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MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

The Neogene Intra-Carpathian Basins [and Discussion]

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The Neogene intra-Carpathian basins

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The Carpathian intra-arc region is composed of two distinct types of sub-basins, the peripheral basins lying adjacent to the arc, and the central basins that lie in the central intra-arc region. Both sets of basins have approximately 2–6 km of Miocene and younger sediment, 20–25 km pre-Miocene crust, and elevated heat flow. Examination of their subsidence histories suggests that the basins are thermal in origin and formed by twofold stretching. In the central basins this twofold stretching must be combined with an additional mechanism for introducing heat into the lithosphere that does not significantly alter the structure or thickness of the stretched crust.

Geology

The Carpathian orogenic belt is the eastward extension of the western and eastern Alps, and is the result of lithospheric convergence and continental collision during Cretaceous to earliest Miocene time. Collision occurred during southward directed subduction of an oceanic terrane when a part of the Apulian continental fragment overrode the European plate (Horváth *et al.* 1977; Mahel 1977). The eastward convex loop of the Carpathians is made up of a stack of externally and radially directed thrust nappes that consist of older crystalline rocks, with and without their Mesozoic cover, flysch and flysch-like rocks, and, in the most external units, shallow water molassic sediments. In places, the flysch units contain exotic blocks of mafic rocks in what appears to be a fore-arc accretionary complex, and parts of dismembered ophiolites (?) (pillow lavas, diabase and gabbro) occur in a few regions. The total shortening across the arc is probably several hundred kilometres (Burchfiel 1980). In the Lower–Middle Miocene, during and after the last stages of thrusting in the Outer Carpathians, a 'back-arc' basin began to form within the Carpathian loop. Basin formation was accompanied by large-scale extension of the intra-arc region, interrupted by brief periods of compression (Balogh & Körössy 1974). At the same time, volcanic activity began in this region, producing andesites, rhyolites and dacites (Pantó 1968). Volcanic activity continued through the Quaternary when basalts were erupted in the final stages of volcanism. Subsidence of the intra-arc region has continued through to the present, with a total sediment accumulation of 2–6 km.

The basins surrounded by the Carpathian arc are not continuous. The Neogene–Quaternary subsidence only affected certain areas, leaving some blocks emergent. These blocks divide the back-arc region into several sub-basins (figure 1). These sub-basins are of two types: (1) those lying in the peripheral regions of the intra-Carpathian lowlands (Vienna, west Danube, Transcarpathian and Transylvania) and (2) those lying in the central intra-Carpathian region (east Danube, Little Hungarian and Great Hungarian (Pannonian)). Although both groups of basins have thin crust, elevated heat flow and similar ages for the onset of subsidence, the detailed subsidence history and present heat flow observed in the two sets of basins are quite different. Because the onset of basin formation coincided with a period of large-scale extension within the basement, and because of the thinned crust and elevated temperatures observed in this region, we believe that these basins are extensional and thermal in origin and are the direct result of early Miocene extension.

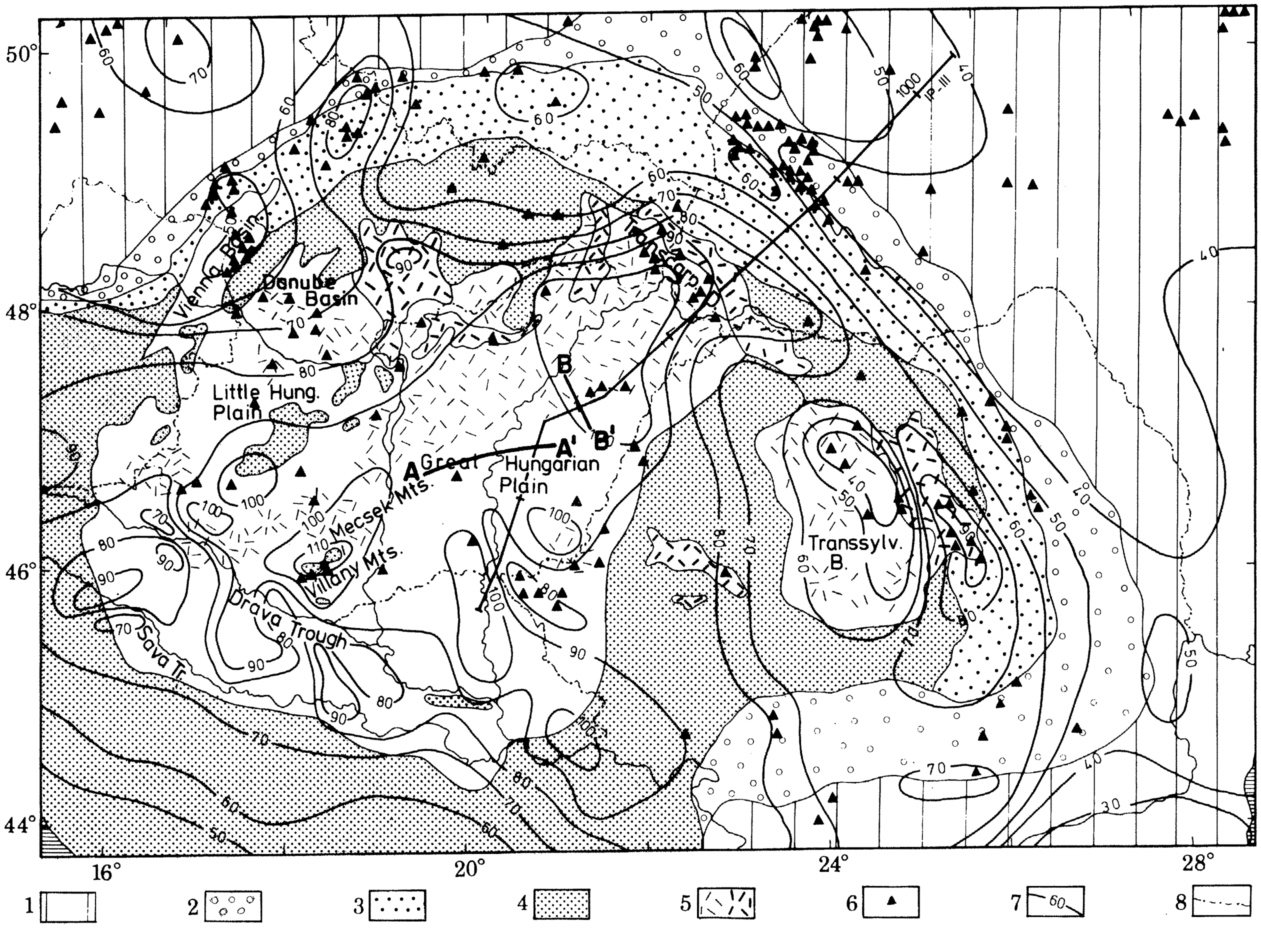

FIGURE 1. Tectonic sketch and heat flow map of the Pannonian basin and the surrounding Carpathian arc. The position of the heat flow stations is shown. Key: 1, European foreland; 2, Oligocene to Pliocene Molasse; 3, Upper Cretaceous and Palaeogene flysch nappes; 4, late Mesozoic (mainly Upper Cretaceous) nappe system of the eastern Alps, Inner Carpathians, Apuseni Mountains and Dinarides; 5, Mio–Pliocene calc-alkaline volcanics (covered and outcropping respectively); 6, Neogene–Quaternary basins and heat flow stations; 7, heat flow isolines in milliwatts per square metre; 8, political boundaries. The heavy continuous lines are the geological cross sections shown in figure 4 and a seismic refraction line.

Thermal models

Recently, McKenzie (1978*a*, *b*) has suggested that simple extension may be the mechanism for basin formation in the Aegean and other subsided sedimentary basins. Royden *et al.* (1980) and Royden & Keen (1980) have suggested similar extensional explanations of these areas, based on an analysis of Atlantic-type continental margins. Since these processes have been discussed in detail in these and other publications, we shall not repeat those discussions here. Figure 2 shows several of these extensional mechanisms: dyke intrusion and uniform and non-uniform stretching. In the Carpathian intra-arc region, surface and subsurface volcanism does not appear sufficiently extensive to be a fundamental factor in basin formation. Therefore, the

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subsidence histories, heat flow, crustal thickness and other geological constraints are examined with respect to stretching and thinning during basin formation. The theoretical subsidence predicted by extensional models can be divided into two distinct parts. During and immediately after stretching there is a rapid change in elevation. This occurs in isostatic response to net density changes resulting from both crustal thinning and from heating and thermal expansion of the lithosphere. This process is called the initial subsidence. The second stage of subsidence is a relatively long-term process caused by cooling and thermal contraction of the lithosphere after the extensional phase. This is called the thermal subsidence. Both the initial and thermal subsidence are generally magnified by the effects of sediment loading, which must be removed by careful decompaction of sediments (Sclater & Christie 1980) and by use of point-wise or flexural loading models (Watts & Ryan 1976; Steckler & Watts 1978). Mathematical calculations for subsidence and heat flow resulting from these extensional mechanisms are given in Sclater *et al.* (1980) and Royden & Keen (1980).

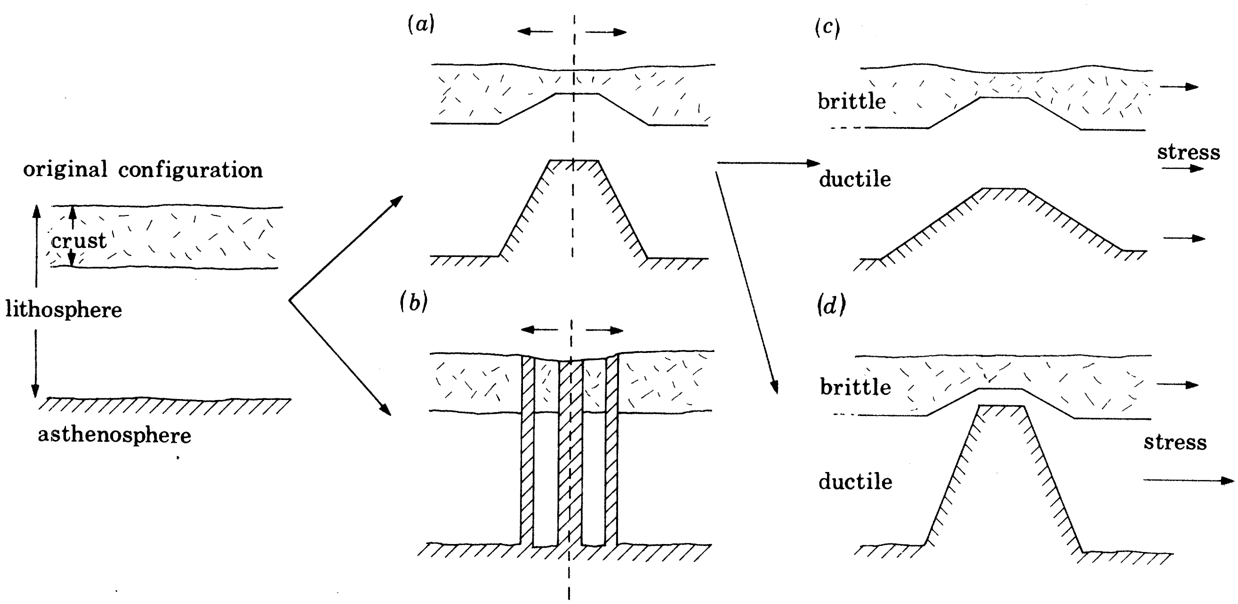

FIGURE 2. Various extensional models for the creation of the initial thermal anomaly that can account for the subsidence and heat flow of intracontinental basins: (*a*) and (*b*) are uniform stretching and dyke intrusion; (*c*) and (*d*) are modifications of the stretching model for a non-uniform rheology. In model (*c*), the ductile lithosphere thins over a wider zone than the rigid crust, although the net extension is the same. In model (*d*), net extension is greater in the lithosphere than in the crust.

A comparison of observed and theoretical subsidence profiles

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 (figure 3). Since sediment in these basins is primarily shallow water or lacustrine, the rate of sediment accumulation is a good indicator of vertical movement of the basement (Steininger *et al.* 1975; Rudinec 1978). We propose that this first phase is an initial isostatic adjustment to stretching of the lithosphere and that the second results from conductive decay of a thermal anomaly. Figure 3 shows subsidence histories for three of the peripheral basins, plotted against a theoretical subsidence curve for twofold stretching for lithosphere originally at equilibrium temperatures and with an initial crustal thickness of 40 km. We have assumed that stretching took place between 16.5 and 13 Ma. The present depth to Moho under the Vienna basin and Transcarpathian Depression is 27–32 km and 26–28 km respectively (Sollogub *et al.* 1973). By subtracting the 4–6 km sedimentary thickness, 20–26 km is obtained, which is the end-product of the suggested twofold stretching (table 1).

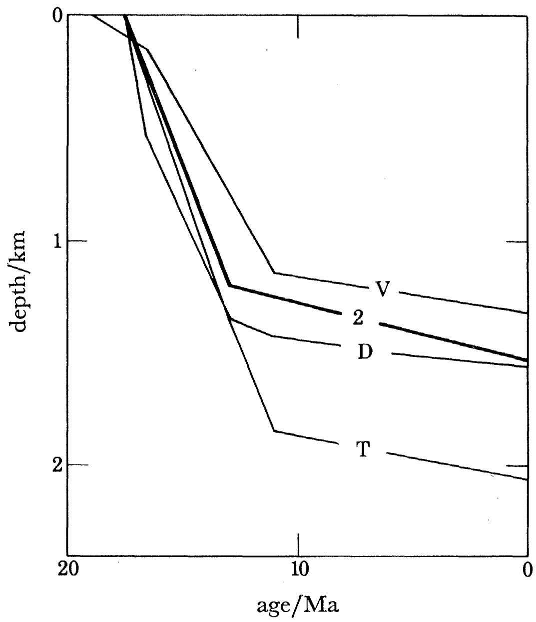

FIGURE 3. Water-loaded subsidence of basement as inferred from wells in the centre of the Vienna basin (V) (Grill *et al.* 1968), the Transcarpathian Depression (T) (Rudinec 1978) and the west Danube Basin (D) (Mahel 1974). These are plotted with theoretical subsidence for a twofold stretching of a 40 km thick continental crust, as discussed in the text. The stretching was assumed to take place between 16.5 and 13 Ma ago.

TABLE 1

	peripheral basins		Pannonian basin		
	observations	predictions $B = 2$	observations	predictions $B = 4$	$B = 4$†
initial subsidence/km	1.4	1.2	none	1.6	none
thermal subsidence/m	200	300	500–800	600	500
crustal thickness/km	25	20	20–25	10	20
heat flow/(mW/m^2)	55	70	100	110	110

† Fourfold subcrustal stretching, followed by twofold whole lithospheric extension.

Initial isostatic adjustment to twofold stretching 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 initial sediment fill, or 1.3 km of water-loaded subsidence. Thereafter, the thermal subsidence (water-loaded) would be about 300 m in 13 Ma (figure 3; table 1). This simple history is in good agreement with what is observed in the centre of the Vienna and Transcarpathian basins. By reducing the amount of stretching it can be made to explain to west Danube Basin. In addition, surface heat-flow measurements in these basins are close to those predicted by twofold stretching (70 mW/m^2 after 15 Ma of conductive cooling), except in the Transcarpathian basin where

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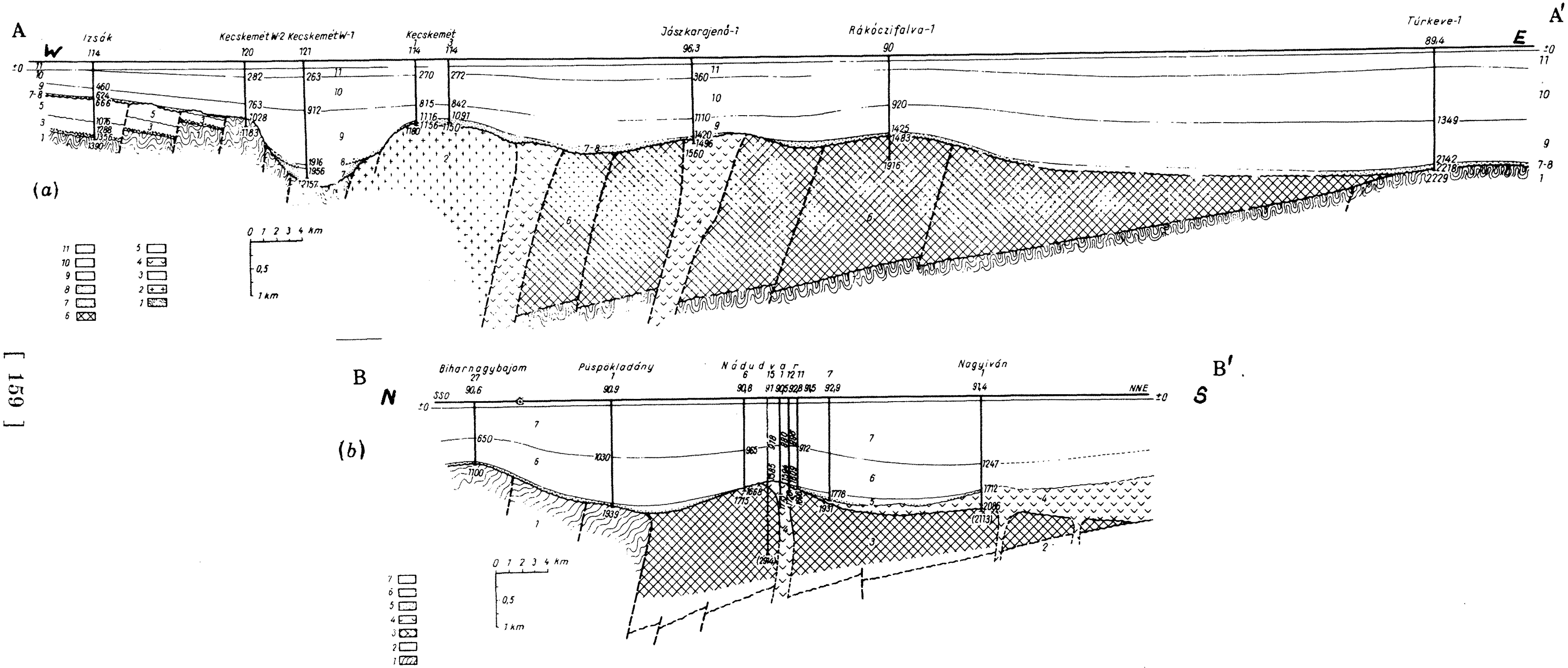

FIGURE 4. (*a*) Geological profile along AA′ in figure 1. Key: 1, crystalline schists; 2, granite; 3, Lower Cretaceous; 4, basic eruptives (diabase); 5, Upper Cretaceous; 6, Cretaceous–Palaeogene flysch; 7, Tortonian; 8, Sarmatian; 9, Lower Pannonian; 10, Upper Pannonian; 11, Pleistocene–Holocene (Körössy 1964).

(*b*) Geological profile along BB′ in figure 1. Key: 1, crystalline schists; 2, Palaeozoic–Mesozoic formations; 3, Upper Cretaceous–Palaeogene flysch; 4, Helvetian–Tortonian; 5, Tortonian; 6, Lower Pannonian; 7, Upper Pannonian–Pleistocene (Körössy 1964).

the heat flow is higher (Horváth *et al.* 1979). This is probably the result of widespread post-extensional volcanism that occurred in the Transcarpathian region.

Although application of this simple stretching model gives good results in the peripheral basins, there are some problems when this model is applied to the central basins. In contrast to the peripheral basins, the central basins show a single phase of subsidence that began about 15 Ma ago (figure 4). The subsidence is quite linear and has continued through to the present.

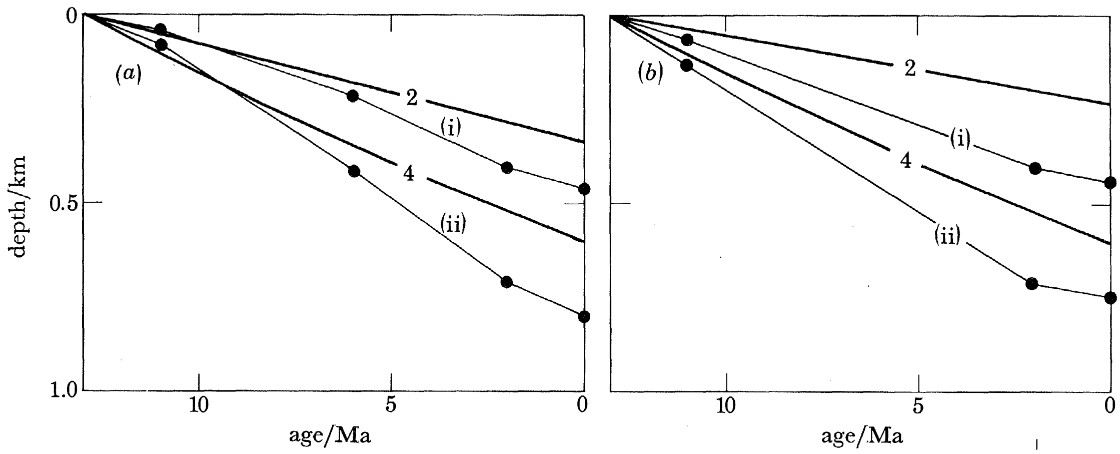

FIGURE 5(*a*). The inferred basement subsidence for the mean of 12 holes in the Great Hungarian Plain (Sclater *et al.* 1980), assuming a basement 2 km below base of base Neogene (i) and basement at base Neogene (ii) compared with the thermal subsidence expected after twofold to fourfold uniform stretching. (*b*) The inferred basement subsidence for the mean depths of various horizons in the Great Hungarian Plain from Horváth & Stegena (1977), compared with the thermal subsidence expected after uniform stretching; symbols as in (*a*).

Total sediment thickness varies from 2 to 5 km in most regions, and is similar to the sediment thickness in the peripheral basins. We believe that the subsidence is thermally controlled and that the initial subsidence is not observed in this region. Figure 5 shows average subsidence curves from 12 selected wells across the Great Hungarian Plain and mean subsidence for various horizons in the Great Hungarian Plain (from Horváth & Stegena 1977) plotted against theoretical subsidence curves for twofold and fourfold stretching. Depending on how the sediments are decompacted, the observed subsidence can be explained as the thermal subsidence resulting from about fourfold stretching. Heat-flow measurements are also in good agreement with those predicted for fourfold stretching (table 1). If we assumed that the Pannonian area were initially high, and subsided to sea level during the isostatic adjustment phase, then fourfold stretching could give the observed basement subsidence due to thermal decay. However, the present crustal thickness is 20–25 km (Sollogub *et al.* 1973). Fourfold stretching implies an original crustal thickness of 80–100 km, and an initial unloaded subsidence of greater than 5 km. It seems unlikely that the Pannonian region stood at an elevation of 5 km before Miocene extension since at least part of the Pannonian region was a site of deposition until the late Oligocene (Körössy 1964). Furthermore, early and mid-Miocene extension in the Pannonian has been estimated at no more than 100–150 km (Burchfiel 1980). Fourfold stretching requires 300–350 km of extension. This is also unreasonable and implies that simple stretching is not the complete explanation for the observed subsidence. In essence, the subsidence and heat flow from the central basins require the rapid shallowing of the isotherms in the mid-Miocene without any initial subsidence or more than twofold stretching.

Additional sources of heat

To explain the observations it is necessary to introduce extra heat into the lithosphere during or immediately before the initial stages of basin formation. The simplest way of doing this is to assume that the initial conditions were incorrect and that the lithosphere was hotter than the equilibrium condition assumed before stretching. It is unlikely that this extra heat was carried upwards by conduction because conductive heating requires tens of millions of years and the lithosphere must have been fairly cool until the late Oligocene because the elevation of this region was low until this time.

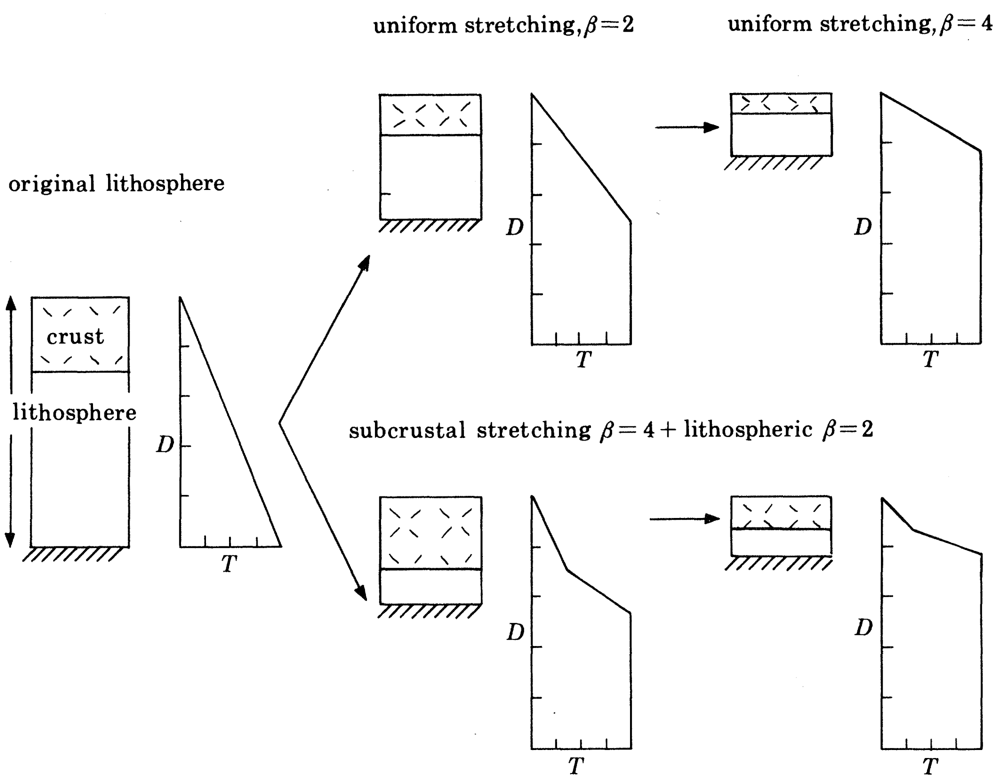

Figure 6. A diagram outlining how lithospheric subcrustal stretching and then stretching can result in almost the same temperature profile as extensive uniform stretching. This enables high-temperature material to be brought close to the surface without excessive crustal thinning, thus removing or reducing the initial subsidence. The same effect could also be created by lithospheric erosion from below.

This extra heat might be introduced into the lithosphere by first melting or eroding the lower lithosphere from below and then extending the entire lithosphere by a factor of two. Alternatively, the extra heat could be derived by subcrustal extension, which does not involve the crust (figure 6). The additional heat results in a temperature profile which is similar to that obtained by fourfold stretching, especially since short-wavelength variations in temperature die out very rapidly. This initial temperature distribution gives a thermal subsidence curve nearly identical to that for fourfold stretching, and after 5–10 Ma also yields similar values of surface heat flow. However, crustal thinning and crustal extension given by this mechanism indicate twofold extension. Furthermore, when the initial isostatic subsidence is calculated, the density changes caused by crustal thinning and by lithospheric heating approximately cancel each other out and the initial elevation change is negligible. These predictions are all in good agreement with observations in the Pannonian Basin (table 1). Whether this process of adding extra heat is directly related to subduction or termination of subduction along the Carpathian arc, or whether it is the result of complex movements of smaller individual blocks within the intra-arc region, is not clear. Since the timing of this event is constrained to be latest Oligocene to early–mid-Miocene, it seems probable that this process is directly linked to the general (twofold) extension that occurred at about the same time.

Conclusions

In summary, 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 twofold stretching of the lithosphere with some contribution from dyke intrusion in the Transcarpathian Depression. The rapid initial subsidence is an isostatic response to stretching that occurred during the early–mid-Miocene. The slower linear subsidence from then onward is thermally controlled.

The second type of basin lies in the central intra-Carpathian region, has high heat flow and a reasonably fast linear subsidence since the mid-Miocene. This type of basin may be formed by twofold stretching accompanied by subcrustal attenuation, melting or erosion of the lithosphere.

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Discussion

L. Stegena (*Department of Cartography, Lorand Eötvös University, Budapest, Hungary*). There are six deep refraction seismic profiles across the Pannonian basin, traversing also the Carpathians, Dinarides and the eastern Alps. Moreover, there are about 12 short profiles of the crust determined by reflexion seismic techniques. The general results of these measurements are: (1) the Pannonian basin has a thin crust (24–28 km); (2) it seems that the upper crust has a normal thickness (*ca.* 16 km) and the lower crust is thin (*ca.* 9 km); (3) there is a velocity inversion in the upper mantle (at a depth of 57 km, from 9.1 to 7.8 km/s). Taking into consideration the magnetotelluric measurements, seismological studies and deep-temperature calculations, it is believed that the velocity inversion coincides with the lithosphere–asthenosphere boundary.

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