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Rheology predictions across the western Carpathians, Bohemian massif, and the Pannonian basin: Implications for tectonic scenarios

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Abstract. On the basis of extrapolation of failure criteria, lithology, and temperature models, we predict the rheology of the lithosphere for several sections through the Carpathians and surrounding regions. Our models show significant lateral variations in rheology for the different tectonic units, with important implications for the tectonic evolution. The rheologically strong lithosphere of the Polish Platform area contrasts with the weak lithosphere of the Pannonian basin, indicating that the arcuate shape of the Carpathian orogen is primarily caused by an inherited curvature of an ancient embayment in the foreland, with the Pannonian units passively filling the space. The Polish Platform and the Moesian Platform exhibit a similar rheological anisotropy caused by NW-SE trending weakness zones paralleling the Tornquist - Teisseyre zone. This anisotropy was the main controlling factor on the behavior of the lithosphere in this area since Cadomian times, as documented by the geological evolution of the Sudety Mountains and the Mesozoic Polish Trough, including the Late Cretaceous Alpine inversion and the Neogene development of the Carpathian foreland. This rheological anisotropy appears to have a major controlling impact on the development of at least the eastern part of the European lithosphere. Rheology predictions for the Bohemian massif support the idea that the rigid lithosphere of the Bohemian massif governed the bending of the Alpine-Carpathian transition zone, expressed in the large-scale wrench movements opening the Vienna basin. In the foreland area, detachment levels are predicted for upper and lower crustal levels, leading to a decoupling of crustal and subcrustal flexure in most areas. Comparison with basin formation models indicates that our predictions for effective elastic thickness (EET) are similar to those derived from flexural models for the foreland area. Also, EET predictions from extensional basin models in the Pannonian region yield values close to our findings.

1. Introduction

The Carpatho-Pannonian region provides a key area to study the influence of different parameters on the rheology of the lithosphere. In a relatively small area, many different thermotectonic units occur. Many extensional basins characterize the young and hot Pannonian lithosphere, whereas the young Carpathian-thickened crust shows mainly strike slip related basins. The thermotectonically old lithosphere underlying the foreland area on the Bohemian massif and the European platform area form a sharp rheological contrast to the former two lithospheric units.

Furthermore, the abundance of geophysical data such as deep seismic reflection profiles, gravity surveys, and surface heat flow data provide valuable constraints on tectonic models of the area. To this purpose, we selected two profiles. The westernmost profile is based on the Deep Seismic Section VI (DSS VI) [*Beránek and Zátopek,* 1981], running WNW-ESE, starting in the Czech Republic, crossing the Bohemian massif, the Vienna basin, and the Malé Karpaty Mountains, and ending in the Danube basin (NW Pannonian basin) (Figure 1). The eastern profile, which includes deep seismic line 2T [*Tomek et al.,* 1989], starts on the Polish Platform, crosses the foreland basin and the western Carpathians, and ends in the Pannonian basin (Figure 1).

Finally, the inferences from many generations of numerical basin models for both the foreland and the extensional hinterland area provide independent constraints on lithospheric rheology. Lateral and temporal changes in lithosphere rheology have been documented to have pronounced effects on lithosphere dynamics [*Lankreijer et al.,* 1997; *Sachsenhofer et al.,* 1997]. Therefore rheological constraints on the proposed geodynamic models for the area are important.

Several authors [*Burov and Diament,* 1995; *Ranalli and Murphy,* 1987] have documented the methods for calculating lithospheric rheology during the last few years. Incorporating rheology predictions into tectonic models has yielded important constraints on those models [*Bassi,* 1995; *Buck,* 1991]. Incorporating lithosphere rheology into regional geodynamic and tectonic studies has only recently started [*Cloetingh and Banda,* 1992; *van Wees and Cloetingh,* 1996]. This approach allows a better quantitative understanding of the role of the lithosphere in tectonic processes such as basin formation and continental collision. We use a fully two-dimensional approach to predict rheology along selected lithospheric profiles, thus allowing a detailed study on the nature of variations in rheology across different tectonic units. Our approach predicts effective elastic thickness (EET) variations and detachment levels in the lithosphere, which are validated by comparison with EET estimates derived from other tectonic modeling techniques and which can be compared to interpretations of deep reflection seismic sections.

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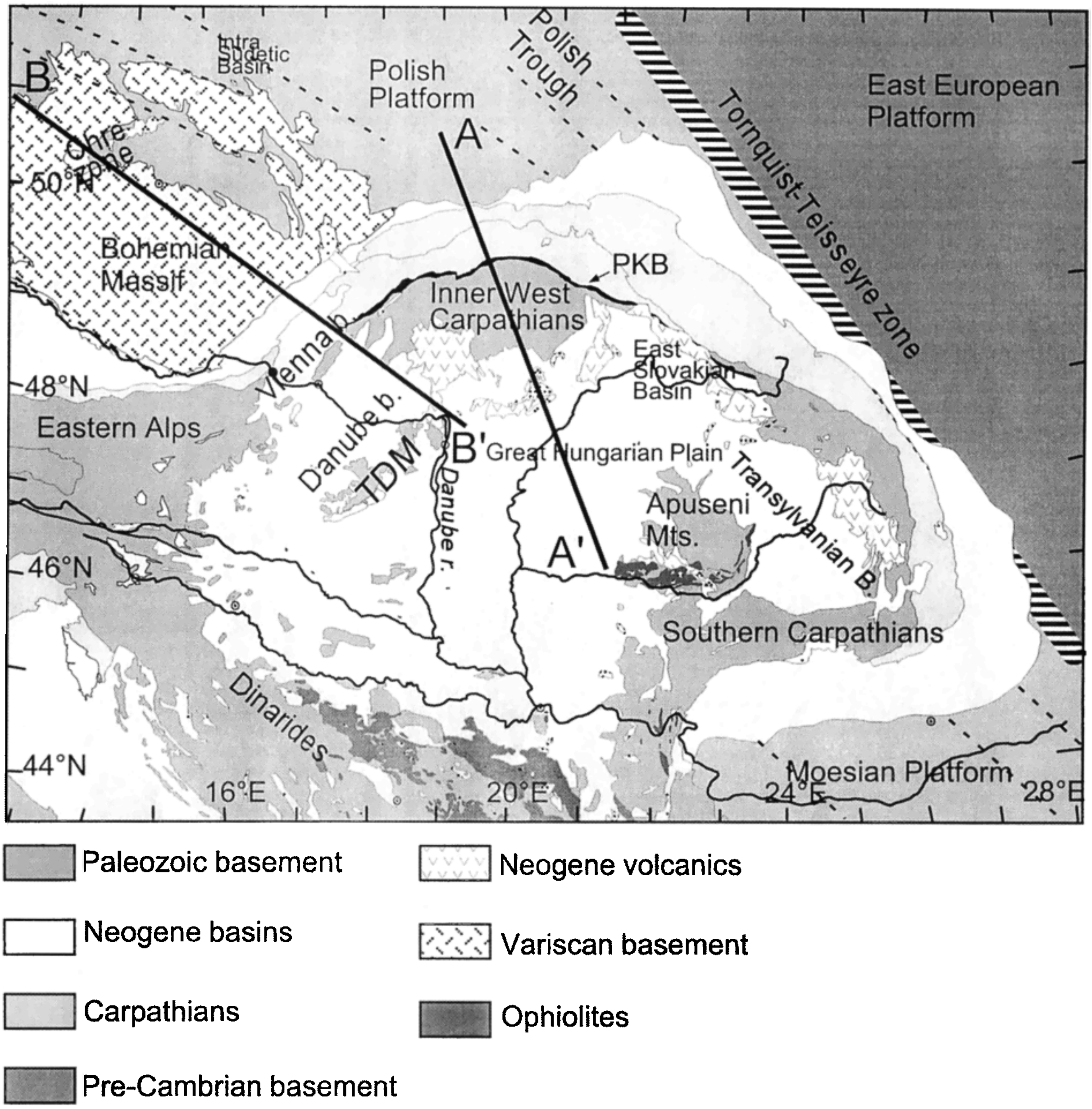

Figure 1. Tectonic map of central Europe. Solid lines show the location of the lithospheric crosssections A-A' and B-B'. TDM, Trans-Danubian Mountain Range; PKB, Pieniny Klippen belt.

2. Tectonic setting

The study area (Figure 1) allows the study of many different tectonic units in close spatial distribution. Tectonic units of different thermotectonic age, lithologic stratification, and crustal and lithospheric thicknesses cause significant variations in rheology with important implications for the tectonic behavior of each unit.

2.1. Bohemian Massif

The Bohemian massif (Figure 1) forms a Variscan and possibly Cadomian core [*Zoubek and Malkovský*, 1974], with an Alpine overprint, resulting in a cold and thickened crust. Surface heat flow values are typically low on the Bohemian massif (45 - 60 mW m^{-2}), especially in the Moldanubian unit.

The Bohemian massif appears to have played an important role in the bending of the Alpine-Carpathian junction and the associated strike-slip motions which opened the Vienna basin [*Royden*, 1988]. The rigidity of the Bohemian massif has a pronounced effect on the width of the foreland basin. Crustal-scale NE-SW strike slip faults of Hercynian age characterize the Saxothuringian part of the Bohemian massif and form major terrane boundaries [*Mahel' and Malkovský*, 1984]. NW-SE trending faults are mainly found in the central Bohemian massif [*Franke et al.*, 1993]. Tertiary reverse movements along the NW-SE trending Franconian lineament are evidenced from fission-track analyses of the German Kontinentales Tiefbohrungsprogramm (KTB) borehole [*Coyle et al.*, 1997]. Quaternary faulting has been observed in the Bohemian massif (Diendorf Fault) [*Hejl et al.*, 1997].

2.2. Polish Platform

The Polish Platform is of Precambrian age [*Znosko*, 1974]. Crustal-scale NW-SE strike-slip zones (Elbe-Hamburg fracture zone, Odra Fracture zone, Main Intra-Sudetic fault, and Lusatian main fault), parallel to the Tornquist-Teisseyre zone crosscut the Polish foreland. A similar set of NW-SE trending shear zones can be observed in the Moessian platform. These weakness zones create a large-scale anisotropic fabric, governing the rheologic behavior of the Polish platform through time. The Polish foreland shows low surface heat flow values and intermediate crustal and lithospheric thicknesses. The Polish Platform is characterized by low surface heat flow values (~ 60 mW m^{-2}). Foreland basin modeling [*Zoetemeijer et al.*, 1999] predicts values for the effective elastic thickness (EET) in the order of 15 km, to explain the bending of the lithosphere along the western Carpathian foreland.

2.3. Western Carpathians

The western Carpathians are the northernmost spur of the central European Alpides. The formation of its structure was influenced by complex processes such as convergence, lateral displacement, collision suturing, accretion, and transpression - transtension [e.g., *Andrusov*, 1968; *Plašienka*, 1997; *Soták*, 1992]. The fundamental feature of the western Carpathians is their nappe structure [e.g., *Maheľ*, 1974; *Sandulescu*, 1994; *Sandulescu and Bercia*, 1974]. The western Carpathian lithosphere has been thermotectonical rejuvenated during the volcanic episodes associated with the Carpathian convergence (22-15 Ma). The mean value of surface heat flow in the central western Carpathians amounts to 60 - 70 mW m^{-2}. However, the heat flow density in the Neogene Danube basin is noticeably higher (70 - 80 mW m^{-2})[*Bodri*, 1981; *Čermák*, 1994].

2.4. Pannonian Basin

The Pannonian basin is a young basin (17-10 Ma), with associated high surface heat flow values (85 - 95 mW m^{-2}). Numerical basin models for the Pannonian basin predict low values for the EET of the order of 5-7 km [*Lankreijer et al.*, 1995; *van Balen and Cloetingh*, 1995]. Earthquake focal depths are limited to the upper 6 km of the crust of the Pannonian basin [*Horváth and Cloetingh*, 1996], supporting this thin upper crustal strong layer. The thickness of the Neogene fill in the Pannonian basin amounts to 9 km in the deepest troughs, but on average, the Neogene sequences are only 2-3 km thick. Lithospheric and crustal thickness maps [*Horváth*, 1988; 1993; *Szafian*, 1999] show a close spatial coupling between thinned crust and the main depocenters in the Neogene Pannonian basin. The lithosphere in the Pannonian area is extremely thin (60 km), giving rise to very high crustal temperature in the region [*Dövényi and Horváth*, 1988; *Lenkey*, 1999]. Crustal thickness amounts to 25-28 km [*Kilènyi et al.*, 1989; *Lillie et al.*, 1994]. Neogene core-complex style deformation along the western margin of the Pannonian basin [*Tari*, 1993] indicates a weak lower part of the crust in these areas during Neogene deformation [*Sachsenhofer et al.*, 1997].

2.5. Vienna Basin

The Vienna basin is a Neogene extensional pull-apart basin located on top of Alpine-Carpathian thrustsheets. Large-scale sinistral strike-slip faults decouple the basin from the surrounding lithosphere. The discussion on the penetration depth of the basin-bounding faults is still going on [*Lankreijer et al.*, 1995; *Royden*, 1985; *Wessely*, 1992].

The Vienna basin opened during Karpatian / early Badenian times (17.5 - 15.5 Ma) and shows a passive postrift subsidence since Sarmatian (14 Ma) times. Changes in the paleostress field, in Pannonian and Pliocene times, are documented by microtectonic fabric analyses [*Bada*, 1999; *Decker and Peresson*, 1996; *Fodor et al.*, 1999].

Paleogeography of the basin [*Seifert*, 1992] indicates an isostatic compensation, where the basin is decoupled along the master faults. A flexural response to loading of the basin [*Watts et al.*, 1982] predicts a general widening of the basin. The paleogeography of the Vienna basin [*Seifert*, 1992] shows a stable position of the paleocoastline through time. Measured surface heat flow values in the basin are relatively low (50 - 60 mW m^{-2}) [*Dövényi and Horváth*, 1988], but this could be due to the effect of blanketing by the sediment pile, which is in places more than 6 km thick.

Many different tectonic models have been proposed for the Carpatho-Pannonian area [*Csontos et al.*, 1992; *Ratschbacher et al.*, 1991a, b; *Royden*, 1988; *Soták*, 1992; *Tari et al.*, 1992]. It is clear that around 20 Ma several microplates filled the space inside the Carpathian arc. Collision ceased diachronously along the Carpathian belt between 22 Ma (in the west) and 5- 0 Ma (in the southern Carpathians). After 17 Ma, extension started in the Pannonian basin system, thinning the crust in general with a factor 1.6 and extending the

Table 1. Thermal Parameters

Parameter	Value
Thermal base depth, km	250
Depth increment, m	1000
Surface temperature, °C	0
Temperature at base of the plate (mantle melt temperature), °C	1300

Thickness, km	Density, kg m^{-3}	Conductivity, W m^{-1} K^{-1}	Capacity, J kg^{-1} K^{-1}	Heat Production, W m^{-3}	Skindepth, km
Upper crust	2650	2.5	1136	2.00×10^{-6}	10
Lower crust	2900	2.0	1029	0.50×10^{-6}	0
Mantle	3300	3.5	1212	0.0	0

Table 2a. General Properties Used for Rheology Models

Definition	Parameter	Value
Acceleration of gravity, m s^{-2}	g	9.81
Universal gas constant, J mol K^{-1}	R	8.314
Surface heat flux, W m^{-2}	q_s	30-100
Temperature base lithosphere, °C	T_m	1300
Static friction coefficient	f_s	0.6
Strain rate, s^{-1}	$\dot{\varepsilon}$	10^{-15}
Hydrostatic pore fluid factor (ρ_w/ρ)	λ	≈0.35

subcrustal lithosphere by a much higher factor [*Lankreijer et al.*, 1995; *Lenkey*, 1999; *Royden and Dövenyi*, 1988].

3. Method

A dependence of rock strength on temperature and pressure has been demonstrated by laboratory experiments [e.g., *Goetze and Evans*, 1979; *Ranalli and Murphy*, 1987]. In the upper region of the mechanically strong part of the lithosphere, rheology is generally governed by brittle failure (Byerlee's law). At temperatures exceeding roughly half the melting temperature of rock, ductile creep processes become the dominant deformation mechanism [*Carter and Tsenn*, 1987]. Therefore the strength in the lower part of the lithosphere and the lower parts of the Earth's crust is mainly governed by the temperature distribution. *Ord and Hobbs* [1989] argue that there must be a breakdown stress for Byerlee's brittle failure law. They infer a value of ~ 260 MPa for this breakdown stress.

Extrapolation of flow laws and laboratory failure criteria [*Brace and Kohlstedt*, 1980; *Byerlee*, 1978], adopting estimates for tectonic strainrates and thermal gradients, provides a firs-order description for the strength distribution within the lithosphere. For each depth interval, strengths for both brittle and ductile deformation are calculated (taking into account the brittle failure breakdown stress), with the lesser of these representing the limiting strength (yield strength) of the lithosphere at that particular depth interval [e.g., *Beekman*, 1994; *Burov and Diament*, 1995; *Cloetingh and Burov*, 1996; *Ranalli*, 1995]. Critical input data for the prediction of lithospheric strength are crustal composition and thermal structure of the lithosphere (Table 1).

Furthermore, predictions of lithospheric strength are strongly influenced by the adopted strain rate. We adopted a bulk lithospheric strain rate for our calculations of $\dot{\varepsilon} = 10^{-15}$ s^{-1}, which is commonly observed in extensional and compressional settings [*Carter and Tsenn*, 1987; *Okaya et al.*, 1996]. Observations on strain rates indicate a range of 10^{-17} s^{-1} < $\dot{\varepsilon}$ < 10^{-12} s^{-1} [*Carter and Tsenn*, 1987; *van den Beukel*, 1990]. Faster strain rates produce greater predicted strengths. Strain rates are typically assessed within the accuracy of an order of magnitude. Such uncertainties in estimation change the predicted lithospheric strength by no more than 10%.

Although the construction of lithospheric strength profiles invokes a number of intrinsic uncertainties, the results of many recent studies support the extrapolation of microphysical models from a laboratory scale to a lithosphere scale [e.g., *Burov and Diament*, 1995; *Cloetingh and Banda*, 1992; *Lankreijer et al.*, 1997; *Ranalli and Murphy*, 1987]. Furthermore, hydraulic fracture tests in the KTB borehole (SW Germany) demonstrate that such an extrapolation is valid for the tested interval (4-6 km) [*Zoback et al.*, 1993a].

We adopted a five-layer rheologic model for the lithosphere along our profiles, consisting of a sedimentary layer (where present), a quartzite layer (representing superficial sedimentary nappes, i.e., Tatric nappe), a granite layer (representing the upper crust), a diorite layer (for the lower crust), and a dunite layer representing the lithospheric mantle. Mantle xenolites indicate a mantle composition beneath the study area consisting of Lherzolite and Habsburgite [*Downes and Vaselli*, 1995]. Tables 2a and 2b summarize the material properties for the adopted lithologies. We adopted a wet rheology for these lithologies, since most recent studies support "wet" rheology rather than the stronger "dry" variant [*Beekman et al.*, 1994; *Cloetingh and Burov*, 1996; *Lankreijer et al.*, 1997].

3.1. Temperature Model

A lithology model, based on gravity and geological interpretation of deep reflection seismic [*Bielik et al.*, 1995; *Tomek et al.*, 1987] served as a base for assigning thermal properties to individual blocks in a model of the lithosphere [*Kutas et al.*, 1989; *Majcin*, 1993]. The temperature distribution was calculated following *Kutas et al.* [1989] and *Majcin and Tsvyashcheko* [1994].

The stationary component of the temperature field is determined as a result of both the effect of heat sources and of background heat flow density from the lower mantle. A second component of the thermal field corresponds to thermotectonic rejuvenation of the area [*Čech*, 1988; *Horváth et al.*, 1989; *Jiříček*, 1979; *Kováč et al.*, 1993].

Table 2b. Material Properties used in Rheology Models

	Upper Crust		Lower Crust		Mantle	
	Granite Dry	Granite Wet	Diabase Dry	Diorite Wet	Dunite Dry	Dunite Wet
Density (ρ), kg m^{-3}	2700	2700	2900	2900	3300	3300
Young's modulus (E), Gpa	50	50	70	90	70	70
Poisson's ratio (ν)	0.25	0.25	0.25	0.25	0.25	0.25
Power law exponent (n)	3.3	1.9	3.05	2.4	4.5	3.6
Power law activation energy (E_p), kJ mol^{-1}	186	140	276	212	535	498
Pre-exponential constant (power law) (A_p), Pa^{-N} s^{-1}	3.16×10^{-26}	7.94×10^{-16}	6.31×10^{-20}	1.26×10^{-16}	7.94×10^{-18}	3.98×10^{-25}

The brittle failure Function is $\sigma_{brittle} = \alpha \rho g z\,(1-\lambda)$, where $\alpha = R-1/R$ for normal faulting, R-1 for thrust faulting, and $(R-1)/[1+\beta(R-1)]$ for strike-slip faulting and $R = [(1+f_s^2)^{1/2} - f_s]^{-2}$. The power law creep function used is $\sigma_{creep} = (\dot{\varepsilon} / A_p)^{1/n} \exp[E_p / nRT]$, after *Carter and Tsenn* [1987] and *Goetze and Evans* [1979]

Table 3. Comparison Between Tectonics Units

Tectonic Unit	HF, mW m^{-2}	EET, km	Thermo Tectonic Age, Ma	Bouguer, mGal
Bohemian Massif (core)	45-50	20-40	660-550	-10 - 20
Bohemian Massif (Tepla-Barrandian)	55-60	8-12	320-260	-20 - 10
Foreland area (Polish Platform)	50-60	7-8	320-260	0 - 20
Vienna Basin	50-60	18-22*	17-15	-40 - 0
Pannonian basin	70-100	5-10	17-14	-40 - 20
West Carpathians	60-70	15-23	22-7	-50 - 0

HF, Heat flow; EET, effective elastic thickness.

* note that the measured heatflow for the Vienna basin reflects the surface heatflow only. The thick sedimentary fill of the basin isolates the basement; therefore thermal and rheological predictions yield estimates that are too cold, i.e., too strong.

The reliability of a temperature model depends mainly on the accuracy and density of measurements of heat flow density in the surroundings of the profile [*Hurtig et al.*, 1992]. The reliability of the temperature model was further increased by fitting the lithosphere thickness along the profile to seismic data [*Babuška et al.*, 1988]. Therefore the relative inaccuracy is not in excess of 10%. Temperature calculations for the Bohemian massif [*Čermák*, 1994], yield values very close to ours. Differences are mainly in temperature predictions for the deepest parts of the model (>150 km), where the effect on rheology is minimal.

3.2. Gravity Model

Gravity modeling along section A-A' was performed using the GM-SYS™ programs of Northwest Geophysical Associates, Inc. Thicknesses of the Pannonian basin sediments are derived from the maps of *Kilényi et al.* [1991]. The thickness of the sediments in the Carpathian foreland and the thickness of Tatricum are taken from deep seismic profile 2T [*Tomek et al.*, 1987; 1989]. Depths of the upper to lower crust boundary are deduced from *Bielik et al.* [1990]. Depths to the Moho and lithosphere / asthenosphere boundary were taken from *Horváth* [*1993*] and *Babuška et al.* [1988], respectively. Density contrasts of the different bodies are similar to those of *Lillie et al.* [1994], *Szafián et al.* [1997] and *Szafian* [1999].

For calculation of the gravimetric model along section B-B', the method of *Talwani* [1973] was used. Density contrasts for the anomalous bodies are relative to the reference model defined by *Bielik et al.* [1994]. The interpretation was based

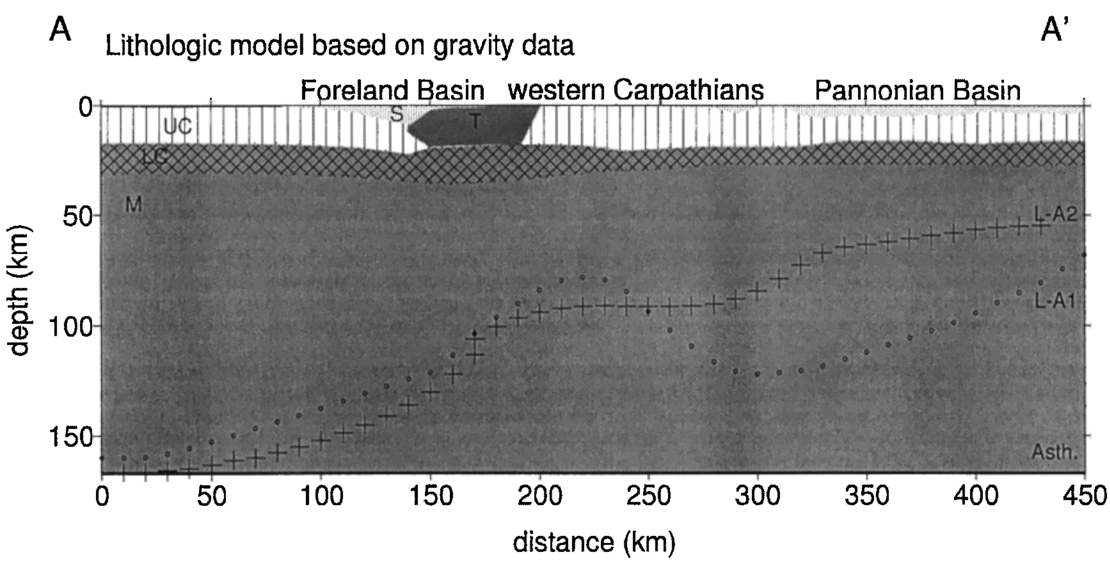

Figure 2. Lithospheric crosssection A-A' through the western Carpathians (For location, see Figure 1.). Lithologic differentiation is based on gravity modelling [*Bielik et al.*, 1994]. Lithologic units as used for rheology calculations: S, sediments; T, Tatricum (quartzite); UC, upper crust (granite); LC, lower crust (diorite); M, upper mantle (olivine); Asth, asthenosphere; L-A1, lithosphere - asthenosphere boundary based on gravity model; L-A2, thermally defined lithosphere – asthenosphere boundary based on model in Figure 3. Lithospheric crosssection cuts, from north to south, the following tectonic units: between km 0 and 130 is the Polish Platform, forming the substratum for a well-developed foreland basin of the outer Carpathians. Between km 130 and 230 is the central western Carpathians, consisting of a series of nappes related to different Alpine convergence events. Between km 230 and 270 is the inner western Carpathian system, probably thermotectonically rejuvenated by Neogene convergence-related intrusions. Farther southward, between km 270 and 450, is the Pannonian basin, consisting of several individual mountain ranges (Matra – Bükk) and grabens (Szolnock and Békés).

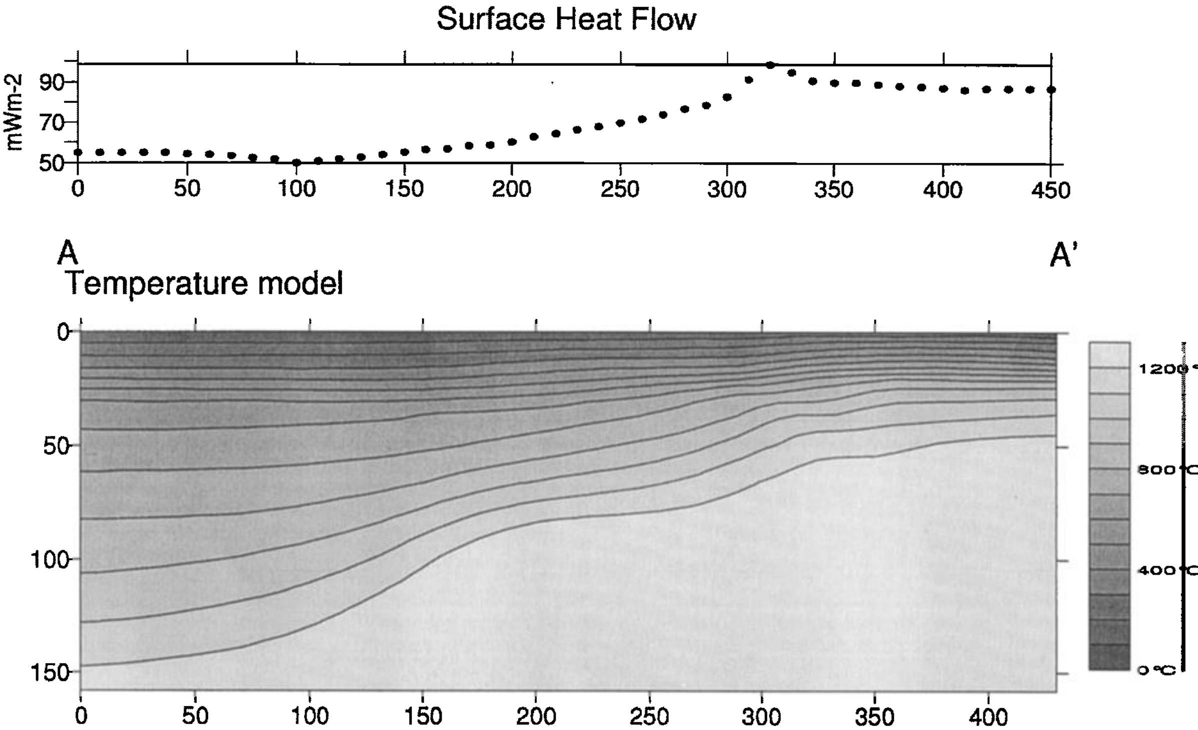

Figure 3. Temperature field calculated for the studied crosssection A-A'. Surface heat flow is from *Hurtig et al.*, [1992]. Crustal heat production is taken into account (model after *Majcin*, [1993] and *Kutas et al.* [1989]). Thermophysical properties are assigned to lithologic units defined by gravity modeling and deep seismics [*Bielik et al.*, 1995; *Tomek et al.*, 1987].

on the following input data: information on the geologic structure to a depth of ~ 5 km, densities of rocks defined according to *Eliaš and Uhmann* [1968], *Ondra and Hanák* [1981], the Moho after *Beránek* [1980], and the seismic velocity after *Beránek and Zátopek* [1981].

3.3. EET

According to *Burov and Diament* [1995], the effective elastic thickness of the continental lithosphere can be calculated from the combined effect of the thicknesses (h_i) of the individual strong layers:

$$EET = \left(\sum_{i=1}^{n} \Delta h_i^3\right)^{1/3}$$

Definition of the exact thickness of the strong layers (values of h_1, h_2, etc.,) remains a matter of debate [*Burov and Cloetingh*, 1997; *Burov and Diament*, 1995; *Cloetingh and Burov*, 1996]. The criterion of *Ord and Hobbs'* [1989] for Byerlee brittle failure breakdown, at ~ 260 MPa, would imply that EET values cannot be higher than 26 km, under the condition that we adopt a pressure-scaled minimum yield strength or minimum vertical stress gradient of 10 Mpa km^{-1}. The conversion of strength predictions to EET values is useful since the latter can be directly compared to inferences from basin modeling and lithospheric flexure studies, providing a means to quantitatively test the merits of the individual modeling techniques (Table 3).

4. Lateral Variations in Lithospheric Properties Along the Profiles

4.1. Western Carpathian Profile

The profile through the western Carpathians (A-A') (Figures 2 and 3) can roughly be subdivided into three zones i.e., the Polish platform - foreland area, the Carpathians, and the Pannonian region. The Polish platform - foreland area is characterized by a relatively thick crust, overlain in the foreland area by a thick sedimentary pile (6 km). Surface heat flow values are intermediate (40 - 60 mW m^{-2}).

In the Carpathian unit, the crustal thickness is notably increased (32 - 35 km). Our model assumes a large body of Tatric low-density material to form the upper crust in the outer part of the central western Carpathians.

The Pannonian region is characterised by a relatively thin crust (25-28 km) and high surface heat flow values (90 - 100 mW m^{-2}). The thickness of the lithosphere is also very small (50 - 80 km).

Profile A-A' (across the western Carpathians) shows a general decrease in strength toward the Pannonian basin (Figure 4). The Polish foreland area (between km 0 and 130) shows a horizontal rheological stratification of the lithosphere. Mechanically strong behavior is predicted for the upper part of the crust, the uppermost part of the lower crust, and the uppermost part of the mantle. The weak lower part of the lower crust is predicted as the most obvious detachment level;

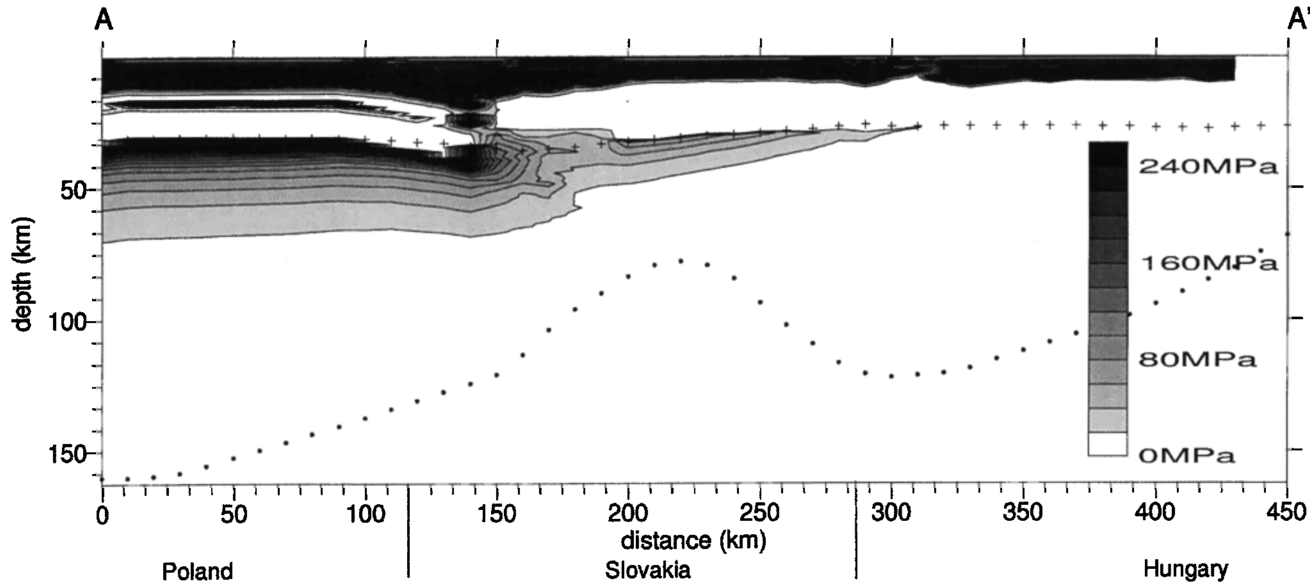

Figure 4. Yield-strength contour plot for compressional deformation. For rheological cross-section A-A', at a strain rate of $\dot{\varepsilon} = 10^{-15}\,s^{-1}$. A clear rheological stratification of the lithosphere is visible. In the foreland area, three individual strong layers are predicted, whereas in the Pannonian part of the profile only one thin strong layer is predicted. Clearly visible is also the lower crustal detachment level in the foreland area. Moho and base of the lithosphere are indicated by crosses and dots, respectively [after *Bielik et al.*, 1995].

possibly also the lower part of the upper crust will act as a detachment level for the adopted strain rate of $\dot{\varepsilon} = 10^{-15}\,s^{-1}$. The combined elastic effect of the three strong layers in this region will govern the flexural behavior of the foreland in this region. An EET of 12 km is predicted for this region on the basis of our strength predictions. Flexural models for this area [*Krzywiec and Jochym*, 1997; *Zoetemeijer et al.*, 1999] predict values between 6 and 15 km for this area.

In the Carpathian part of the profile (km 150-250), lower crustal strength completely disappears as a result of crustal thickening and increased crustal temperatures. The lithospheric strength gradually decreases toward the SE along this profile; this is a direct result of the increasing temperatures toward SE and the corresponding decrease of the (thermally defined) lithospheric thickness. The EET for this region is mainly governed by the thickness of the upper crustal brittle part; we predict an EET of 15 - 23 km. There are no independent estimates for EET in this area.

The Pannonian part of the profile (km 300-450) displays a typical Pannonian rheological structure, characterized by one relatively thin strong layer in the uppermost 10 km of the crust and a complete absence of strength in the lower crust and lower lithosphere. The extreme weakness of the Pannonian lithosphere is a direct result of the high heat flow density and it is related to the extremely shallow asthenosphere in this area. EET values of 5 - 10 km are predicted. Results from extensional basin modeling in the Pannonian basin yield EET values of 5 -10 km [*van Balen et al.*, 1999]. Rheology predictions based on a technique similar to that adopted here yield EETs of 8 km for the Romanian part of the Pannonian basin [*Lankreijer et al.*, 1997]. Earthquakes in the Pannonian basin are limited to the upper 6 – 10 km of the crust [*Szíros et al.*, 1987], supporting the interpretation of a thin strong layer.

4.2. Bohemian Profile

The profile through the Bohemian massif (B-B') (Figures 5 and 6) can be subdivided into three main different units, based on crustal structure: the Krušné hory - Saxothuringian zone, the Bohemian core zone (includes Tepla-Barrandian, Moldanubicum, and Brunovistulicum) and the Carpathians - Pannonian zone [*Bielik et al.*, 1994]. The Krušné hory - Saxothuringian zone is characterised by a relatively thick upper crust (16-20 km) in comparison to the thin lower crust (12-14 km), resulting in negative Bouguer anomalies. Intermediate surface heat flow values (60 -70 mW m^{-2}) are found in this unit, whereas the Cretaceous basin shows slightly increased heat flow values. The Ohre zone, a whole crustal fault zone, is characterized by a steep gravity gradient and separates this unit from the Bohemian core unit.
The Bohemian core unit shows a thicker crust (30 - 39 km), whereas the upper crust is remarkably thinner (9 - 15 km). The gravity effect of the depressed Moho is almost completely compensated by the presence of high-density rocks in the upper and lower crust [*Bielik et al.*, 1994]. The Bohemian core typically shows low heat flow values (40 - 60 mW m^{-2}).

The Carpathian - Pannonian area is characterized by a reduced crustal thickness (25 - 28 km) and a thinner lower crust in comparison with that of the Bohemian unit. The upper crust is ~ 16 km thick in this unit. The Pannonian unit especially is characterized by extremely high surface heat flow values (85 - 95 mW m^{-2}).

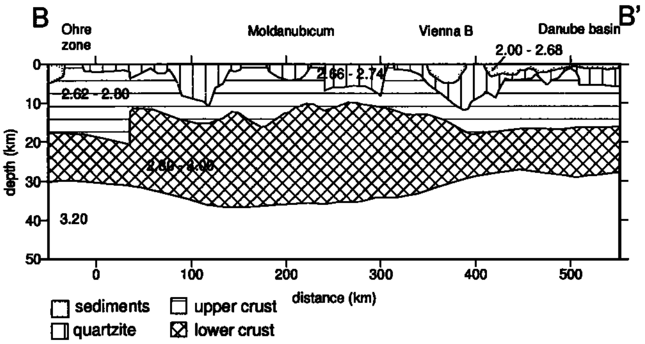

Figure 5. Lithologic stratification of crosssection B-B' (For location, see Figure 1) based on gravity models of Deep Seismic Section VI, [*Bielik et al.,* 1994]. Densities are in kg m^{-3}. The profile shows three different tectonic units, separated by the Ohře zone and the Peripienian lineament, which are interpreted as whole crustal faults. The Ohře zone marks the sharp transition between the Krušné hory - Thuringian region (characterized by a thick upper crust and a relative thin lower crust) and the Bohemian core (with its thickened lower crust and thinned upper crust). The Bohemian core consists of the Teplá-Barrandian, the Moldanubicum, and the Brunovistulicum regions. The third region comprises the Carpathians and the Pannonian basin system (including the Vienna basin).

The profile through the Bohemian massif (B-B') shows a three-layer rheological stratification in the Saxothuringian part of the profile (km -50 - 0). EET values predicted for this area are 8 - 12 km.

Underneath the Cretaceous basin (km 0-50) and the Ohre zone, an absence of strength in the lower crust and mantle is predicted associated with the high surface heat flow values [*Čermák,* 1994] measured in this area (Figure 7). We predict EET's between 5 and 8 km for this area (Figure 8a and 8b).

The remarkable increase in thickness of the lower crust on the SE side of the Ohře fault is also visible in the predicted strength distribution, since lower crustal material is present at shallower depths, causing strong layers. This exceptionally thick and shallow lower crust continues along the entire Moldanubicum and is possibly a result of earlier crustal thickening.

The Moldanubicum (km 150 - 300) displays the mechanical strong core of the Bohemian massif. At the adopted strain rate, only a very thin zone in the lowest part of the lower crust allows detachment between crust and mantle. All other crustal layers are nondetached and behave as one single thick, rigid layer. Additionally, the lower lithosphere shows a very strong and deep keel, with a thickness of over 60 km. We predict EETs of the order of 20 – 40 km for this area. Extremely low heat flow density values, a shallow lower crust, and a cool mantle are the main causes for the predicted extreme values of lithospheric strength. The stiff behavior of the Bohemian massif allowed transmission of Alpine compressional stresses far from the orogenic front, causing inversions of the northern margins of the Bohemian massif (P.A. Ziegler, personal communication 1997).

In the Bohemian foreland area (km 300-350), we still observe the shallow lower crust, producing a strong upper layer in the lower lithosphere. The lower part of the lower crust is again a weak zone. The seismically observed Moho in

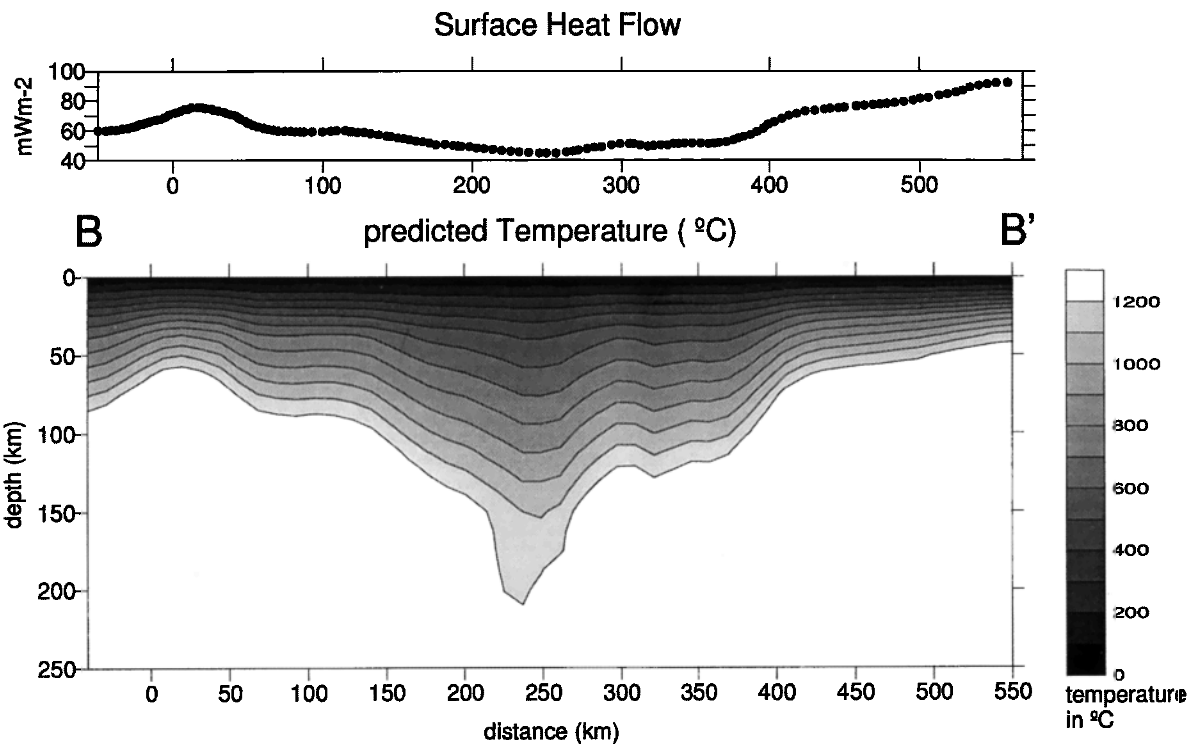

Figure 6. Temperature distribution along section B-B'. Model is partly based on the work of *Čermák* [1994]. Crustal heat production is taken into account. Surface heat flow and crustal structure are after *Čermák* and *Bielik et al.* [1994].

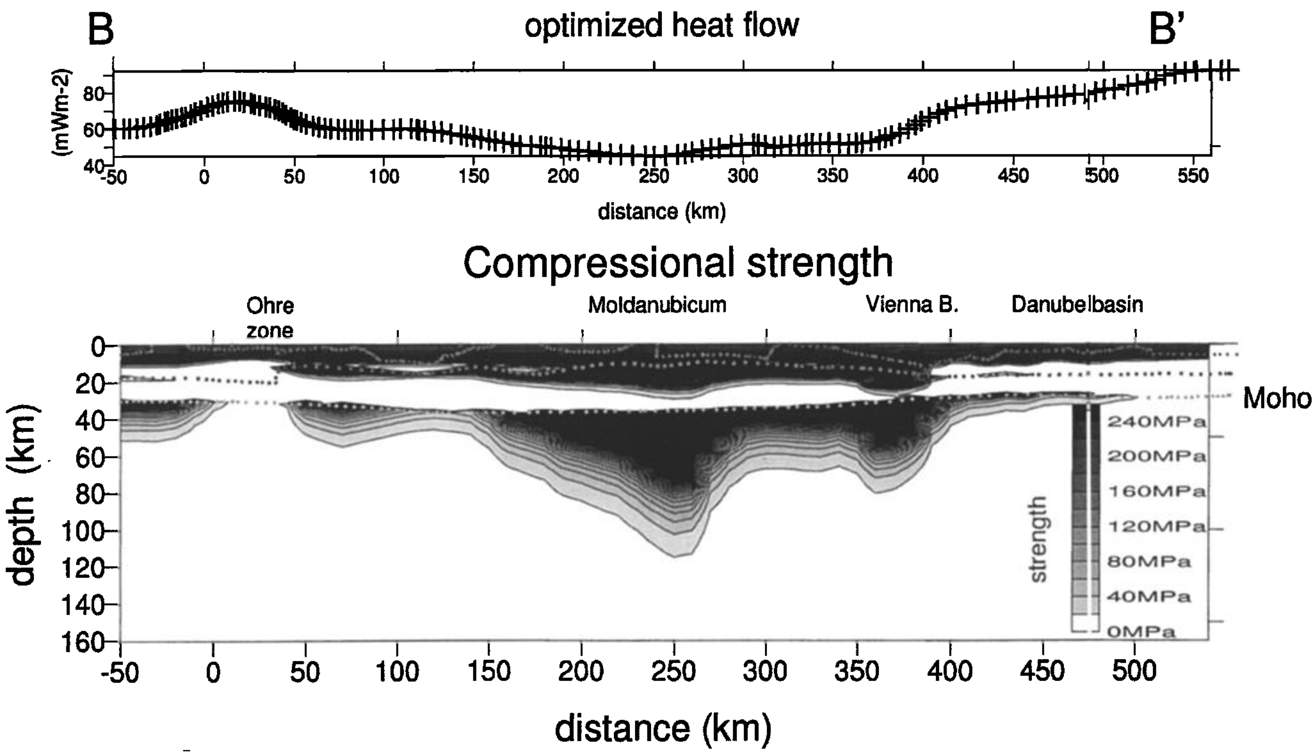

Figure 7. Yield-strength contour plot for compressional deformation for cross section B-B', at a strain rate of $\dot{\varepsilon}$ = 10^{-15} s^{-1}. A clear rheological stratification of the lithosphere is visible. The Moldanubian core of the Bohemian massif shows a strong keel, where crustal layers are mechanically attached to lower lithospheric layers. The strength rapidly decreases away from the strong center toward SE and NW. The Carpathian foreland area is, however, still relatively strong. The Vienna basin shows a remarkably strong lithosphere owing to the fact that we did not correct for the insulation effect of thermal blanketing of the over 6 km sedimentary fill of the basin. The Pannonian basin typically only displays lithospheric strength in the uppermost parts of the crust.

this area is supposed to be of very recent age, since it is not down-flexed, as is the case in all other Carpathian foreland areas [*Tomek et al.,* 1987; *Tomek and Hall,* 1993]. The predicted weak lower crusts possibly aided the process of the formation of a new Moho after flexure and possibly slab detachment. We predict EET values of 10 km for this area.

Strength values inferred for the Vienna basin are assumed to be overestimated, since the relatively low surface heat flow values do not take the thermal effect of sediment blanketing into account. Basement heat flow values are probably much higher than the used surface heat flow, since the basin is filled with over 6 km of Neogene sediments. EET values based on our strength estimates are 18 – 22 km.

Basin models show a southward increase of detachment depth in the Vienna basin [*Lankreijer et al.,* 1995]. However, since our profile crosses the basin at an unfavorable angle, no lateral changes can be observed in the rheology of the lithosphere underneath the Vienna basin.

The Danube basin, on the SE side of the profile, shows very low values of lithospheric strength, associated with the increased heat flow density and deeper lower crust. Only the upper part of the upper crust and the uppermost part of the lower lithosphere show some strength in our calculations. The EET is mainly governed by the thickness of the strong part of the upper crust and amounts to 8 km.

Seismic interpretations of the Danube basin [*Posgay et al.,* 1986; 1996; *Tari,* 1994; 1996] show SE dipping crustal detachments. Detachment along discrete deep faults is not completely in accordance with the predicted recent rheology. On the basis of the recent crustal strength predictions, a shallower detachment (at the base of the strong part of the upper crust) is expected. Furthermore, these detachments show a synsedimentary behavior for Karpatian - Sarmatian times and do not affect Pannonian strata. Therefore the deep detachments represent an earlier (Karpatian - Sarmatian) rheological situation.

The absence of lithospheric strength in the lower crust and upper mantle has major implications for basin models. Lithospheric loads, imposed by sedimentary fill of the basins overlying this extremely weak lithosphere, must be largely compensated in a local isostatic manner, that is, flexural support of the lithosphere is almost absent. However, the predicted rheology is only valid for the present situation, and caution must be taken in extrapolating these findings to previous times. The Pannonian part of the profile (km 500 - 550) shows the typical Pannonian rheology, a total absence of

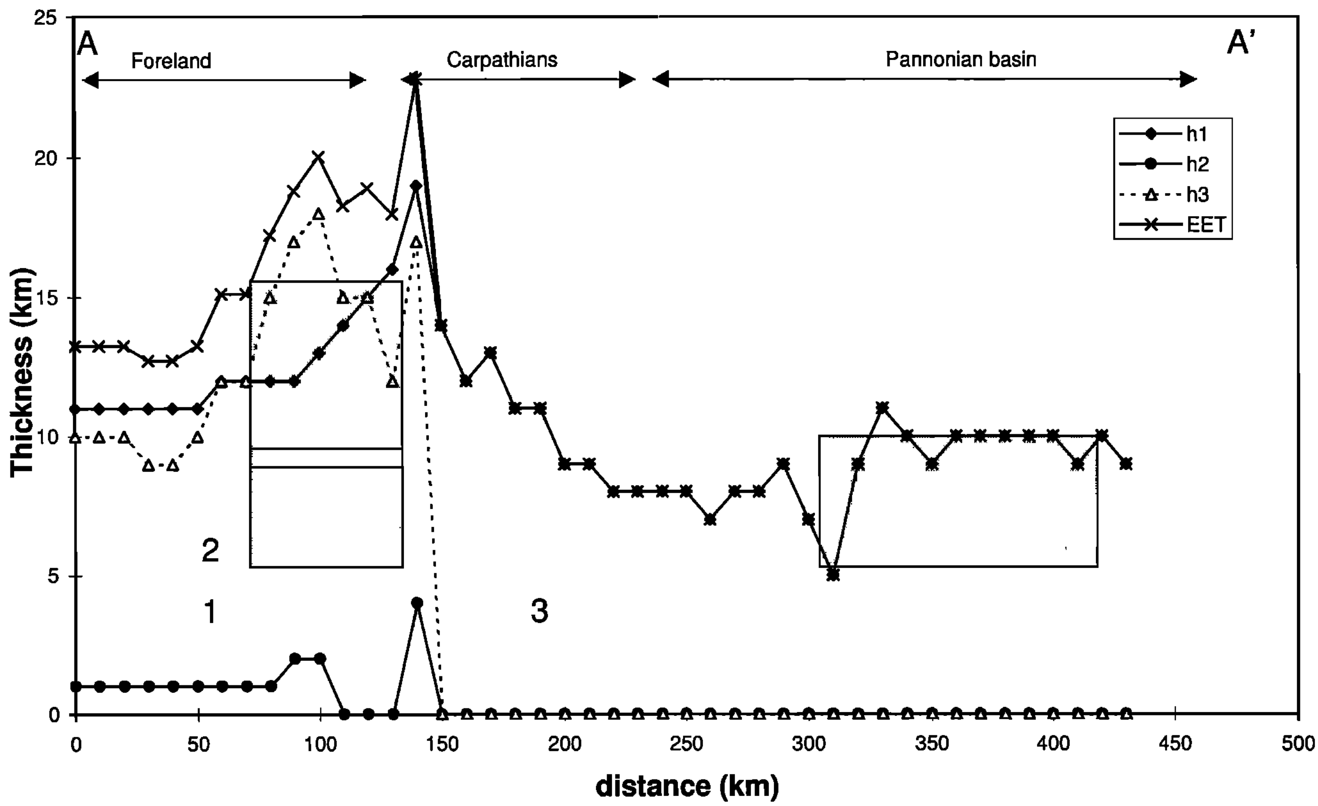

Figure 8a. Effective elastic thickness (EET)(T_e) distribution along profile A-A'. The thicknesses of the mechanically strong layers h_i are shown. The combined effect of n detached layers can be calculated using $Te = (\Sigma \Delta h_i^3)^{1/3}$ [*Burov and Diament,* 1995]. The EET is mainly governed by the thickness of the uppermost strong layer h_1. In the Polish Platform, a significant contribution to EET is also added by the strong part of the mantle (h_3). Boxes indicate independent EET estimates based on foreland basin models (EET 6-10 km [*Zoetemeijer et al.,* 1999] and EET 10-15 km [*Krzywiec and Jochym,* 1997]) and extensional basin models (EET 5-10 km, [*van Balen et al.,* 1999; *van Balen and Cloetingh,* 1995]).

lithospheric strength except for the uppermost few kilometers of the crust, similar to that observed in section A-A' and described by *Lankreijer et al.,* [1997] and *Lankreijer* [1998].

5. Discussion

5.1. Validation

As pointed out in section 4, EET predictions derived from alternative modeling techniques like extensional basin modeling and those from flexural modeling yield independent estimates on the lithospheric rheology in the studied area. The observed close fit between the rheology predictions obtained using the different methods makes us confident in our own rheology predictions, which not only predict EET but also allow identification of detachment zones.

The predicted EETs for the foreland areas do not take into account the weakening effect imposed by the bending of the lithosphere. The effect of far-field stresses causing weakening was also not taken into account, since the predictions reflect a static situation in the absence of actual deformation and stress. Incorporation of these effects requires limiting the calculations to a single well-defined tectonic scenario, with its intrinsic uncertainties. Furthermore, the complex feedback mechanisms operating in the relation between stress and strain through rheology do not permit such complex calculations. The stress field is directly influenced by the strength distribution, and the predicted rheology is partially dependent on the applied stress. Deformation induces direct geometrical changes, thus influencing the strength distribution. Additionally, strain hardening, weakening, or localization as a function of deformation is difficult to quantify in a kinematic model. Far-field stresses do probably play an important role in the areas where the lithosphere is weak, i.e., the central parts of the Pannonian basin system. In order not to make too many concessions on the spatial geometry of the system, a static model was used, rather than a dynamic model that would take into account the above mentioned processes.

Deformation velocities can be derived from extensional basin models [*Lankreijer,* 1998]. Typical extension values for the Pannonian basin system are of the order of β=1.6 [*Horváth et al.,* 1975; *Lankreijer et al.,* 1995; *Royden and Dövenyi,* 1988; *Sclater et al.,* 1980; *Stegena et al.,* 1975]. The duration of the rift period in the Pannonian basin system is ~ 2 Myr

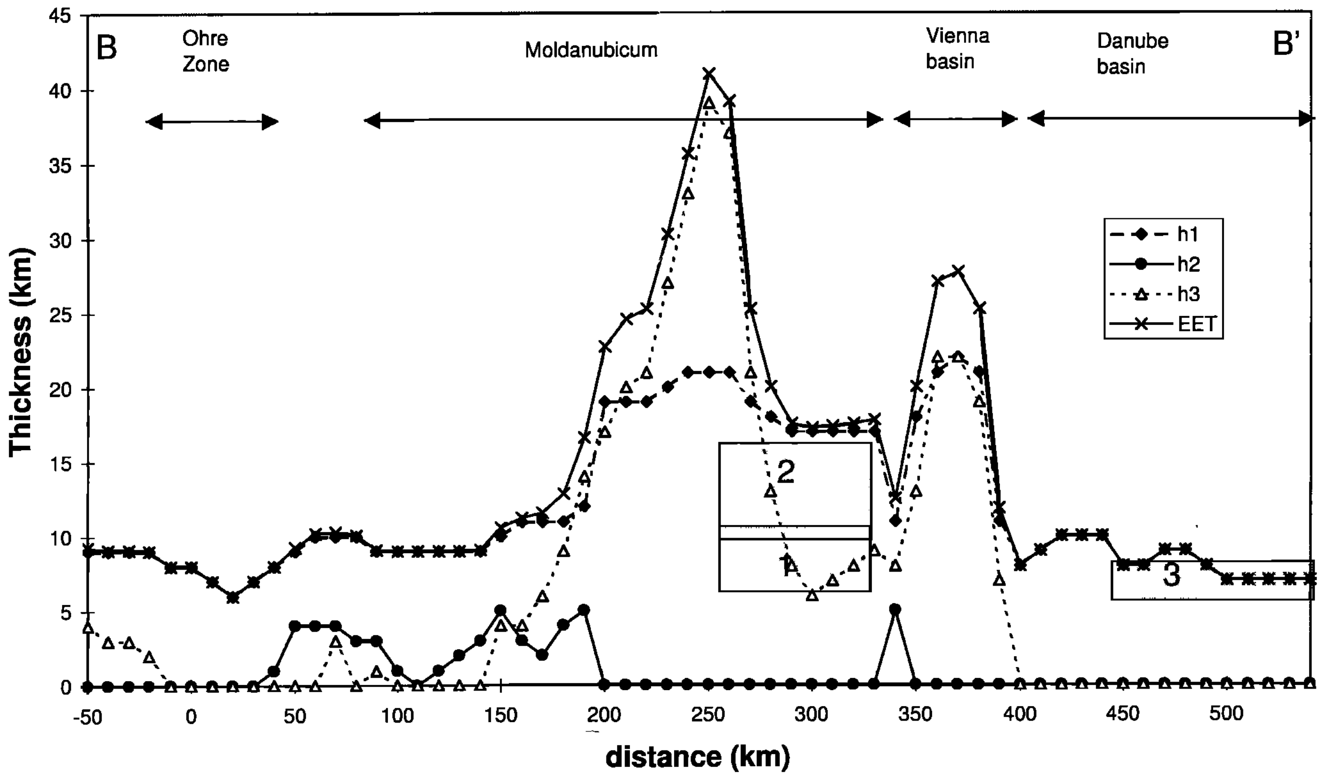

Figure 8b. EET ditribution along profile B-B'. EET is mainly governed by the uppermost part of the crust (h_1). In the Bohemian massif, however, the EET is controlled by the mantle rheology (h_3), although the upper crust also adds considerably to the EET. The strong layers of the upper and lower crust underneath the Bohemian massif are attached and have a combined thickness (h_1) of ~ 20 km, leaving the thickness of h_2 as zero. Boxes 1, 2, and 3 indicate independent EET estimates based on foreland basin models of 5 – 10 km [*Zoetemeijer et al.,* 1999] and 8 – 16 km [*Krzywiec and Jochym,* 1997] and extensional basin models yielding an EET of 7 km [*Lankreijer et al.,* 1995; *van Balen et al.,* 1999], respectively.

(Karpatian and early Badenian). This yields a strain rate of 9.5×10^{-15} s^{-1}.

Palinspastic reconstructions of the late Oligocene - early Badenian deformation in the Carpathian thrust belt yield shortening values of between 130 km (original length of 190 km) and 180 km (original length of 230 km) [*Ellouz and Roca,* 1994; *Roure,* 1994]. This produces strain rates of 3.6×10^{-15} s^{-1} and 4.1×10^{-15} s^{-1}. Strainrates for Magura and Silesian nappe deformation amount to 10^{-15} s^{-1} - 10^{-16} s^{-1} [*Nemčok et al.,* 1997].

Geodynamic reconstructions provide similar amounts of displacement. *Csontos et al.* [1992] shows an estimate of 150 km displacement for the Carpathian front during mid-Miocene and younger times. The deformed area includes in Csontos et al.'s model the extensional basin areas in the Pannonian basin system, ~ 400 - 600 km . The time involved in this displacement is difficult to quantify and is dependent on the location in the Carpathian arc due to the migration of thrusting along the arc. Estimates are of the order of 3 - 6 Myr, yielding strain rates roughly between 1.3×10^{-15} s^{-1} and 3.9×10^{-15} s^{-1}. Short-term strain rates, based on seismicity of the Vrancea area, yield estimates of 1.1×10^{-14} s^{-1} [*Oncescu and Bonjer,* 1997].

Since most observed deformation related to the formation of the Neogene Pannonian-Carpathian system is of the order of between 20 and 60% and the time involved is of the order of a few million year, strain rates typically are of the order of 10^{-15} s^{-1} to 10^{-14} s^{-1}. In summary, differences in strain rate of one order of magnitude induce differences in strength predictions of no more than 10%, which is well within the uncertainties introduced by the thermal model and the gravity model.

5.2. Tectonic Implications

The Polish Platform is characterized by significant NW-SE trending shear zones parallel to the Tornquist - Teisseyre zone (e.g., Odra Fracture zone, Elbe Fracture zone, Main Intra-Sudetic fault, and Lusatian main fault), possibly of Cadomian or older origin [*Żelaźniewicz and Bankwitz,* 1995]. These zones have been reactivated by subsequent various stress regimes.

During Variscan times, large-scale dextral strike-slip movements (up to 300 km [*Aleksandrowski,* 1995]) along these faults occurred, accommodating the northward Variscan compression [*Franke et al.,* 1993]. Also, the dextral Intra-Sudetic strike-slip basin opened along the Intra-Sudetic NW-SE trending strike-slip fault. The Polish Trough opened

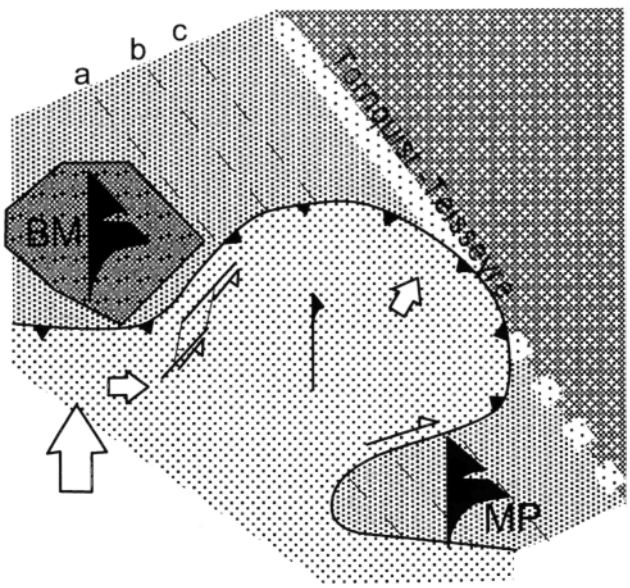

Figure 9a. Tectonic sketch of the effect of the rigid Bohemian massif (BM) and Moesian Platform (MP) on the north-south compression in the Alpine-Carpathian transition zone.

parallel to the Tornquist-Teiseyre zone in Mesozoic times [*Ziegler,* 1990].
Alpine inversion movements inverted the Polish Trough and the northern Bohemian margin, including the Intra-Sudetic basin, reactivating these NW-SE trending faults. Carpathian foreland flexure is largely governed by these NW-SE trending weakness zones [*Krzywiec,* 1997]. Large lateral displacements along the Odra fault zone [*Mastalerz and Wojewoda,* 1990] and the Diendorf Fault [*Leichman and Hejl,* 1996] occurred during Quaternary. In Tertiary times, the western border of the Bohemian massif (Franconian lineament) was heavily inverted [*Coyle et al.,* 1997].
In the Moesian Platform (Romania), similar NW-SE trending shear zones, influencing to a large extent the flexure of the foreland, have been documented [*Maţenco,* 1997; *Sandulescu and Visarion,* 1978; *Visarion and Sandulescu,* 1979]. We can only speculate on the extent of this fault system, since the basement areas do not outcrop.

Ziegler's [1990] map of Permian paleogeography indicates that it is likely that this fault system coincides with the northern boundary of the London-Brabant massif and that the entire area northward of the Alpine - Carpathian orogenic front and south of the Tornquist-Teisseyre line, including the west Netherlands basin, is influenced by this fault system. There could be a causal relation between the dominant NW-SE direction of the recent stress field [*Müller et al.,* 1992; *Zoback et al.,* 1993b] and the main direction of large-scale basement shear zones in this area still governing the recent rheology.

These shear zones are not manifested as discrete faults, but in the case of the Polish Platform, it is better to refer to an anisotropic fabric with two different axes. This implies that because of this megascale foliation, material properties like grain size and thus rheology are different for shear zone parallel and shear zone perpendicular directions. In shear zone perpendicular directions (NE-SW), extensional stresses will be able to reactivate the inherited shear fabric as normal faults, thus utilizing the minimal yield strength. NW-SE oriented extensional stresses will have to overcome the maximum strengths.

The rheology we predict for the Polish foreland does not take into account the above described NW-SE trending rheology anisotropy but only describes the maximum strengths. Since it is difficult to assess the actual failure mechanisms in the NW-SE shear zones with depth, the calculation of failure envelopes will not provide a satisfactory minimum yield-strength envelope.

Studies of extensional basins, like the Polish Trough, can provide this minimum rheology. The anisotropy of the rheology of the Polish Platform will have important consequences for the flexural behavior of the foreland downbending underneath the arc-shaped Carpathians, thus loading in different directions with respect to the weakest direction.

The strength maximum calculated in the foreland area of the Carpathians (Figures 4 and 7) places important constraints on the evolution of foreland basins. Downbending of the lower plate implies the introduction of relative cool material at greater depths, thus increasing the strength. This mechanism puts a limit to the rate of downbending. Bending rates in excess of the thermal relaxation rate will lead to an increase of strengths in the plate automatically blocking the movement by the increased flexural rigidity. The vertical loads associated with flexural basins induce fiber stresses that reduce the strength of a bending lithosphere severely [*Bertotti et al.,* 1997; *Cloetingh and Burov,* 1996]. If we can extrapolate our rheology predictions to the geologic past, they may shed some light on tectonic models for the area during Neogene times.

A striking feature in our predictions of lithospheric strength is the extremely strong Bohemian core, rooting deep into the lower lithosphere. It is likely that its core acted as a rigid anchor, blocking the northward movement of the colliding Alpine region, causing for example, large-scale sinistral strike-slip movements in the eastern Alps (Salzachtal – Ennstal fault and Mur – Mürz fault zone) opening the Vienna basin in Karpatian times (Figure 9a).

The proposed rheologic anisotropy in the Polish Platform and the Moesian Platform has probably determined the precollisional continental margin in the western and southern Carpathians. The NW-SE trending weakness zones favor a jagged edge (Figure 9b) to a straight or slightly curved edge. How such a margin with internal weakness zones reacts to loading is unclear. Each slab, separated by major shear zones, probably flexes individually as a reaction to the load. This creates a distinct shape of reactivated structures perpendicular to the axis of the foreland basin, like that described by *Krzywiec* [1997]. In the eastern Carpathians, the weakness zones run parallel to the margin, favoring a staircase geometry. Furthermore, differences in foreland basin development [*Zoetemeijer et al.,* 1999] between the western Carpathians, where the foreland basin axis is at high angles to the trend of the anisotropy, and the eastern Carpathians, where the foreland basin axis runs parallel to the anisotropy, can be due to this effect.

The predicted weakness of the lithosphere underlying the Pannonian basin makes it highly unlikely that it can

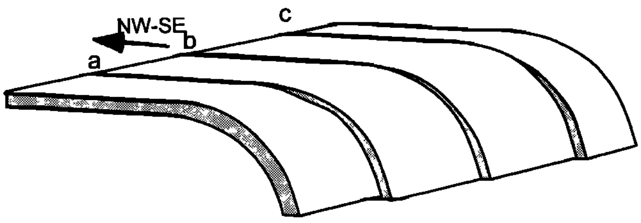

Figure 9b. Tectonic sketch of the effect of rheologic anisotropy on the precollisional margin of the Polish platform and the Moesian Platform. NW-SE trending, inherited weakness zones cleave the slab into separated segments, causing differential flexing of the separate parts along the western and southern Carpathian margin, where the rheologic anisotropy is at high angle to the margin. Along the eastern Carpathians margin, the anisotropy is parallel to the margin.

compensate the load of the shallow asthenosphere, which produces a gravity effect of +50 mGal at least [*Bielik et al.*, 1994]. Therefore a passive subsidence mechanism, caused by re-equilibration of the thinned lithosphere, as described by *Huismans et al.* [*1999*] is likely to occur in the Pannonian basin. The two-phase subsidence history of the Neogene Pannonian basin system [*Lankreijer et al.*, 1999] and the spatial correlation of the youngest extension phase with the weakest lithosphere [*Lankreijer*, 1998] indicates a causal relationship between the asthenospheric dome, the extension processes, and the weak rheology. The observed good fit between our strength predictions and those inferred from the approaches of others (Figure 8a and 8b) provides an independent validation of our results.

6. Conclusions

We predict a detached behavior of the crust and mantle for all study areas for the adopted strain rate. However, a faster strain rate will cause coupling of the strong lithosphere in the Bohemian massif. We speculate that the stiff Bohemian massif causes major implications for the eastern Alpine – Carpathian tectonic evolution. A similar behavior was predicted for the Moesian Platform [*Lankreijer et al.*, 1997]. Furthermore, the strong rheologic contrast between the Pannonian area and the surrounding platform areas supports scenarios in which the shape of the Pannonian embayment was predetermined by the passive margins of the lithosphere surrounding the present-day Carpathian arc. We also predict a strong anisotropy in the rheology of the Polish Platform, possibly extending to the North Sea area, directly linked to the occurrence of NW-SE basement faults, parallel to the Tornquist-Teisseyre zone.

Large lateral variations in present-day lithospheric rheology are predicted for the study area, corresponding to the broad spectrum of thermotectonic ages encountered. The Bohemian massif forms a relatively strong lithospheric block. The rigid behavior is responsible for complex large-scale strike-slip movements along these blocks in order to accommodate the emplacement of the internal microplates of the Pannonian system.

The Polish platform is characterized by a rheologic anisotropy induced by large-scale NW-SE trending shear zones that form prominent weakness zones controlling reactivation since Variscan times. This anisotropy controls the shape of the Carpathian foreland basin, the tectonic history of the Polish Trough in Mesozoic times, and the Intra-Sudetic basin in Paleozoic times. The Pannonian basin system is dominated by a weak rheology, owing to high lithospheric temperatures. In general, the peripheral basins of the Pannonian basin show a relatively stronger present-day rheology than the central basins do.

The inferred large variations in lithospheric strength suggest that tectonic models should be based on units with similar rheology (i.e., the strong part of the Bohemian massif or the Polish Platform - Moesian Platform rheologic anisotropy), rather than primarily based on geographical units. Additionally, temporal changes of rheology should be taken into account in such models.

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