate" along a zone where the crust of the Carpathian flysch basin had largely been consumed. There was molassic sediment deposited on the European foreland while more towards the inner West Carpathians (leading edge of the Pannonian plate) the piling up of the flysch nappes commenced (Fig. 11a). The Pannonian plate is under both compression and tension. Compression in the crust is due to the beginning of the collision with the European plate and tension at depth resulting from the subduction of part of the ductile lower lithosphere of the Pannonian plate (Fig. 11a). Sometime in the early Middle Miocene the paroxysm of the continental collision occurred. The flysch sequence is compressed against and overrides the European plate. There is some extension at depth under the Pannonian Plate and the lithosphere starts to thin by either subcrustal erosion and melting or by simple extension (Fig. 11b). The considerable Middle Miocene andesitic and rhyolitic volcanism on the Pannonian plate is related to this process. This is the last stage of compression and thrusting in the Carpathians, the Vienna Basin, West Danube Basin, and the other peripheral basins are formed by simple stretching within the nappe type structures. At this time the lithosphere under the Pannonian Basin is further thinned by continued subcrustal erosion or by stretching. Such stretching may be associated with the downward rotation of the subducted portion of the European plate (Fig. 11b). By the end of Middle Miocene the active extension in peripheral basins has terminated leaving the lithosphere boundary close to the base of the crust, at a depth of some 50–60 km under the Vienna and West Danube Basins (Fig. 11c). The uniform stretching by a factor of two and the subsequent cooling of the thin lithosphere explains the rapid early subsidence and slow post-Sarmatian subsidence of the peripheral basins. In the central basins, either twofold crustal stretching accompanied by extra thinning of the lithosphere or subcrustal erosion and lower crustal melting accounts for the absence of initial subsidence in these basins. The more rapid post-Sarmatian subsidence is a result of the cooling of an extremely shallow (20–30 km) lithosphere. Between the Late Miocene and the present there is no major tectonic activity and, except for the slight Quaternary uplift in the Transdanubian Mid-Mountains and Transylvanian Basins, the basins have subsided quiescently to their present depths.

The model we have presented above is necessarily much simpler than the actual development of the Carpathians and Pannonian. However, if it is in essence correct, it may have major implications for creation of black-arc basins and for extension behind continental collision zones and Andean type margins. If subcrustal attenuation of the overriding plate occurs continuously during subduction of oceanic material, and if the rate of subduction is so great that cracks form in the rigid part of the overriding plate, then passive penetration of basaltic material may result in formation of back-arc basins.

Careful reconstruction of thermal histories from subsidence data has great importance for the maturation of hydrocarbons in these basins. Although the total thickness of sediment in the peripheral and central intra-Carpathian basins is roughly equivalent, their temperature histories are radically different. For example, the Vienna Basin started cold and has never been very hot. In contrast, the Pannonian Basin started hot and has cooled down very little. This may explain why source material for hydrocarbons in the Vienna Basin is Mesozoic while that in the Pannonian Basin is probably Neogene. Our analysis explains how adjacent basins which are tectonically related can have very different thermal histories. Such ideas could be very important in the search for hydrocarbons and geothermal reservoirs in similar tectonic areas.

10. Conclusions

(1) We have shown that the intra-Carpathian basins are thermal in origin and are separated into two distinct categories. The first type of basin lies in the peripheral regions of the intra-Carpathian lowlands and has very fast initial subsidence followed by a period of slower, linear subsidence. These basins were formed by about two-fold stretching of the lithosphere with some contribution from dike intrusion in the Trans-Carpathian Depression. The rapid initial subsidence is an isostatic response to stretching which occurred during the Early/Middle Miocene (Eggenburgian to Sarmatian). The slower linear subsidence from the Sarmatian onward is thermally controlled.

(2) The second type of basin lies in the central intra-Carpathian region, has high heat flow and a reasonably fast linear subsidence since the late Middle Miocene (Sarmatian). This type of basin may be formed by either (a) two-fold stretching accompanied by subcrustal attenuation of the lithosphere, or (b) attenuation of the whole subcrustal lithosphere and part of the crust by subcrustal melting and erosion. These may be the result of lithospheric drag and/or secondary convection induced by subduction along the Carpathian arc.

(3) Careful analysis of sedimentation and subsidence rates enables us to reconstruct the thermal history of these complex intermountain basins. When detailed records of subsidence history are not available, models for the genesis of intracontinental sedimentary basins are unconstrained and open to debate. The subsidence history gives quantitative, rather than qualitative, constraints on basin evolution and should prove to be a valuable tool in sorting out how and why intracontinental basins and continental margins are formed.

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Appendix

In general we may consider the lithosphere as a slab whose upper surface, $z = a$, is maintained at temperature $T = 0$ and lower surface, $z = 0$, at $T = T_1$. If the slab is thermally perturbed at time $t = 0$, and subsequently cools through vertical conduction, the temperature distribution may be given by:

$$T = T_1\left(1 - \frac{z}{a}\right) + T_1 \sum_{n=1}^{\infty} \left(\frac{2}{n\pi}(-1)^{n+1}\right) x_n \times \sin\frac{n\pi z}{a} \exp\left(\frac{-n^2\pi^2 kt}{a^2}\right) \tag{A-1}$$

where $[(2/n\pi)(-1)^{n+1}]x_n$ is the nth Fourier coefficient of the initial temperature distribution. Likewise, surface heat flux, Q, and elevation above final equilibrium elevation, U, may be written:

$$Q = \frac{KT_1}{a}\left[1 + \sum_{n=1}^{\infty} 2x_n \exp\left(\frac{-n^2\pi^2 kt}{a^2}\right)\right] \tag{A-2}$$

and:

$$U = \alpha a T_1 \frac{4}{\pi^2} \sum_{m=0}^{\infty} \frac{x(2m+1)}{(2m+1)^2} \exp\left[1 - (2m+1)^2 \frac{\pi^2 kt}{a^2}\right] \tag{A-3}$$

We can calculate the x_n's which correspond to the three thermal processes described in the text; uniform stretching, dike intrusion, and subcrustal attenuation. If the lithosphere stretches by a factor δ from the surface to a depth y, and stretches by a factor β from y to the base of the lithosphere and is finally intruded by dikes, comprising a fraction γ of the lithosphere, the initial temperature distribution is:

$$T = T_1\,\delta\left(1 - \frac{z}{a}\right) + \gamma T_1 \qquad \text{for } \left(a - \frac{y}{\delta}\right) \leqslant z \leqslant a$$

$$T_1\,\beta\left(1 - \frac{z}{a}\right) + \frac{y}{a}\left(1 - \frac{\beta}{\sigma}\right)(1-\gamma) + \gamma T_1$$

$$\text{for } a - \frac{y}{\sigma} - \frac{(a-y)}{\beta} \leqslant z \leqslant \left(a - \frac{y}{\delta}\right)$$

$$T = T_1 \qquad \text{for } 0 \leqslant z \leqslant a - \frac{y}{\delta} - \frac{(a-y)}{\beta}$$

and x_n is:

$$x_n = \gamma + \left\{(1-y)\left[(\delta - \beta)\sin n\pi H + \beta \sin n\pi G\right] \times \frac{(-1)^{n+1}}{n\pi}\right\} \tag{A-4}$$

where:

$$H = 1 - \frac{a}{y}$$

$$G = 1 - \frac{y}{a\delta} - \left[\left(1 - \frac{y}{a}\right)\Big/\beta\right]$$

The total amount of heat (dimensionless) added to the lithosphere is:

$$\Delta H = \left[\left(1 - \frac{1}{\beta}\right) + \left(\frac{y^2}{a^2} - \frac{2y}{a}\right)\left(\frac{1}{\delta} - \frac{1}{\beta}\right)\right](1-\gamma) + \gamma \tag{A-5}$$

where $\Delta H = 1$ is "total oceanization", or non-equilibrium heat input at a mid-ocean ridge. $\Delta H = 0$ represents a stable thermal equilibrium in the lithosphere (cold continent).

In addition to the thermal subsidence caused by conductive cooling of the lithosphere there is an immediate isostatic elevation change at $t = 0$ in response to deformation of the lithosphere. There are two contributing factors: (1) density changes due to elevated temperatures in the lithosphere and thermal expansion, and (2) density changes due to crustal thinning and replacement of light crustal rocks by denser ultrabasic material. Elevation changes associated with thermal expansion are proportional to the total amount of heat added to the lithosphere (equation A-5) and may be written:

$$E = \frac{T_1 \alpha a\,\Delta H}{2(1 - \alpha T_1)} \tag{A-6}$$

Elevation changes due to crustal thinning or replacement can be written:

$$-\left(\frac{\rho_m - \rho_c}{\rho_m(1 - \alpha T_1)}\right) t_c \left(1 - \frac{1}{\delta} + \frac{\gamma}{\delta}\right)\left(1 - \frac{T_1 \alpha t_c}{2a}\right) \qquad (y \geqslant t_c)$$

$$\frac{-(\rho_m - \rho_c)}{\rho_m(1 - \alpha T_1)}\left\{y\left(1 - \frac{1}{\delta} + \frac{\gamma}{\delta}\right)\left(1 - \frac{\alpha T_1 y}{2a}\right) + (t_c - y)\left(1 - \frac{1}{\beta} + \frac{\gamma}{\beta}\right)\left[1 - \frac{\alpha T_1}{2a}(y + t_c)\right]\right\}$$

$$(y \leqslant t_c) \tag{A-7}$$

where t_c is the original crustal thickness. Positive values indicate uplift, negative values indicate subsidence. The parameters used in the calculations are

ρ_m, the density of the mantle = 3.3 g/cm^3; ρ_c, the mean density of the crust = 2.9 g/cm^3; α, the thermal expansion coefficient = 3.2×10^{-5} °C^{-1}; a, the thickness of the lithosphere = 125 km; T_1, the temperature at the bottom of the plate = 1350°C; KT_1/a, the equilibrium heat flux = 0.8×10^{-6} cal/cm^2 s; and k, the thermal diffusivity = 0.0075 cm^2/s; and all are taken from Parsons and Sclater [1].

References

1 B. Parsons and J.G. Sclater, An analysis of the variation of ocean floor bathymetry and heat flow with age, J. Geophys. Res. 82 (1977) 803–827.

2 N.H. Sleep, Thermal effects of the formation of Atlantic continental margins by continental breakup, Geophys. J. R. Astron. Soc. 24 (1971) 325–350.

3 M.S. Steckler and A.B. Watts, Subsidence of the Atlantic-type continental margin off New York, Earth Planet. Sci. Lett. 41 (1978) 1–13.

4 C.E. Keen, Thermal history and subsidence of rifted continental margins – evidence from wells on the Nova Scotian and Labrador shelves, Can. J. Earth Sci. 16 (1979).

5 A.B. Watts and W.B.F. Ryan, Flexure of the lithosphere and continental margin basins, Tectonophysics 36 (1976) 25–44.

6 B.J. Collette, On the subsidence of the North Sea Area, in: Geology of Shelf Seas, D.T. Donovan, ed. (Oliver and Boyd, Edinburgh, 1968) 15–30.

7 R.J. O'Connell and G.J. Wasserburg, Dynamics of the motion of a phase boundary to change in pressure, Rev. Geophys. 5 (1967) 329.

8 M.H.P. Bott, Formation of sedimentary basins of graben type by extension of the continental crust, Tectonophysics 36 (1976) 77–86.

9 D. McKenzie, Active tectonics of the Alpide-Himalayan belt: the Aegean Sea and surrounding regions, Geophys. J. R. Astron. Soc. 55 (1978) 217–254.

10 D. McKenzie, Some remarks on the development of sedimentary basins, Earth Planet. Sci. Lett. 40 (1978) 25–32.

11 L. Royden, J.G. Sclater and R.P. Von Herzen, Continental margin subsidence and heat flow: important parameters in formation of petroleum hydrocarbons, Am. Assoc. Pet. Geol. Bull. 64 (1980) 173–187.

12 J.G. Sclater and P. Christie, Continental stretching – an explanation of the post-mid-Cretaceous subsidence of the central North Sea basin, J. Geophys. Res. (in press).

13 K. Burke, Hot spots and aulacogens of the European margin: Leicester/Shropshire; Oslo/Fen; Geol. Soc. Am. Abstr. 7 (1975) 34–35.

14 C.H. Scholz, U. Barazangi and M.L. Sbar, Late Cenozoic evolution of the Great Basin, Western United States, as an ensialic inter-arc basin, Geol. Soc. Am. Bull. 82 (1971) 2979–2990.

15 A.H. Lachenbruch, Heat flow in the Basin and Range Province and thermal effects of tectonic extension, Pageoph 117 (1978) 34–50.

16 L. Royden and C. Keen, Rifting process and thermal evolution of the continental margin of eastern Canada determined from subsidence curves, Earth Planet. Sci. Lett. 51 (1980) in press.

17 G.T. Jarvis and D. McKenzie, Sedimentary basin formation with finite extension rates, Earth Planet Sci. Lett. 48 (1980) 42–52.

18 O. Fusan, F. Ibrmajer, F. Plancar, F. Slavik and M. Smisek, Geological structure of the basement of the covered southern part of the inner west Carpathians, Zb. Geol. Vied, Zapadne Karpaty 15 (1971) 115–173.

19 Gy. Wein, Eine strukturgeologische Skizze des vorneogenen Untergrund des klienen Tiefebene, Proc. 10th Congr. Carpatho-Balkan, Geol. Assoc., Geol. Ust., Bratislava 3 (1975) 36–345.

20 F. Steininger, A. Papp, I. Cicha and J. Senes, Marine neogene in Austria and Czechoslovakia, in: Excursion "A", Reg. Comm. Med. Neog. Stratigr., 6th Congr. (Veda Publ. House, Slov. Akad. Sci., Bratislava, 1975) 1–96.

21 D. Ciupagea, M. Pauca and Tr. Ichim, Geologica Depresiunii Transilvaniei, Bucuresti, Ed. Acad. (1970) 1–256.

22 Gy. Wein, Tectonic review of the Neogene-covered areas of Hungary, Acta Geol. Acad. Sci. Hung. 13 (1969) 399–436.

23 K. Szepesházy, Structural and stratigraphical connection between the basement of the Great Hungarian Plain east of the river Tisza and the Apusei Mts. in western Transylvania, Ált. Földt. Szemle, Budapest, 12 (1979) 178–198.

24 B. Géczy, Plate tectonics and paleogeography of the East-Mediterranean Mesozoic, Acta Geol. Acad. Sci. Hung. 17 (1973) 421–428.

25 J. Channel and F. Horvàth, The African/Adriatic promontory as a paleogeographical premise for Alpine orogeny and plate movements in the Carpatho-Balkan region, Tectonophysics 35 (1976) 71–101.

26 F. Horvàth, A. Vörös and K.M. Onuoha, Plate tectonics of the Western Carpatho-Pannonian region, a working hypothesis, Acta Geol. Acad. Sci. Hung. 21 (1977) 207–221.

27 R. Marschalko and T. Korab, The position of the east Slovakian flysch in the Carpathians, Mineral. Slovaca 7 (1975) 53–80.

28 D.P. Radulescu and M. Sandulescu, The plate tectonics concept and the geological structure of the Carpathians, Tectonophysics 16 (1973) 155–161.

29 M. Mahel, Development model of the West Carpathians, Geol. Zb., Geol. Carpath. 28 (1977) 203–218.

30 M. Mahel, Some particularities of development of the European Alpides and West Carpathians, mainly from the viewpoint of new global tectonics, Geol. Zb., Geol. Carpath. 21 (1978) 1–18.

31 Z. Roth and B. Lesko, The Outer Carpathian flysch belt in Czechoslovakia, in: Tectonics of the Carpathian-Balkan Regions, M. Mahel, ed. (Geol. Inst. D. Stur, Bratislava, 1974) 158–163.

32 F. Rögl, F.F. Steininger and C. Müller, Middle Miocene

salinity crisis and paleogeography of the Paratethys (Middle and Eastern Europe), in: K. Hsü, L. Montadert et al., Initial Reports of the Deep Sea Drilling Project, 42, part 1 (1978) 985–990.
33 F. Senes and I. Cicha, Neogene of the West Carpathian Mts. in: Guide to Excursion F, 10th Congr. Carpatho-Balkan, Geol. Assoc. (Geol. Inst. D. Stur, Bratislava, 1973) 1–46.
34 R. Rudinec, Paleogeographical, lithofacial and tectogenetic development of the Neogene in eastern Slovakia and its relation to volcanism and deep tectonics, Geol. Zb., Geol. Carpath. 29 (1978) 225–240.
35 G. Pantó, Cenozoic volcanism in Hungary, in: Guide to Excursion 40C, 23rd Sess. Int. Geol. Congr., Prague (1968) 1–106.
36 G. Hámor and A. Jámbor, A magyaror szági közepsömiocén (The middle Miocene in Hungary), Föld. Közl. 101 (1971) 91–102.
37 E. Szádeczky-Kardoss, Gy. Pantó, T. Poka, G. Pantó, V. Szèkyne-Fux, F. Kiss and I. Kubovics, Die neovulkanite Ungarns, Acta Geol. Acad. Sci. Hung. 11 (1967) 161–180.
38 K. Balogh and L. Körössy, Hungarian mid-Mountains and adjacent areas, in: Tectonics of the Carpathian-Balkan Regions, M. Mahel, ed. (Geol. Inst. D. Stur, Bratislava, 1974) 391–403.
39 F. Kokay, Geomechanical investigation of the south-eastern margin of the Bakony Mts. and the age of the Liter fault line, Acta Geol. Acad. Sci. Hung. 20 (1976) 245–257.
40 D. Radulescu and M. Borcos, Spätsubsequenter alpiner Magmatismus in Rumänien, Acta Geol. Acad. Sci. Hung. 11 (1967) 139–152.
41 A. Ronai, Size of Quaternary movements in Hungary's area, Acta Geol. Acad. Sci. Hung. 18 (1974) 39–44.
42 R. Grill, J. Kapounek, H. Kupper, A. Papp, B. Plöchinger, S. Prey and A. Tollmann, Guide to Excursion 33C Austria, Neogene Basins and Sedimentary Units of the Eastern Alps near Vienna, International Geological Congress XXIII Session (Geological Survey of Austria, Vienna, 1968) 1–75.
43 M. Mahel, Tectonics of the Carpathian-Balkan Regions (Geol. Inst. D. Stur, Bratislava, 1974),
44 F. Horváth and L. Stegena, The Pannonian Basin: a Mediterranean interarc basin, in: International Symposium on the Structural History of the Mediterranean Basins, Split(Yugoslavia), B. Biji-Duval and L. Montadert, eds. (Editions Technip, Paris, 1977) 333–340.
45 L. Körössy, Tectonics of the Basin Areas of Hungary, Acta Geol. Acad. Sci. Hung. 8 (1964) 377–394.
46 L. Stegena, B. Geczy and F. Horvâth, Late Cenozoic evolution of the Pannonian Basin, Tectonophysics 26 (1975) 71–90.
47 L. Stegena, The structure of the earth's crust in Hungary, Acta Geol. Acad. Sci. Hung. 8 (1964) 413–431.
48 F. Horvàth, L. Bodri and P. Ottlik, Geothermics of Hungary and the tectonophysics of the Pannonian basin "red spot", in: Terrestrial Heat Flow in Europe, V. Cermak and L. Rybach, eds. (Springer, Berlin/Heidelberg/New York, N.Y., 1979) 206–217.
49 V.B. Sollogub, D. Prosen and co-workers, Crustal structure of Central and Southeastern Europe by data of explosion seismology, in: The Structure of the Earth's Crust Based on Seismic Data, S. Mueller, Tectonophysics 20 (1973) 1–33.
50 K. Posgay, Mit Reflexionsmessungen bestimmte Horizonte und Geschwindigkeitsverteilung in der Erdkruste und in Erdmantel. Geophys. Trans., Budapest 23 (1975) 13–17.
51 A. Adam, F. Horvàth and L. Stegena, Geodynamics of the Pannonian basin: geothermal and electromagnetic aspects, Acta Geol. Acad. Sci. Hung. 21 (1977) 251–260.
52 C. Morelli, S. Belleno and G. De Visintini, Non-directional travel-time anomalies for European stations, Bull. Geofis. Teor. Appl. 10 (1968) 164–180.
53 B.C. Burchfield, Eastern European Alpine systems and the Carpathian Orocline as an example of collision tectonics, Tectonophysics 63 (1980) 31–62.
54 C. Tomek, F. Svancara and L. Budik, The depth and origin of the West Carpathian gravity low, Earth Planet. Sci. Lett. 44 (1979) 39–42.
55 C. Roman, Seismicity in Romania – evidence for the sinking lithosphere, Nature 228 (1970) 1176–1178.
56 D.F. Andrews and N.H. Sleep, Numerical modeling of tectonic flow behind island arc, Geophys. J. R. Astron. Soc. 38 (1974) 237–251.
57 W.F. Haxby, D.L. Turcotte and J.M. Bird, Thermal and mechanical evolution of the Michigan Basin, Tectonophysics 36 (1976) 57–75.
58 H. Pichler and W. Ziel, The Cenozoic rhyolite-andesite associations of the Chilean Andes, Bull. Volcanol. 35 (1972) 424–452.
59 J. Klerkx, S. Deutsch, H. Pichler and W. Ziel, Strontium isotope composition and trace element data bearing on the origin of Cenozoic volcanic rocks of the central and southern Andes, J. Volcanol. Geotherm. Res. 2 (1977) 48–71.