J Geophys (1981) 49:212–216

Journal of Geophysics

A Comparison of the Thermal and Mechanical Structure of the Lithosphere Beneath the Bohemian Massif and the Pannonian Basin

K.M. Onuoha

Institut für Geophysik, TU Clausthal, Postfach 230, D-3392 Clausthal-Zellerfeld, Federal Republic of Germany*

Abstract. Temperature-depth profiles for the Bohemian Massif and the Pannonian Basin have been determined on the basis of: the known surface heat flow values in both regions, a vertical crustal radiogenic heat production that is estimated from seismic velocity (v_P) profiles, and a thermal conductivity that varies with depth. From these profiles, effective viscosity and stress-depth curves have been determined to a depth of 250 km assuming both linear and non-linear mantle rheology. There is a relative dominance of Newtonian Nabarro-Herring creep in the asthenosphere beneath the Pannonian Basin. Stress values are also significantly lower here than in the lithosphere beneath the Bohemian Massif. The transition from lithosphere to asthenosphere appears to occur at depths of 48–60 km under the Pannonian Basin. The absence of a viscosity minimum under the Bohemian Massif renders an estimate of the depth of this transition difficult, but stress curves appear to level off around depths of 180–200 km, and may thus provide estimates of the depth to the asthenosphere under the Bohemian Massif. Models of the lithosphere beneath these two tectonically different regions are also presented.

Key words: Stress – Effective viscosity – Brittle lithosphere – Ductile lithosphere – Temperature regime

Introduction

The Bohemian Massif of Czechoslovakia is a platform-type cratonic block, consolidated during the Variscan orogeny. Approximately rhombic in shape and bordered by deep SW-NE, NW-SE trending faults, the western margin of the Massif is in the Federal Republic of Germany while the southern margin is in Austria. It is bordered towards the north and north-east by the North-German-Polish lowland. The core of the Bohemian Massif is the Moldanubicum, a metamorphosed Precambrian unit interspersed with Hercynian granites (Máska et al. 1960). The Pannonian Basin on the other hand is a Late Miocene-Pliocene feature whose formation is believed to have been genetically connected with the Late Oligocene-Early Miocene subduction in the Carpathian realm (Stegena et al. 1975; Horváth et al. 1977).

Both the Pannonian Basin and the Bohemian Massif have been relatively well studied geologically and geophysically. In this paper the temperature regimes of the lithosphere in both regions are compared, and the effects of the differences on such rheological parameters as effective viscosity and stress are discussed. The depths to the zone of transition between brittle and creep dominated behaviour in the lithosphere beneath both regions are examined, and their bearing on seismicity and faulting is discussed. The Moldanubicum which forms the core and the best consolidated part of the Bohemian Massif is the actual area referred to in the discussions because the temperature-depth distribution for this unit can be regarded as characteristic for most of the Bohemian Massif (Cermák 1975). The temperature-depth distribution for the Karcag-Debrecen region of Eastern Hungary is taken as representative for most of the Pannonian Basin. These specific locations have been chosen for the determination of the profiles because in each area reliable data exist for heat flow and crustal structure. Moreover the calculated temperature, stress and viscosity profiles help to illustrate the differences in lithospheric structure between two characteristically different regions of the earth's crust: the Pannonian Basin representing a young intermontane basin and the Bohemian Massif representing a consolidated Variscan platform.

Temperature-Depth Profiles

Results of heat flow and geothermal studies in the Bohemian Massif and the surrounding regions have been reported by Cermák (1975, 1977). Important results of heat flow studies in the Pannonian Basin were reported by Stegena (1976), Stegena et al. (1975) and Horváth et al. (1979).

Temperature-depth profiles have been calculated for both regions using a computational model of steady-state conduction of heat in a layered medium. The surface heat flow Q is assumed to be in equilibrium with the heat flowing into a slab of the earth at its base, plus the heat generated by sources within it. Neglecting the earth's curvature and assuming that the temperature T depends only on depth z, we have the one-dimensional heat conduction equation:

$$\frac{d}{dz}\left[k(z)\frac{dT}{dz}\right] = -A(z) \tag{1}$$

where k is the coefficient of thermal conductivity, and A is the heat generation.

If $T(z_0)$ is the temperature at the top of the first layer (with conductivity $k_1=$const. and heat generation $A_1=$const.) and $Q(z_0)$ the heat flow through the upper boundary of the first layer, the temperature in the first layer is

* *Present address:* Department of Geology, University of Nigeria, Nsukka, Nigeria

0340-062X/81/0049/0212/$01.00

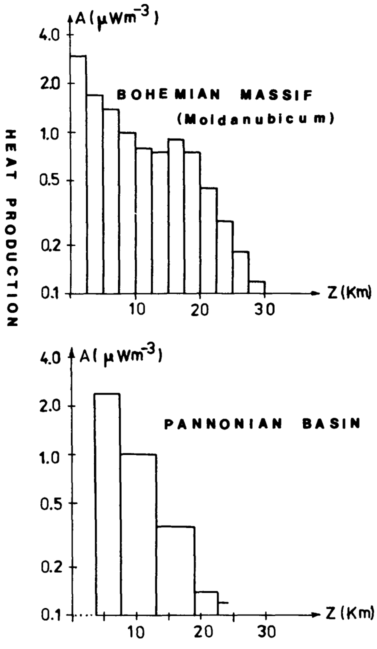

Fig. 1. Heat production-depth relationships for the Moldanubicum (Bohemian Massif) and the Pannonian Basin, based on v_P-depth profiles of Beránek (1971), Mituch and Posgay (1972), Ádám et al. (1979)

$$T(z)=T(z_0)+\frac{Q(z_0)}{k_1}\,z-\frac{A_1}{2k_1}z^2. \tag{2}$$

$(z_0 \leqq z \leqq z_1)$

Pollack (1965) has shown that for any arbitrarily layered model (with homogeneous individual layers) the temperature at the base of the n-th layer is:

$$T(z)=T(z_0)+\sum_{i=1}^{n}\left\{\frac{z_i-z_{i-1}}{k_i}\left[Q(z_0-\sum_{j=1}^{i-1}A_j(z_j-z_{j-1})\right]-A_i\frac{(z_i-z_{i-1})^2}{2k_i}\right\}. \tag{3}$$

In the present paper the vertical profile of crustal heat generation A has been determined from the seismic velocity (v_P) profiles, making use of the implicit relationship between A and v_P as described by Rybach (1973) and Buntebarth (1976). Following closely the same assumptions as those of Pollack and Chapman (1977) the value of k was taken as $2.5\,\mathrm{Wm^{-1}K^{-1}}$ for $T \leq 500\,°\mathrm{C}$, while for higher temperatures, the values of k were obtained using the results of Schatz and Simmons (1972). The heat generation-depth relations are shown in Fig. 1, while the temperature-depth profiles are displayed in Fig. 2. The sedimentary cover in the Karcag-Debrecen area of the Pannonian Basin is between 2 and 3.5 km thick in many places. The $A-v_P$ relation is not strictly valid for such sedimentary series, and this explains why the $A-z$ relation for the Pannonian Basin is given only from the depth $z=4$ km in Fig. 1. The heat production in this uppermost layer has no appreciable effect on the overall shape of profile 2 in Fig. 2. It should also be noted that, in this figure, profiles 1 and 2 are mean curves of a family of profiles corresponding to the range of surface heat flow data available for each of the two regions under consideration.

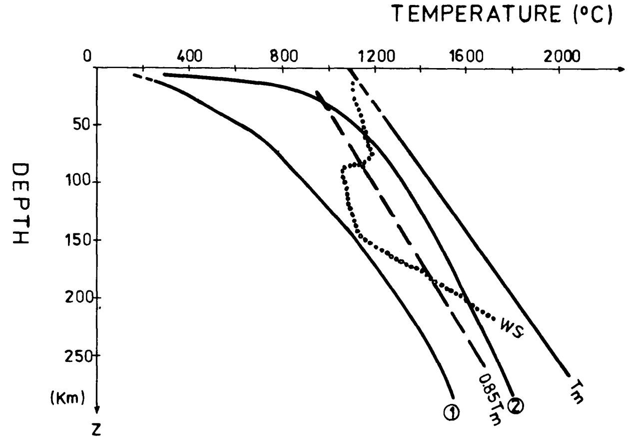

Fig. 2. Temperature-depth profiles: 1 = Bohemian Massif; 2 = Pannonian Basin; T_m = dry pyrolite sodius; WS = solidus for pyrolite containing 0.1 % water [T_m and WS are after Ringwood (1975)]

Continuation of the geotherms further downwards into the asthenosphere would necessitate a departure from the purely conductive model of heat transfer which is assumed in the calculations. It should be noted that the solidus temperature in an anhydrous mantle would increase with depth. The actual level at which the solidus is reached in the mantle remains debatable, but it is now agreed that this depends on the role of volatiles in depressing the solidus temperature. The Pannonian geotherm intersects the solidus for pyrolite containing 0.1% water at a depth of less than 60 km. It is thus likely that a partial melt zone should characterise the upper mantle in this region at this shallow depth. Beneath the Bohemian Massif such an intercept does not occur within the depth range investigated.

Stress-Depth and Viscosity-Depth Profiles

Following Weertman and Weertman (1975) or Meissner and Vetter (1976), we take the following general creep equation to be valid for the creep properties of rocks at high temperatures:

$$\dot{\varepsilon}=C_i\sigma^n\exp\left[-(Q^*+PV^*)/RT\right] \tag{4}$$

$$\dot{\varepsilon}\approx C_i\sigma^n\exp(-g^*T_m/T) \tag{5}$$

where

g^*	= constant	P	= pressure
$\dot{\varepsilon}$	= strain rate	V^*	= activation volume
C_i	= quasi constant	R	= gas constant
σ	= shear stress	T	= absolute temperature
n	= creep exponent	T_m	= absolute melting temperature
Q^*	= activation energy		

For Nabarro-Herring or diffusion (Newtonian) creep $n=1$, and

$$\dot{\varepsilon}_{\mathrm{NH}}=C_1\,\sigma\exp(-g^*T_m/T) \tag{6}$$

with

$$C_1=\alpha_1\Omega D_0/d^2kT \tag{7}$$

and

α_1 = constant
k = Boltzmann's constant
D_0 = diffusion constant.
d = grain size
Ω = atomic volume

For dislocation glide (or power law) creep $n=3$ and

$$\dot{\varepsilon}_{PL} = C_3 \sigma^3 \exp(-g^* T_m/T) \tag{8}$$

with

$$C_3 = \alpha_3 \Omega D_0 / \mu^2 kT$$

and

α_3 = constant,
μ = shear modulus.

Introducing the effective viscosity η

$$\eta \approx \sigma / \dot{\varepsilon} \tag{9}$$

From these equations the viscosities for both creep laws can be expressed explicitly:

$$\eta_{NH} = (1/C_1) \exp(g^* T_m/T) \tag{10}$$
$$\eta_{PL} = (1/C_3)(1/\sigma^2) \exp(g^* T_m/T). \tag{11}$$

These relations have been discussed in detail by Meissner and Vetter (1976).

The stress-depth curves (Figs. 3 and 4) and the Newtonian creep viscosity profiles (Fig. 5) have been calculated on the basis of Eqs. (5)–(9) assuming a grain size of 5 mm for the material of mantle rock, and with T/T_m values obtained from Fig. 2. The dry solidus was used for the calculations, not the hypothetical 0.1 % H_2O curve. The values of the constants C_1, C_3 and g^* were taken from Vetter and Meissner (1977, 1979). The non-linear stress-strain relation with $n=3$ has been shown to be more appropriate for much of the lithosphere (Neugebauer and Breitmayer 1975; Mercier et al. 1977).

Discussion

It is obvious from the above figures that the upper mantle regions beneath the Pannonian Basin and the Bohemian Massif differ significantly from each other not only in their temperature regimes, but probably also in composition, at least within the depth range considered here. The temperature-depth curves for the two regions do not intersect even at depths of about 250 km. There is however no information on whether these differences might exist up to and possibly beyond a depth of 400 km. In the sense of Oxburgh and Turcotte (1976) the lithosphere can be divided into an upper elastic layer characterised by brittle response to stresses, and a lower ductile or plastic layer where elastic stresses are relaxed by creep processes.

The temperature at which the transition from elastic to creep-dominated behaviour occurs in the lithosphere depends on the rheological properties of the material of the lithosphere. These are in turn dependent on the homologous temperature (T/T_m), and on the variation of T_m with depth. A value of 0.62 for T/T_m has been assumed in the present paper to define this transition. Other authors (e.g., Vetter and Meissner 1979) have also assumed T/T_m values close to this.

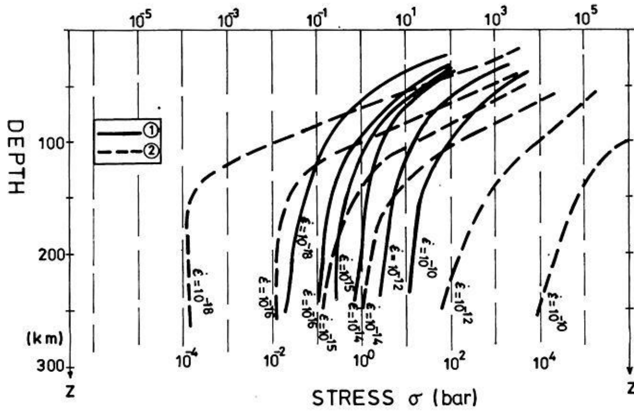

Fig. 3. Stress-depth curves for the Bohemian Massif: *1* = dislocation (power law creep), *2* = Newtonian diffusion creep, with strain rate values (in s^{-1}) as parameter

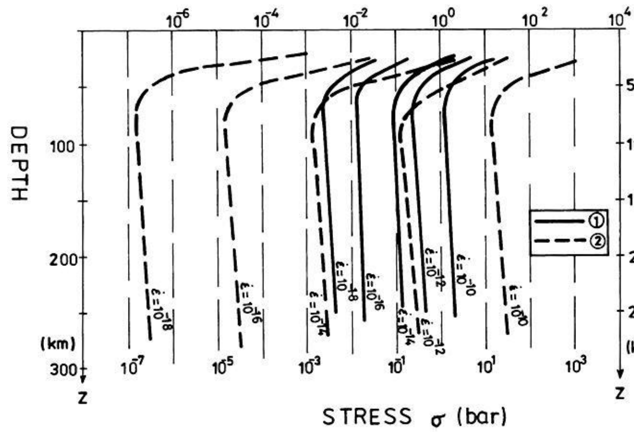

Fig. 4. Stress-depth curves for the Pannonian Basin: key as for Fig. 3

Under the Pannonian Basin the elastic lithosphere according to the above definition is only 10–12 km thick while under the Bohemian Massif its thickness lies between 58 and 65 km (see Fig. 5). The occurrence of exceptionally deep faults and grabens in the Bohemian Massif and their conspicuous absence within the Pannonian Basin can now be understood considering the fact that the maximum depth of faults in a region seems to depend on the thickness of the brittle layer. The Central Bohemian suture, the Pribyslav fault zone and the Ohře graben are examples of deep faults of great length found in the Bohemian Massif. Their extents, laterally and in depth, have been investigated by Beránek and Dudek (1972).

The deformation diagrams (Figs. 3 and 4) show that beneath the Bohemian Massif, power-law creep dominates for strain rates higher than $10^{-15}\,s^{-1}$. For lower strain rates (and at depths greater than 185 km for $\dot{\varepsilon}=10^{-15}\,s^{-1}$) Newtonian creep appears to dominate. The transition between the two creep laws is at about 0.6 bar. This corresponds to the stress level at which the two mechanisms contribute equally to the strain rate. It is however known that the actual value of this transition depends mainly on the assumed grain size in Eqs. (7) and (10) (Stocker and Ashby 1973).

Beneath the Pannonian Basin stresses are very much lower because of the higher temperature regime. Consequently diffusion (Nabarro-Herring) creep dominates even for strain rates as high as $10^{-12}\,s^{-1}$. For $\dot{\varepsilon} < 10^{-12}\,s^{-1}$ and for depths greater than

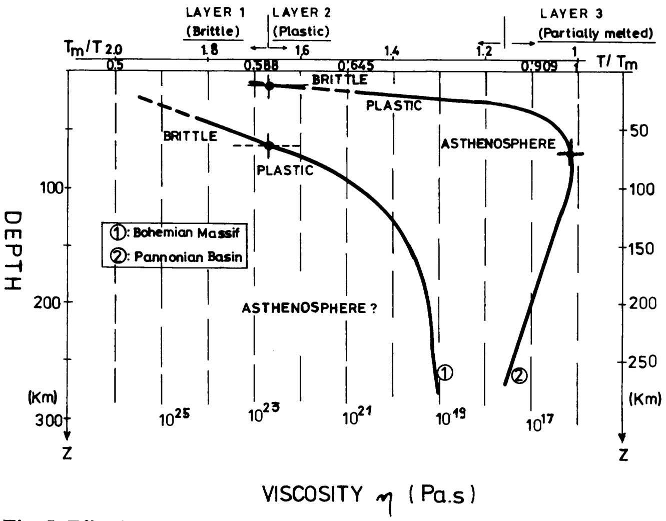

Fig. 5. Effective viscosity-depth profiles

60 km diffusion creep governs all creep processes in the Pannonian upper mantle. Comparing these results with those of Vetter and Meissner (1977), the Pannonian upper mantle can be seen to show mechanical properties reminiscent of the oceanic upper mantle. The transition from lithosphere to asthenosphere appears to occur at depths of 48-60 km under the Pannonian Basin. These results are in good agreement with geoelectric induction and magnetotelluric results which have located the intermediate conductive layer (ICL) at depths of 60 km below the Pannonian Basin (Ádám 1976). Figure 4 also shows that stresses under the Pannonian Basin appear to level off for all curves with constant strain rate at the depth of about 50—60 km. The depth at which this levelling off of stress occurs is usually identified as marking the top of the asthenosphere (Mercier et al. 1977).

Very often there have been speculations about the possible absence of the asthenosphere under shields and stable platforms. In the case of the Bohemian Massif where the absence of a viscosity minimum (see Fig. 5) renders an estimate of the depth of transition from lithosphere to asthenosphere difficult, this view might appear to be substantiated. However, if there is no viscous layer undergoing shear in the upper mantle under stable platforms and shields, it is then difficult to imagine how such parts of the lithospheric plate could take part in the world-wide plate motions. Deep geoelectric soundings (Pecová et al. 1976) located some conductive layers at shallower levels of 60–70 and 100 km, but it appears likely that the conductive zone located near 200 km under the Bohemian Massif might correspond to the base of the lithosphere. The stress curves (Fig. 3) show signs of levelling off only at depths of about 170–200 km. It is thus likely that this zone may really mark the top of the asthenosphere beneath the Bohemian Massif. Figure 6 shows the preferred models of the Pannonian and Bohemian lithospheres.

In the Pannonian Basin dispersed earthquakes occur with shallow focal depths (5–14 km). With the exception of the Dunaharaszti earthquake (12 January 1956; $M = 5.5$-6.0) which was apparently related to sinistral faulting along the Balaton line (Csomor 1967), the seismic activity in the Pannonian Basin shows no correlation with tectonic lines. It is thus likely that this activity is due to the relaxation of thermoelastic stresses in the brittle upper part of the Pannonian lithosphere which is only 10–12 km thick. The Bohemian Massif shows very little seismic activity except along the deeply faulted margins. However, it has long been known that seismic energy originating from earthquakes in the Eastern and Southern Alps was transmitted with little or no absorption through the Bohemian Massif, while seismic energy propagating towards the Western Carpathians and the Pannonian Basin became strongly damped (Kárnik 1967). Zátopek (1957) sought to explain this damping by suggesting the existence, in the contact zone between the two tectonic regions, of deep-seated barriers which reflect the energy further downwards. The difference in the observed teleseismic delay-time residuals between the two regions under consideration is of the order of $+2s$. The stations in the Pannonian Basin are far "slower" (i.e., they record far higher positive residuals), than those in the Bohemian Massif. It is easy to see using simple model calculations that the differences in the range of P_n anomalies or teleseismic delay-time residuals cannot be explained in terms of differences in crustal structure alone, between the Bohemian Massif and the Pannonian Basin. The differences rather reflect the variations in the physical state of the upper mantle under these regions. This in turn means that differences in the sub-Moho temperature regime and rheology are the obvious causes of the variations in P_n anomalies.

Gravity anomalies in the Bohemian Massif seem to be very

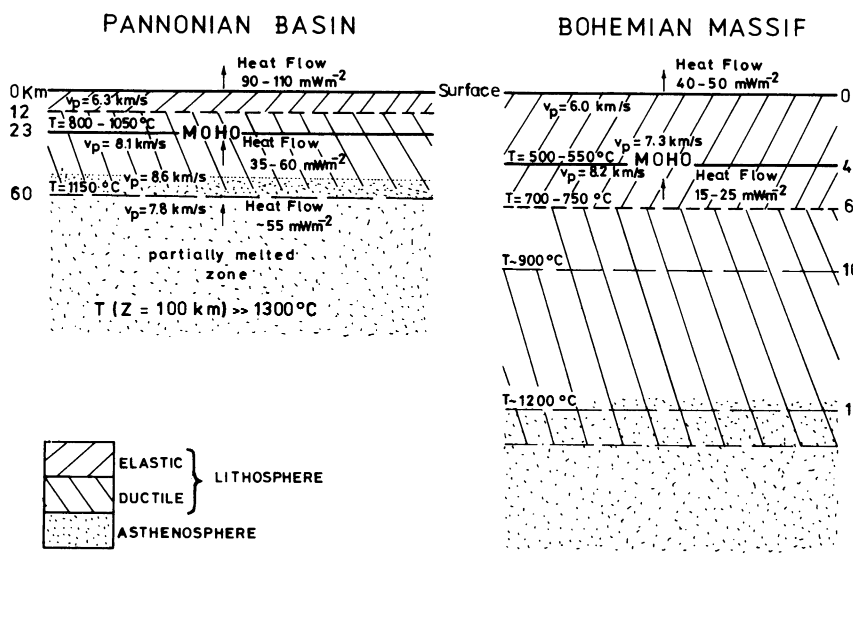

Fig. 6. Preferred lithospheric models: surface heat flow data from Cermák (1975), Horváth et al. (1979); v_P values from Benárek (1971), Ádám et al. (1979), temperature values from this study

well accounted for by granitic intrusions and other near-surface density inhomogenities (Tomek, personal communication). The Pannonian Basin on the other hand is characterized by average Bouguer anomalies between +10 and +15 mgal. Taking into consideration the anomalously thin crust and the elevation of the high-density upper mantle material, these anomalies are strikingly low. Deep seismic soundings have revealed that beneath the Pannonian *M*-discontinuity, *P*-wave velocities are in the range 8.0–8.1 km/s (see Fig. 6). These velocity values indicate the range of probable density values under the Moho. It then becomes apparent that a discrepancy exists in the observed gravity anomalies. The mass which compensates the missing positive anomalies probably exists deeper down in the asthenosphere and is likely to be of thermal origin.

Concluding Remarks

It should again be pointed out that the profiles calculated in this study are models and refer to dry conditions in the lithosphere. The presence of water-bearing minerals (e.g., phlogopite and amphibole) in some xenoliths from the upper mantle beneath many continental regions does indicate that the upper mantle probably contains reasonable quantities of H_2O. The depth of the lithosphere/asthenosphere transition estimated from deformation diagrams (Figs. 3 and 4) or viscosity profiles (Fig. 5) may not necessarily agree with that obtained from purely thermal models (Pollack and Chapman 1977) or with the depth to the low-velocity zone (LVZ). The strong differences in the stress-depth and viscosity-depth profiles in both regions are enough to explain the seismological, teleseismic and other geological differences such as the occurrence and depth of faults. These differences between the two tectonic regions are all thermally controlled and remain valid even if other models are assumed for the melting conditions in the mantle.

Acknowledgement. The author would like to thank Dr. F. Horváth and Professor L. Stegena (both of Eötvös University, Budapest) for instructive discussions. Dr. G. Buntebarth and Professor H.J. Neugebauer suggested improvements to the manuscript. This work was carried out at the Techn. Univ. Clausthal during the tenure of a fellowship from Deutscher Akademischer Austauschdienst (DAAD).

References

Adám, A.: Results of deep electromagnetic investigations in the Pannonian Basin. In: Geoelectric and geothermal studies, A. Adám, ed.: pp. 547–561. Budapest: Akadémiai Kiadó 1976

Adám, A., Bisztricsány, E., Wallner, A., Horváth, F., Meskó, A., Stegena, L., Tarcsai, Gy., Posgay, K.: Hungarian IASPEI report for the general assembly of IUGG, Canberra, Australia, 1979

Beránek, B.: Study of velocity conditions in the earth's crust in the regions of the Bohemian Massif and the Carpathian system. Stud. Geophs Geod. **15**, 316–330, 1971

Beránek, B., Dudek, A.: The results of deep seismic sounding in Czechoslovakia. J. Geophys. **38**, 415–427, 1972

Buntebarth, G.: Methoden zur Abschätzung der Wärmeflußdichte aus dem Oberen Mantel. Geol. Rundsch. **65**, 809–919, 1976

Cermák, V.: Temperature-depth profiles in Czechoslovakia and some adjacent areas derived from heat flow measurements, deep seismic sounding and other geophysical data. Tectonophysics **26**, 103–119, 1975

Cermák, V.: Heat flow map of the Bohemian Massif. J. Geophys. **42**, 455–458, 1977

Csomor, D.: Focal mechanism solution of the Hungarian earthquake 12:1:1956 (in Russian). Ann. Univ. Sci. Budap., Rolando Eotvos Nominatae Sect. Geol. **10**, 3–8, 1967

Horváth, F., Vörös, A., Onuoha, K.M.: Plate tectonics of the western Carpatho-Pannonian region: a working hypothesis. Acta Geol. Acad. Sci. Hung. **21**, 207–221, 1977

Horváth, F., Bodri, L., Ottlik, P.: Geothermics of Hungary and the tectonophysics of the Pannonian Basin "red spot" In: Terrestrial heat flow in Europe, V. Cermák, L. Rybach, eds.: pp. 206–217. Berlin, Heidelberg, New York: Springer 1979

Kárnik, V.: Crustal and upper mantle seismic activity in the European Area. Stud. Geophs. Geod. **11**, 324–334, 1967

Máska, M., Matejka, A., Zoubek, V.: Tectonic development of Czechoslovakia (Collected papers and the tectonic map 1:1,000,000), pp. 224. Praha: Nakl. CSAV 1960

Meissner, R.O., Vetter, U.R.: Isostatic and dynamic processes and their relation to viscosity. Tectonophysics **35**, 137–148, 1976

Mercier, J.C., Anderson, D.A., Carter, N.L.: Stress in the lithosphere: Inferences from the steady state flow of rocks. Pure Appl. Geophys. **115**, 199–226, 1977

Mituch, E., Posgay, K.: Results of DSS measurements in Hungary. In: The crustal structure of Central and Southeastern Europe based on the results of explosion seismology. Geophysical Transactions, Gy. Szénás, ed.: pp. 118–129. Budapest: ELGI 1972

Neugebauer, H.-J., Breitmayer, G.: Dominant creep mechanism and the descending lithosphere. Geophys. J. R. Abstron. Soc. **43**, 873–895. 1975

Oxburgh, E.R., Turcotte, D.L.: The physico-chemical behaviour of the descending lithosphere. Tectonophysics **32**, 107–128, 1976

Pecová, J., Petr, V., Praus, O.: Depth distribution of the electric conductivity in Czechoslovakia from electromagnetic studies. In: A. Adám (editor), Geoelectric and geothermal studies, A. Adám, ed.: pp. 517–537. Budapest: Akadémiai Kiadó 1976

Pollack, H.N.: Steady heat conduction in layered mediums: the half-space and sphere. J. Geophys. Res. **70**, 5645–5648, 1965

Pollack, H.N., Chapman, D.S.: On the regional variation of heat flow, geotherms, and lithospheric thickness. Tectonophysics **38**, 279–296, 1977

Ringwood, A.E.: Composition and petrology of the earth's mantle. Int. Ser. Earth Planet. Sci. New York: McGraw-Hill, 1975

Rybach, L.: Wärmeproduktionsbestimmungen an Gesteinen der Schweizer Alpen. Beitr. Geol. Schweiz., Geotech. Ser., Lief. **51**, Bern, 1973

Schatz, J.F., Simmons, G.: Thermal conductivity of earth materials at high temperatures. J. Geophys. Res. **77**, 6966–6983, 1972

Stegena, L.: The variation of temperature with depth in the Pannonian Basin. In: Geoelectric and geothermal studies, A. Adám, ed.: pp. 425–438. Budapest: Akadémiai Kiadó 1976

Stegena, L., Géczy, B., Horváth, F.: Late Cenozoic evolution of the Pannonian Basin. Tectonophysics **26**, 71–90, 1975

Stocker, R.L., Ashby, M.F.: On the rheology of the upper mantle. Rev. Geophys. Space Phys. **11**, 391–426, 1973

Vetter, U.R., Meissner, R.O.: Creep in geodynamic processes. Tectonophysics **42**, 37–54, 1977

Vetter, U.R., Meissner, R.O.: Rheologic properties of the lithosphere and applications to passive continental margins. Tectonophysics **59**, 367–380, 1979

Weertman, H., Weertman, J.R.: High temperature creep of rock and mantle viscosity. Annu. Rev. Earth Planet. Sci. **3**, 293–315, 1975

Zátopek, A.: Zu einigen Problemen der Krustendynamik im Karpatengebiet. Geofiz. Kozl. **8**, 196–110, Budapest, 1957

Received July 4, 1980; Revised Version September 11, 1980
Accepted November 6, 1980