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Thermal lithosphere across the Trans-European Suture Zone in Poland

Jacek A. MAJOROWICZ

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Significant lateral variations of surface heat flow occur in the Polish Lowland area, ranging up to 30+/–10 mW/m^2 across the transition from the East European Craton (EEC) and the northeastern part of the Teisseyre-Tornquist Zone (TTZ) to the accreted terranes in the south-west (Palaeozoic Platform) and up to 25 mW/m^2 of change within the Trans-European Suture Zone (TESZ). Modelling of the crustal temperatures for the deep seismic profiles parallel to TESZ (P1, P5 and TTZ) and perpendicular to it (LT-7, P2, LT-2, P4, LT-4 and LT-5) shows evidence of extensive crustal-mantle warming (elevated mantle heat flow in the area between the Sudetes and the EEC). The EEC and the northern part of the TTZ have a much lower mantle heat contribution. Mantle heat flow variations are significant (approximately 20–40 mW/m^2). Significant are also variations in thermal lithosphere thickness ranging from *ca*. 150–200 km in the craton and the northern part of the TTZ to 100–150 km (locally less than 100 km) in the accreted terranes to the south-west of the TTZ and in the central part of the TTZ. The TTZ is a thermally inhomogeneous zone.The thermal transition between the Palaeozoic Platform and the EEC is not a sharp one. Significant variations in the thickness of the thermal lithosphere do not follow major tectonic units of the crust.

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INTRODUCTION

The first pilot numerical modelling of the thermal structure of crust in the Polish Lowland based on the seismic results of the POLONAISE'97 experiment used 3 profiles for which numerical data were available (LT-7, P1 and P3; Majorowicz *et al*., 2003). The high quality seismic data base includes other POLONAISE'97 profiles (P2, P4, P5, TTZ) and recently reinterpreted older profiles LT-2, LT-4, and LT-5 which can now be used for thermal modelling (see Fig. 1 for locations). Previous heat flow modelling (Majorowicz, 1978; Milanovsky, 1984; Čermak *et al*., 1989) used older generation deep seismic sounding (DSS) data.

The heat flow at the Moho and the temperature distribution in the crust and upper mantle were studied for the northern part of the TESZ (Balling, 1995). The mantle heat flow of 20–30 mW/m^2 was estimated for the northern and central Baltic Shield transects and Mazury area of the East European Craton (EEC) in northeastern Poland, increasing to 30–40 mW/m^2 beneath the shield margin, Danish and North-German deep basins and Tornquist Zone (Balling, 1995) and 30–45 mW/m^2 in the Variscan Platform in Poland (Majorowicz *et al*., 2003, POLONAISE'97 profiles LT-7 and P1). Mantle heat flow of 20–30 mW/m^2 was estimated by Förster and Förster (2000) for the Variscan Erzgebrige in Germany and is likely characteristic also for the whole Bohemian region.

The lithospheric temperatures were modelled along the two-dimensional distance-depth slices cutting through the transitional zone between the EEC, the TTZ and accreted terranes south-west of the TTZ (Fig. 1). The latest heat flow map of Poland (Fig. 2) constructed by Szewczyk and Gientka (2003) was used for the present modelling.

Several profiles parallel to the strike of the TTZ have been also analysed (seismic profiles TTZ and P5; Fig. 1). The results of this modelling were used in construction of maps of mantle heat flow contribution and distribution of thermal lithosphere thickness in the Polish Lowland, as defined by the 1300°C isotherm.

GEOLOGICAL AND GEOPHYSICAL BACKGROUND

The SW margin of the EEC, between the North Sea and the Carpathians–Black Sea area, termed the Tornquist Zone (TZ), consists of two segments. The first one, the Sorgenfrei-Tornquist

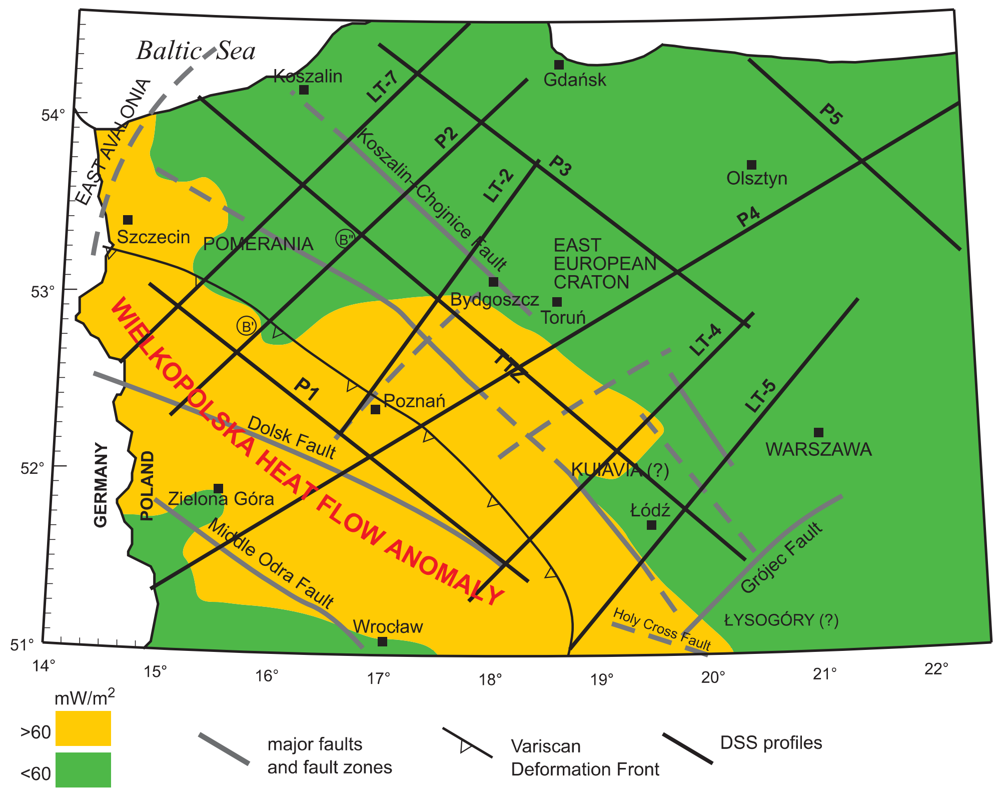

Fig. 1 Main structural elements of the Polish Lowland according to Dadlez (2000) and locations of the new generation seismic profiles shown against the heat flow map of Majorowicz (1984) modified by Majorowicz *et al*., (2002)

Zone, extends NW of Bornholm Island, and the other, the Teisseyre-Tornquist Zone, runs SE from the Baltic coast in Poland to the Carpathians–Black Sea area in the Ukraine. In general, the Tornquist Zone is the longest tectonic zone in Europe, separating the East European Craton with its Palaeozoic cover from the Phanerozoic mobile belts of Central and Western Europe.

The area of present investigations is that of the Polish Lowland located at the border between the Precambrian crustal terranes of the EEC and the younger Phanerozoic terranes in the south-west (Berthelsen, 1992*a*, *b*, 1998). Much of Poland and northern Germany is covered by a deep (>10 km) sedimentary basin, filled with Permian and Mesozoic rocks. The Polish Permian–Mesozoic Basin forms the easternmost part of this Central European Basin (Dadlez, 1989; Ziegler, 1990; Pharaoh *et al*., 1997; Krzywiec, 2002*a*, *b*). The axis of the basin, the Mid-Polish Trough (MPT), parallels the edge of the East Euro-

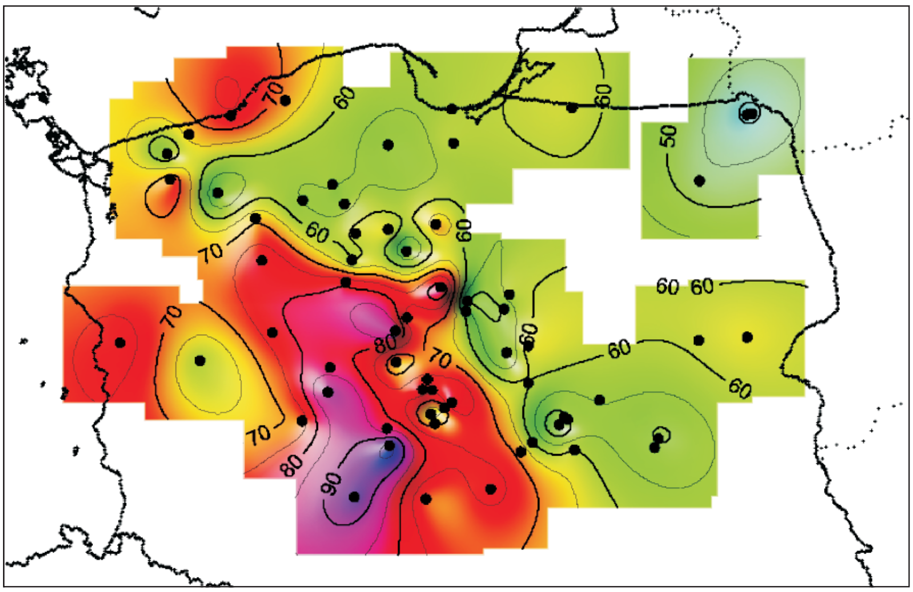

Fig. 2. Revised heat flow map of the Polish Lowland area according to Szewczyk and Gientka (2003)

Isolines described in mW/m^3 (compiled according to method of Wybraniec, 1999) pean Craton. The MPT is a part of the Trans-European Suture Zone (TESZ), a first order geotectonic unit, stretching from the British Isles, through Poland and to the Black Sea.

The axial area of the Polish Basin has a long history of subsidence, starting possibly already from the Early Palaeozoic (and probably from the Late Precambrian); the exact nature of this subsidence remains, however, a matter of discussion. The subsidence during the Permian–Mesozoic episode is attributed by Kutek (1997) and Karnkowski (1999) to a basin formed by an asymmetrical fault-bounded rift structure equivalent to the simple-shear model (Wernicke, 1981, 1985), with associated volcanism at 295–285 Ma (Early Permian). Other authors stress the generally symmetrical shape of the Mesozoic infill and attribute observed small-scale irregularities within the Triassic–Jurassic–Cretaceous sedimentary sequences to a highly asymmetrical inherited TTZ crustal structure instead of an asymmetrical extension mechanism (Dadlez *et al*., 1995). Stephenson *et al*. (2002) stressed the importance of basement heterogeneity in the response of the Polish Basin (PB) to Alpine tectonics. The heterogeneous nature of the basement structure of the PB does not permit an easy interpretation of first — order observations such as Moho depth in terms of simple models of basin subsidence and evolution. The inherited crustal thickness variations have played a role in the development of the basin (Stephenson *et al*., 2003). To some degree, local and regional irregularities observed within the Mesozoic sedimentary infill could also have developed because of a basin-scale decoupling caused by thick Zechstein evaporites (Krzywiec, 2002*a*, *b*) instead of the basin-scale asymmetrical extension.

The crust underneath the Mid-Polish Trough has been studied for many years (Guterch *et al*., 1986, 1991, 1992, 1994). The seismic experiment POLONAISE'97 conducted in May 1997 targeted the deep structure of the TESZ in northwestern Poland (profiles P1 to P5; Guterch *et al*., 1997, 1998, 1999). The results were presented by Guterch *et al*. (1999), Jensen *et al*. (1999), Środa *et al*. (1999), Wilde-Pirko *et al*. (1999) and Janik *et al*. (2002). In addition to the POLONAISE'97 profiles, two DSS profiles, LT-7 and TTZ, across the Polish Basin were reinterpreted (Guterch *et al*., 1994; Grad *et al*., 2002). The older ideas of deep Moho below the Polish Trough/TTZ have been significantly revised in recent publications (*cf*. Janik *et al*., 2002; Jensen *et al*., 2002). The TTZ in Poland is clearly situated in or close to a major Moho gradient zone.

The crustal structure of the EEC is revealed by profiles P3, P5 and the northeastern part of profiles P2, P4 LT-2, LT-4, LT-5 and LT-7 (Fig. 1). All models of the crust for this area exhibit nearly horizontal uniform structure. The crystalline crust consists of three parts: upper, middle and lower with P-wave velocities of 6.1–6.4, 6.5–6.7 and 7.0–7.2 km/s, respectively. The crystalline basement lies at a depth of 0.5–5 km, dipping steeply to the SW. In the northwestern part of the P5 profile, a body with high seismic velocities of 6.6 km/s was found 2–10 km deep, coinciding with the rapakivi-like and anorthosite Mazury complex. The depth of the Moho ranges from 39–45 km in northeastern Poland, reaching 50 km beneath Lithuania. Below the Moho, the P-wave velocity is 8.05–8.1 km/s.

The crustal structure of the Polish Basin and Palaeozoic Platform is characterised by profile P1, the southwestern parts of profiles P2, P4, LT-2, LT-4, LT-5 and LT-7, as well as the profile TTZ. Profiles P1 and TTZ are located parallel to the TESZ; profile P1 transects the zone of Variscan influence. Profiles P4, P2 LT-4, LT-5 and LT-7 cross the above-mentioned profiles and the edge of the East European Craton almost perpendicularly. In general, the P-wave velocities of the upper crust in the Polish Basin are low (< 6.1–6.2 km/s) down to 20 km depth. This can be interpreted as an evidence for sediments subjected to low-grade metamorphism. The lower crust has a P-wave velocity of 6.5–7.3 km/s, a high velocity gradient, and strong ringing reflectivity. The velocity of the uppermost mantle is high (> 8.2–8.4 km/s).

Recent reinterpretation of the seismic data along older profiles (LT-2, LT-4 and LT-5) generally confirms the results from the POLONAISE'97 profiles. Using the geometry of seismic boundaries and seismic velocities of P-waves, the successive layers in the models of the upper crust were classified as sediments (Cenozoic and Mesozoic), compact sediments (meta sediments, lower Palaeozoic–Neoproterozoic) and granitoids (upper crust). Medium to high grade metamorphic crystalline rocks are likely beneath these. A peridotite uppermost mantle is also likely. All cross-sections clearly show differentiation of the seismic structure, strong asymmetry of the Polish Basin both in the basement and in the shape of the Moho, variation in crustal layering and thickness, as well as sub-Moho velocity changes beneath both platforms (Krysiński and Grad, 2000; Grad *et al*., 2002).

The TESZ is clearly outlined by the morphology of magnetic and gravity fields (Tornquist, 1908; Jankowski, 1967; Grabowska and Raczyńska, 1991; Grabowska *et al*., 1991; Królikowski and Petecki, 1995; Królikowski and Wybraniec, 1996; Petecki, 2001, 2002) and to a lesser extent by the heat-flow data (Majorowicz and Plewa, 1979; Majorowicz, 1984; Čermak *et al*., 1989; Plewa, 1994; Gordienko and Zavgorodnaya, 1996; Karwasiecka and Bruszewska, 1997; Majorowicz *et al*., 2003; Szewczyk and Gietka, 2003).

The area of the TESZ in Poland is devoid of magnetic anomalies. Within the EEC marked, large scale (100 km) oval and belt-shaped magnetic anomalies are related to granitoid massifs, metamorphic belts and metamorphic-magmatic complexes of the crystalline basement (Ryka, 1984). The Palaeozoic Platform in southwestern Poland exhibits numerous local small-scale (10 km) magnetic anomalies (Karaczun *et al*., 1978; Królikowski and Petecki, 1995; Królikowski and Wybraniec, 1996). West of the TTZ in northeastern Poland a magnetic source at a depth of 18.5 km has been discerned and interpreted as being related to the top of the main anomalous structures in this area (Petecki, 2001). The northwestern part of Poland is the area in which Curie point is at a considerable depth of some 20–25 km according to Dąbrowski and Majorowicz (1977). This is also the area of high-density lower crust and upper mantle as shown from gravity and magnetic modelling along the LT-7 profile (Petecki, 2002). The area may represent a wedge of the EEC crust indenting the Palaeozoic crust or possibly the crust of a proximal or exotic terrane or of an island arc (Petecki, 2002). This area is also characterised by heat flow less than 60 mW/m^2, comparable to that of the marginal zone of the EEC. The central part of TTZ to the south of the above block is hotter, with heat flow comparable to that of the Variscan platform to the west.

Information on the depth of the asthenosphere is not available, although, "ringing reflectors" are found at the depth interval from 90–110 km (Grad *et al*., 2002). These are explained by relatively small scale heterogeneities beneath the depth interval from less than 90 km down to as deep as 110 km. A series of reflectors range from 50–90 km as shown by seismic data of the POLONAISE'97 experiment. Qualitative interpretation of the observed wave field points to a differentiation of the reflectivity in the lower lithosphere. The seismic reflectivity of the uppermost mantle is stronger beneath the Palaeozoic Platform and TESZ than beneath the EEC. The deepest interpreted seismic reflector with a zone of high reflectivity may mark a change in upper mantle structure from an upper zone characterised by seismic scatters of small vertical dimensions to a lower zone with vertically greater seismic scatters, possible caused by inclusions of partial melt (Grad *et al*., 2002). This can be related to estimates of the thermal lithosphere thickness of some 100 km or less in the area of high heat flow (Majorowicz *et al*., 2003 and this study). A thickness of 100 km is probable although the lithosphere reflectivity is not a sufficient proof.

GEOTHERMAL DATA AND HEAT FLOW

Heat flow in Poland has been studied from continuous temperature logs obtained in 231 wells with commercial logging equipment of low accuracy (0.1–0.5 K) and 10 wells with precise (better than 0.03 K) temperature-depth measurements made by thermistor probes, recorded in wells not deeper than 0.5 km in southwestern Poland (Plewa, 1994; Karwasiecka and Bruszewska, 1997; Majorowicz *et al*., 2002, 2003; Szewczyk, 2002; Szewczyk and Gientka, 2003). Data limitation is mainly due to well conditions. The main uncertainty is due to perturbed borehole temperatures measured 200–700 hours after drilling in deep wells (2000–6000 m). In some cases logs were taken only days after drilling ceased. Location of wells with logs taken at least 10 days after the drilling had stopped and the ones logged earlier were shown in Majorowicz *et al*. (2002). In some cases, temperatures, especially in the upper section, differ from equilibrium temperatures by as much as 10–15 K. Temperatures extrapolated to the surface differ more than expected from the ground surface mean annual temperatures (8°C+/–1; Plewa, 1994). Corrections applied to the geothermal gradient for non-equilibrium due to drilling are 10–15% in most cases. Another source of uncertainty is limited core sampling for thermal conductivity and limited accuracy of their measurements. Lack of sufficient thermal conductivity data due to scarce core sampling was substituted by the estimates of conductivity from the analysis of geophysical well logs (Szewczyk, 2001). It is estimated that in most cases heat flow values are accurate to within 20% +/–10%.

The temperature change of the Pleistocene–Holocene palaeoclimatic event in periglacial northern Europe and likely also in Poland could be as high as 10°C (Šafanda and Rajver, 2001). This may explain very low heat flow values (< 40 mW/m^2) observed in the northeastern part of Poland (Majorowicz, 1976; Majorowicz *et al*., 2002; Szewczyk, 2002), where wells are shallower than 1.5 km. The disappearance of the last ice sheet from the Polish Lowland some 15 ka ago most likely affected the subsurface thermal regime by lowering the geothermal gradient in the upper part of the sedimentary fill. The magnitude of this temperature change at the surface could amount to 10–15°C in the northern wells as the average temperature at the surface during Late Pleistocene (Weichselian) is estimated at –6.7°C (Szewczyk, 2002). Evidence of such disturbance is clearly evident from wells in northeastern Poland (Majorowicz, 1976; Szewczyk, 2002) and is well documented in wells north of Poland (Kukkonen *et al*., 1998). The most prominent case of disturbance due to palaeoclimatic conditions was observed in a group of wells in northeastern Poland in the low heat flow area with strong inversion of temperature gradient with depth (Majorowicz, 1976; Szewczyk, 2002). It is assumed that the correction in all of the area should be nearly the same.

The lowest heat flow values in northeastern Poland (EEC) are calculated in wells shallower than 1.5 km (Majorowicz *et al*., 2003). This may be a result of the climatic reduction of heat flow as described above. The corrections due to climatic change (Pleistocene–Holocene) are 12–15 mW/m^2 for the upper 800 m as shown by Kukkonen *et al*. (1998).

Heat flow contour maps of the study areas constructed by Majorowicz (1984), Plewa (1994), Gordienko and Zavgorodnaya (1996) and Karwasiecka and Bruszewska (1997) were based on data uncorrected for the Pleistocene–Holocene climatic change effect upon the upper hundreds of metres and down to a kilometre depth. Maps of heat flow data from Karwasiecka and Bruszewska (1997) were used in previous modelling. A new heat flow map (Szewczyk and Gientka, 2003; Fig. 2) was based on corrected heat flow values (with correction for the Pleistocene–Holocene change in surface temperature due to climatic changes and related withdrawal of the ice sheet from northern Poland) and is used for modelling in this paper. This map shows an increase in heat flow for northeastern Poland (especially for the shallower wells) when compared to the map by Karwasiecka and Bruszewska (1997).

The analysis of the heat flow pattern (Figs. 1 and 2) across the area in which about 3000 km of new DSS profiles exist shows significant variation, from low heat flow values of 40 mW/m^2 in NE corner of Poland to high values reaching 90 mW/m^2 in the south-western section of the TESZ. The analysis of the main heat flow features (Figs. 1 and 2) shows evidence of extensive crustal-mantle warming (Q > 60 mW/m^2) in the area between the Sudetes to the south and the TESZ to the north. The axis of the anomaly is aligned parallel to the Variscan deformation front and it cuts the central part of the TTZ.

New heat flow map based on 59 wells in the Polish Lowland (Fig. 2) for which correction for Pleistocene–Holocene surface temperature change was applied (Szewczyk and Gientka, 2003) confirms the general pattern as shown by the other heat flow maps cited above. It shows that the change in heat flow between the southwestern part of the TESZ and the northwestern part of the EEC is some 30 mW/m^2, well above the uncertainty limits of heat flow estimates (15–20%).

LITHOSPHERE TEMPERATURE MODELLING

The previous assumption of constant mantle heat flow of 20–30 mW/m^2 (25 mW/m^2) was derived from earlier models of heat generation in the crust based on the relationships between heat generation and seismic velocities for the older DSS deep seismic sounding profiles and the surface heat flow (Čermak *et al*., 1989; Majorowicz *et al*., 2002). That approach results in large lateral variations of crustal contribution to the heat flow from as high as 65 mW/m^2 in southwestern Poland to 5 mW/m^2 in northeastern Poland. It was found with the use of modelling methods described in detail by Stromeyer (1984) and Šafanda (1985), Čermak *et al*. (1989) and Majorowicz *et al*. (2003), that derived crustal heat flow variations show little correlation between the thickness of the upper crustal high-heat generating zones (sediments, meta sediments and granitoids) and heat flow for the area west of the EEC (Majorowicz *et al*., 2003). The obvious discrepancy of crustal heat flow values between the EEC and the Palaeozoic Platform is due to a large contrast in the surface heat flow. Allowing a larger contrast in the heat flow from the mantle would solve these discrepancies.

In this work I re-analyse (in the light of a new heat flow map) the earlier models based on profiles P4, LT-7, P3 (Majorowicz *et al*., 2003) in addition to the newly studied profiles P2, P4, P5, TTZ, LT-2, LT-4 and LT-5. Changes in the heat flow map due to corrections for Pleistocene–Holocene surface temperature change were used in reinterpretation of previously published models for LT-7 (its northern part), P4 and P3.

Finite difference numerical modelling has been carried out for the above profiles in vertical 2D geothermal models running along the seismic profiles. The geothermal model is determined by heat production, thermal conductivity and surface heat flow.

The length of each profile was determined by the length of the seismic lines and by the availability of heat flow data of high quality and the depth to the 1300–1400°C isotherm. For temperatures below 1300–1400°C, the model assumption about pure conduction is no longer valid. In the asthenosphere, significant transport of heat by movements of material (advection) is present. I have omitted modelling results below that depth. The surface temperature was kept at 0°C. The observed surface heat flow along the profiles (Majorowicz *et al*., 2003; Szewczyk and Gientka, 2003; Figs. 1 and 2) was used to calculate the mantle heat flow at a given heat-generation model.

The model of heat production of the layer just below the sedimentary-metamorphic complex for which we used constant values (described below) is based on the empirical relationship between radiogenic heat production and compression seismic velocity, proposed by Rybach and Buntebarth (1982, 1984). The seismic velocity cross-sections along the profiles were first corrected for the effects of temperature and pressure, and then converted into the heat production values according to Čermak and Bodri (1986), which distinguishes between Phanerozoic and Precambrian rocks. I assumed a correlation of sub-sedimentary heat flow with heat generation at the top of the granitic layer underlying the sedimentary layer. The sub-sedimentary heat flow is here defined as surface heat flow less heat flow contribution of the upper layers of sedimentary rocks and metamorphosed sedimentary rocks.

The heat production of the uppermost sedimentary layer was not calculated from the heat generation vs. seismic velocity relationship. It was estimated at 1.0 μWm^{-3}–1.2 μWm^{-3} (EEC) and 2.0 μWm^{-3} (west of the TTZ) for the thick layers of metamorphosed sedimentary rocks reaching to a depth of 20 km. Heat generation of the mantle was estimated at 0.01 μWm^{-3} (Precambrian) and 0.015 μWm^{-3} (Phanerozoic). Heat production variations in the sedimentary layer influence significantly our estimate of the mantle heat flow and mantle temperature. Lower heat production values than the ones above, tried for several models, increase mantle heat flow and mantle temperatures to unreasonably high values. Heat flow from below the sedimentary rocks and metamorphosed sedimentary rocks is Q_r+*Ab* as:

$$Q = Q_s + Q_r + Ab$$

where: Q — observed heat flow at the surface; Q_s — heat flow generated in the upper layers of the sedimentary rocks and metamorphosed sedimentary rocks; Q_r — the "reduced heat flow" originating from below the layer of thickness (b) and heat generation (A) taken to be dependent on seismic velocity.

According to the heat generation model of Lachenbruch (1971) any vertical distribution of heat production $A(z)$ satisfies the above equation provided that the integral of $A(z)dz$ over the depth from 0 to infinity is $b\,A$. This is satisfied by the model of heat generation distribution:

$$A(z) = A\exp(-z/b)$$

A(z) calculated from the above formula was used for the "granitic" crust assuming b = 7 km for the Variscan areas and b = 10–15 km for the Caledonian–Precambrian areas. The values of A at the top of the granitic layer were estimated from the relationship between heat generation and composition (based on seismic velocity) described by Rybach and Buntebarth (1982, 1984). The heat generation for the lower crust was assumed to be 0.2 μWm^{-3}.

The model of thermal conductivity is based on the general model published by Chapman and Furlong (1992). The conductivity (k) is explicitly pressure- (depth) and temperature-dependent in the model according to the formula:

$$k(T,z) = k_0(1+cz)/(1+bT)$$

where: T — temperature (°C); z — depth; b and c — constants; k_0 — the conductivity measured at 0°C and atmospheric pressure.

We used a coefficient (b) of 0.0015 K^{-1} for the upper crust which is a value intermediate between experimentally determined values for granite and granodiorite. For the middle and lower crust a temperature coefficient (b) of 0.0001 K^{-1} was used. A negative value of b = –0.001 K^{-1} was assumed within the sedimentary layer in order to simulate diagenesis. A pressure coefficient (c) of 0.0015 km^{-1} is used throughout the crust and uppermost mantle. Surface (z = 0) thermal conductivity k_0 is 2.2 Wm^{-1}K^{-1} for sedimentary rocks, 3.0 Wm^{-1}K^{-1} for the upper (granitoids) and 2.6 W m^{-1}K^{-1} for the middle and lower

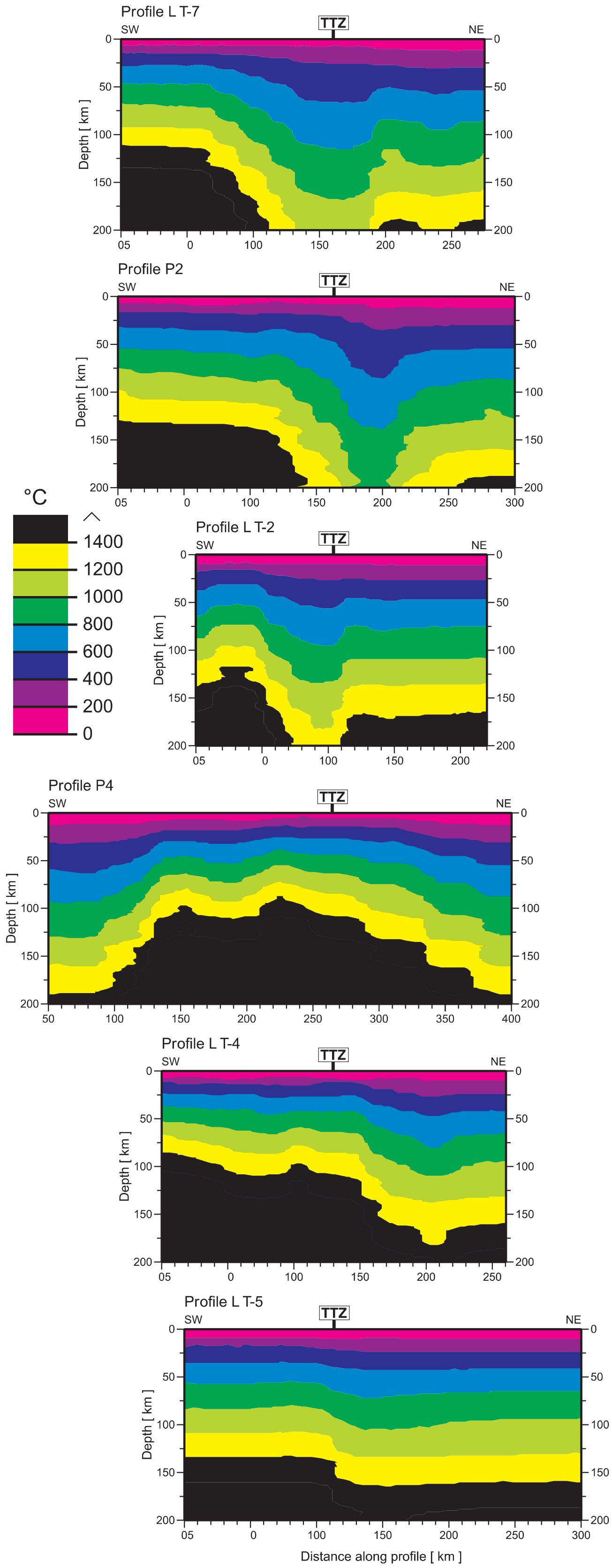

Fig. 3. Model of temperature distribution in the upper 200 km for the profiles perpendicular to the strike of the TTZ

crust, respectively. The mantle in the models between the Mohorovicic discontinuity and the depth of 50 km, probably of peridotitic composition, was assigned values of $k_0 = 4$ Wm^{-1}K^{-1} and $b = 0$.

Using the above model assumption, numerical models of the crustal and upper mantle temperatures were constructed along two dimensional distance-depth slices cutting through the transitional zone between the EEC, the TTZ and the Palaeozoic Platform south-west of the TTZ (Fig. 3). Profiles TTZ and P5 parallel to the strike of the TTZ have also been analysed (Figs. 4 and 5). Profiles P1 and P3 were analysed in a previous study (Majorowicz *et al*., 2003).

I have constructed maps of heat flow at the Moho surface (mantle heat flow contribution — Fig. 6), and of the distribution of thermal lithosphere thickness as defined by the 1300°C isotherm in the Polish Lowland (Fig. 7).

MODEL RESULTS

The new modelling results show that mantle heat flow contribution (heat flow at the Moho) increases after crossing the TTZ southward by 10–20 mW/m^2. In the northern part of Poland the mantle heat flow is low (20 mW/m^2) considerably increasing towards the Variscan area to the south-west (30–40 mW/m^2). There is a significant difference in a mantle heat flow contribution between the northern and central-southern parts of the TTZ, with the respective values increasing from 20 mW/m^2 to 30–40 mW/m^2. The large-scale variations of mantle heat flow are one of the possible causes of large-scale regional variations of the observed surface heat flow. Due to the limited accuracy of observed surface heat observations minor variations in mantle heat flow cannot be resolved. Surface heat flow variations can also be caused by lateral changes in heat production or thermal conductivity.

Results of calculation of the depth to the 1300°C isotherm which approximately can be considered the temperature of the top of the asthenosphere (Fig. 7) show variations similar to those reported by Balling (1995). The thermal lithosphere (peridotite solidus) is less than 100 km thick in the Variscan Platform while being thicker (100–200 km) in the craton and in the northern part of the TTZ in northwestern Poland. These values are similar to the thickness of thermal lithosphere obtained from the modelling of heat flow, gravity and topography in the Western Carpathians and their foreland basin by Zeyen *et al*., 2002. They gave a value of 115–40 km for the region between the present study area and the Carpathians.

Lithosphere thickness variations from 85 to 110 km beneath the Danish and German basins and the northern Tornquist Zone, to about 150 km in the central Baltic Shield areas, and 200–250 km in the areas of low heat flow along the northern Baltic Shield profiles (Balling, 1995) are closely associated with variations in heat flow from the uppermost mantle and perhaps with lateral differences in upper mantle heat flow caused by geodynamic process. These results are in a very good

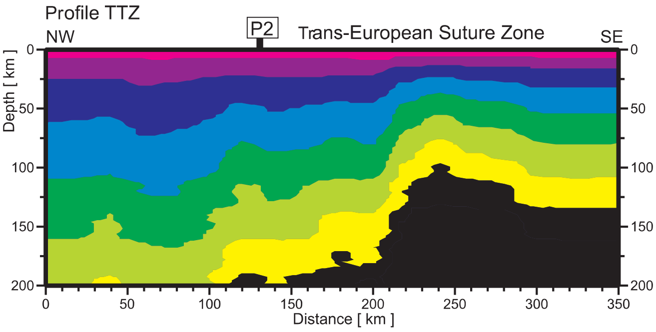

Fig. 4. Model of temperature distribution in the upper 200 km for the profile TTZ

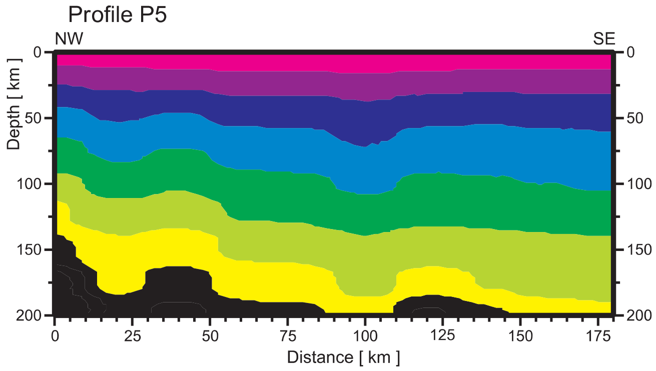

Fig. 5. Models of temperature distribution in the upper 200 km for the profile P5

agreement with seismological information from surface wave dispersion analysis and the refraction wide angle seismic interpretation of FENNOLORA data (Balling, 1995). Zeyen *et al.*, (2002) showed that much thinner lithosphere underlies the Pannonian Basin (75 km) whereas it is much thicker under the Bohemian Basin (90–120 km).

LEVEL OF UNCERTAINTY

New modelling study allows estimation of the uncertainty of the result due to uncertainty in the heat generation model. Heat generation from the upper metamorphosed sedimentary rocks beneath the sedimentary layer was assumed to be 2 $\mu W/m^3$ and its uncertainty is likely +/–0.5 $\mu W/m^3$. I have calculated that the increase of the heat production in that layer by only 0.5 $\mu W/m^3$ leads to a temperature decrease at 40 km depth (approximately Moho position) by 30–50°C. This is due to a decrease in the mantle heat flow. Therefore, the results of modelling presented need to be treated with caution as temperature distribution depends strongly on the heat generation model assumptions.

To test uncertainty in the temperature calculation with depth a simple 1D model was chosen. Temperature variation with depth can be calculated using equation:

$$T(z) = T_b + Q_0 z K^{-1} + A_0 D^2 k^{-1}(1 - \exp(-z/D))$$

where: $Q_0 = Q - A_0 D$; T_b — temperature at the basement surface and D = 10 km.

Conductivity (k) for the crust and upper mantle was assumed according to Jessop (1990, table 11). The contribution of heat generation from sedimentary rocks is estimated to be low (1–3 mW/m^2). Heat generation values A_0 = 2 mW/m^3 (low) and 4 mW/m^3 (high) were assumed for the study area.

Calculated lithospheric geotherms (Fig. 8) are for two heat flow regions in the Palaeozoic Platform: 70 mW/m^2 (low) and 90 mW/m^2 (high). Assumption of lower heat generation results in an increase of the reduced heat flow (deep heat flow from below the highly radioactive upper 10 km of the crust) and higher temperatures at the crust and upper mantle. Assuming Q = 90 mW/m^2 and A_0 = 2mW/m^3 the geotherm reaches the peridotite solidus boundary at 50 km which is improbable as the crustal thickness in the region can reach up to 40 km. For values of Q = 70 mW/m^2 and A_0 = 2mW/m^3, and Q = 90 mW/m^2, and A_0 = 4 mW/m^3 geotherms are more realistic as they both meet the peridotite solidus boundary at some 70 km. Low heat flow Q = 70 mW/m^2 and high heat generation A_0 = 4mW/m^3 gives the largest thickness of the solid lithosphere (130 km). The preferred isotherms are thus the ones for Q = 70 mW/m^2 and A_0 = 2mW/m^3, and Q = 90 mW/m^2 and A_0 = 4mW/m^3 combinations. In these combinations high heat flow is related to higher heat generation and lower heat flow to a lower heat generation in compliance with theory.

The above uncertainty in heat generation knowledge influences models of the strength of the lithosphere which varies in time and space and is a function of composition, temperature and pressure. The predicted depth of the brittle-ductile transition depends strongly on composition and geotherm. At the first order of approximation, the crustal transition is at the depth $z = z^*$ where the brittle and ductile critical stress difference is equal (Ranalli, 2000) to :

$$\beta \rho g z(1-\lambda) = (\varepsilon / A)^{1/n} \exp(E / nR(T(z))$$

where: β — a function of frictional parameters and the orientation of the stress field; ρ — the average density above depth (z); g — the gravitational acceleration; λ — the ratio between hydrostatic and lithostatic pressure; ε — the ductile strain rate; R — the universal gas constant; A, n, and E — the power-law creep parameters; $T(z)$ — temperature in (K).

Geotherms, depth to Moho, and crust and mantle structures allow estimation of the rheological profiles (strength envelopes) and study of the uncertainty of rheology due to temperature uncertainty. Rheological profiles (extension) based on the crustal composition parameters from Ranalli (2000, table 1) and known geotherms for the study area (Fig. 8) are shown in Figure 9. Frictional strength was calculated for λ = 0.4 for simplification. Ductile creep strength is related to different geotherms, which vary for different Q and A_0 parameters. When heat flow is low and heat generation is high the crust and upper mantle are brittle (Fig. 9a). With low heat flow and low heat generation and with high heat flow and high heat generation (similar geotherms, Fig. 8) the brittle/ductile transition is at

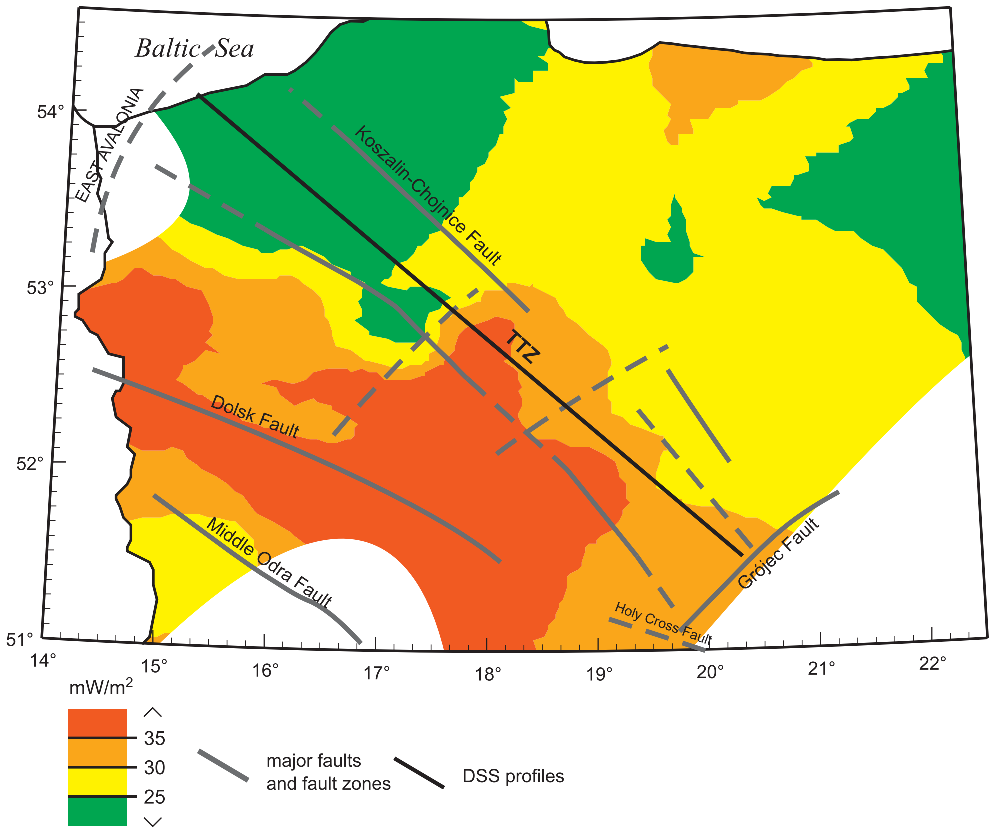

Fig. 6. Contour map of the Moho heat flow derived from geothermal modelling of seismic profiles: P1, P2, P3, P4, P5, TTZ, LT-2, LT-4, LT-5 and LT-7

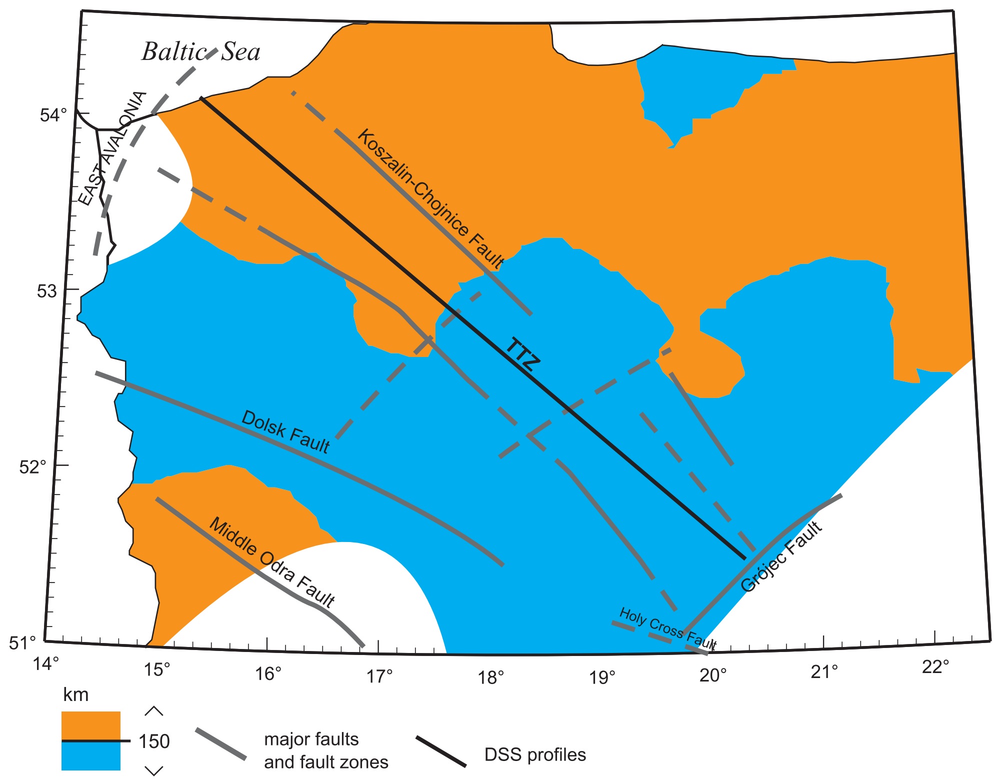

Fig. 7. Contour map of the depth to the asthenosphere (approx. at 1300°C) derived from geothermal modelling of seismic profiles P1, P2, P3, P4, P5, TTZ, LT-2, LT-4, LT-5 and LT-7

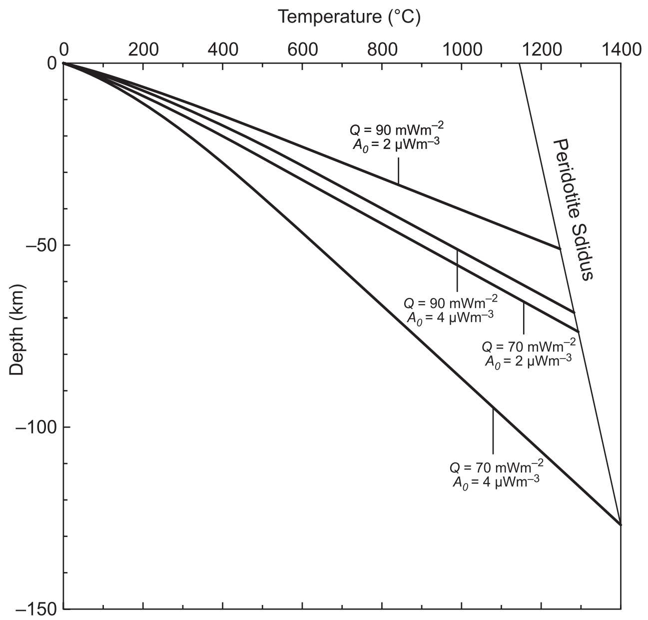

Fig. 8. Lithospheric geotherms — sensitivity analysis (different heat flows — different heat generation assumed)

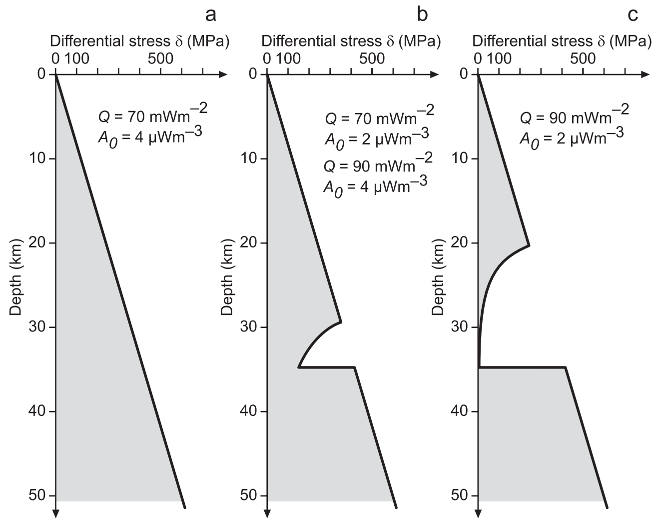

Fig. 9. Crustal and upper mantle extension rheology (for dry Pg) — sensitivity analysis

some 30 km (Fig. 9b). Crust-mantle strength is relatively high with weakening of the lower 5 km of the crust. In our case (Fig. 9b) calculated temperatures result in strong lower crust and no pronounced lower crustal weak zone. This would be different if high heat flow were explained by high mantle upflow as in Figure 9c. The calculations suggest that a high strength of the lithosphere is typical for the study area for the lower heat flow zone of the basin. However, a larger strength discontinuity at the Moho is a possibility in the case of high heat flow and moderate heat generation (90 mW/m^2 and 2 mW/m^3 respectively) as shown in Figure 9c. In that case high heat flow and large reduced heat flow (Q_r) would cause much higher lithosphere temperatures and thinner thermal lithosphere. Wetter rheology would weaken the lower crust even more.

DISCUSSION

Relatively high temperatures at the Moho in the high heat-flow area in southwestern Poland (600–750°C) contrast with high (P_n) seismic velocities at the top of the mantle (8.25–8.4 km/s; Jensen *et al.*, 1999; Guterch and Grad, 2000). These high (P_n) velocities suggest a rather cool top of the mantle and low heat flow through the study area, including the frontal zone of the Variscan area (Fore-Sudetic Monocline) with 30–40 km crust and the areas with thicker crust (40–50 km) in the craton (EEC). Statistical relationships between heat flow, and P_n velocities and heat flow and temperature are shown to exist world-wide (Black and Braile, 1982) including in Europe (Kubik, 1986).

The latter relationship is expressed as:

$$P_n = 8.55 - (0.73 \times 10^{-4}) \times T(P_n)$$

where: $T(P_n)$ — temperature at the Moho.

In the case of similar mineralogy the above formula allows the use of the (P_n) velocities to estimate deep temperatures. A high (P_n) velocity of 8.1–8.35 km/s was interpreted from seismic experiments for the Palaeozoic and Precambrian platforms in Poland (Jensen *et al.*, 1999; Guterch and Grad, 2000) resulting in temperatures between 550 and 270°C, respectively. While the first number is in fairly close agreement with the modelled Moho temperature in the area of thick Precambrian crust, the second number is too low when compared with modelling (600–750°C) of the Palaeozoic Platform, where P_n is 8.25–8.4 km/s (Jensen *et al.*, 1999). It is also quite improbable. High (P_n) velocities also suggest low heat flow in this area in disagreement with the observed high heat flow (Figs. 1 and 2).

The following considerations need to be taken into account:

1. The temperature coefficient of velocity for mantle composition rocks is quite well established in laboratory experiments. Thus, if there is not a similar relationship between (P_n) velocity and Moho temperature, we need to seek an explanation.

2. Areas of localised structural complication must be avoided in using this relationship. In particular we exclude the velocity data near the area of the fossil subduction zone, where uncertainties are large.

3. We can recognise at least two primary sources of error for application of this relationship:

— compositional variations, such as eclogite masquerading as "mantle"; craton mantle being more refractive than younger areas *etc.*;

— anisotropy.

This is well recognised in the oceans, but not often resolved beneath continents. Only the TTZ has azimuthally sufficient coverage and some anisotropy was found.

4. None of our methods for constraining deep temperatures are completely reliable on their own, including models based upon heat flow-surface heat generation. We need to apply all of the available constraints, including (P_n)vs. Moho temperature, thermal isostasy elevation vs. crustal thickness, the elastic lithosphere thickness analysed from gravity-topography coherence with wavelength, xenoliths *etc.*

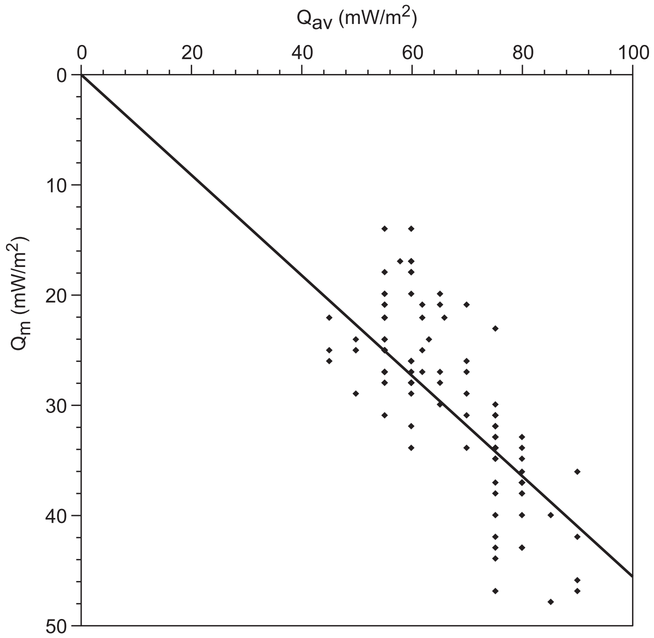

Fig. 10. Relationship between observed surface heat flow (Q_{av}) and mantle heat flow (Q_m) derived from geothermal modelling of seismic profiles: P1, P2, P3, P4, P5, TTZ, LT-2, LT-4, LT-5 and LT-7; mantle heat flow contribution is approximately 46%; lower crustal contribution to surface heat flow is approx. 14%

At this point we recognise that the relationship between (P_n) velocity and temperature does not work for the study area of the Variscan Platform and farther study needs to be undertaken to explain this discrepancy.

Mantle heat flow variations modelled for the Polish area (20–40 mW/m^2) are smaller than contrasting values found for the structurally oldest and youngest regions along the European Geotraverse (20–60 mW/m^2). Our model temperatures and heat flow at the Moho agree with models across the Baltic Shield and the Tornquist Zone north of Poland (Balling, 1995). Balling (1995) estimated that the mantle heat flow varies from 20–25 mW/m^2 along the northern and central shield transect increasing to 30–40 mW/m^2 beneath the North-German and Norwegian–Danish Basins and northern Tornquist Zone. In Poland, mantle heat flow along the TTZ is about 20–40 mW/m^2 and increases to some 40 mW/m^2 in the area of the Variscan Deformation Front. Lower values of mantle heat flow (20–30 mW/m^2) were estimated by Förster and Förster (2000) for the Bohemian Massif (Variscan Erzgebrige in Germany) to the south-west of the Polish high heat flow zone in the Variscan Front area. These values are less by 10–20 mW/m^2 than in the Variscan area of Poland and better agree with the TTZ values than with those for the Variscan Deformation Front areas of the Polish Basin (this work) and German and Danish basins to the north (Balling, 1995).

Mantle heat flow contribution (Q_m is heat flow below the Moho) was statistically correlated with surface heat flow. It turns out that in Poland mantle heat flow is some 50% of the surface heat flow and good statistical correlation exists between Q_m and Q (Fig. 10). For the comparison, the worldwide relationship between continental heat flow and reduced heat flow (Q_r) (Pollack and Chapman, 1977; Vitorello and Pollack, 1980) shows that Q_r is some 60% of the surface heat flow. Q_r represents heat flow from deeper sources below the upper crust. Our relationship suggests that the lower crustal contribution would be close to 10%.

The problem is how to explain the differences in mantle heat flow density (HFD). They are hardly caused by heat production heterogeneity. Even if the mantle production varies by 100%, for instance 0.02 µW/m^3 instead of 0.01 µW/m^3, a difference in the contribution of 100 km of mantle to the HFD is only 1 mW m^{-2}. But the differences in the mantle HFD (if any) might be caused by different cooling of the lithosphere, which was heated during its tectonic history. According to Vitorello and Pollack (1980) this transient thermal perturbation can be 5–10 mWm^{-2} in Variscan units as the thermal process with time depends on the duration of changes, the thickness of the basin fill, the crust, the thermal lithosphere, and thermal properties within the lithosphere.

A thermal time constant of changes controlled by conduction may be derived from the dimensionless time parameter that is common in all analysis of the transient conductive problem. This parameter was defined by Jessop and Majorowicz (1994) as follows:

$$Pt = st / a^2$$

where: s — diffusivity; t — time; a — the size of the zone or body, equivalent to depth in a one dimensional context.

This parameter usually appears as an exponent and if Pt is made equal to unity, we obtain the time constant:

$$t = a^2 / s$$

The above time constant can be regarded as the time taken for the transient thermal event to progress to about 63% of its completion. If we assume s = 1.2 mm^2/s^{-1} a crust of thickness 30 km will have a time constant of 30 Ma. For a 50 km thermal lithosphere with its base at approx. 1300°C it will take some 70 Ma for the transient cooling to progress to 63% of its completion and thicken the lithosphere, and for 100 km lithosphere some 200+Ma. This is relevant to change of the thermal state (in this case cooling of the crust and lithosphere) by conduction only. As it takes millions of years for the crust/lithosphere to cool, few processes in sedimentary basins are in a steady state. This may allow to explain the differences in temperature of the lithosphere/crust and basin between different tectonic units as observed in Poland. It can also be speculated that regional continental drift, which occurs mainly towards the north and north-east (J. Nawrocki, pers. comm.) has shifted the lithosphere by several degrees over a potential hotter mantle.

The difference in calculated mantle heat flow contribution (heat flow at Moho surface, Fig. 5) is larger than 10 mW/m^2, and this cannot be explained by variations in the mantle heat generation as discussed above. The explanation of such variations in Moho heat flow by the Variscan cooling remnant heat (5–10 mW/m^2) also falls short of the observed differences. However, it is possible that Alpine rejuvenation of the mantle heating engine occurred in the Polish Basin. The other explanation would be for much more modest mantle heat flow variations (10 mW/m^2) and larger lateral variations in heat generation of the upper sedimentary and granitic layers. It should be also noticed that a high contrast in heat flow between the EEC and the Palaeozoic Platform existed in the past as shown by analysis of vitrinite reflectance data (Majorowicz *et al*., 1984).

It is assumed that due to a decrease in sediment porosity and permeability with depth the heat flow advection caused by hydrodynamic flow would be low at greater depths. Lateral variations of heat flow in the central part of the TESZ in the Polish Basin could be also explained in some areas by the presence of highly conductive salt domes. Majorowicz *et al*. (2003) concluded that salt tectonics would not explain major differences between low and high heat flow areas, although measured heat flow values in the TTZ area can be 10% higher than in areas devoid of salt structures. The solution to this problem requires 3D modelling not addressed here.

Isostasy offers another explanation of heat flow variations and the thermal state of the crust and upper mantle. Thick crust and low elevation indicate low temperatures (thick lithosphere). Thin crust and high elevation indicate high temperatures and thin lithosphere (Bodri and Bodri, 1985; Hyndman and Lewis, 1999). High heat flow and hot crust and mantle in the Palaeozoic Platform characterised by low crustal thickness versus low heat flow in the areas of thick crust of the Eastern European Craton and the TTZ are confirmed by the first order model for the variations in crustal thickness required for isostatic balance (Bodri and Bodri, 1985; Hyndman and Lewis, 1999). Small elevation changes observed along the low elevation area of LT-7 (altitude less than 200 m) and high variations in crustal thickness from 30–45 km require heat flow varying from 35–75 mW/m^2, respectively. This is close to the variation in the observed heat flow between profiles of the Palaeozoic and Precambrian platforms respectively.

Large variations in mantle heat flow contribution and variable topography of the asthenosphere are only two possible explanations of the observed variations in surface heat flow. Large lateral variation of heat generation in the upper sedimentary zone and through the thick low velocity zone interpreted as low grade meta sedimentary rocks may be another possibility. There are independent constraints obtained, for example, from deep seismic data. The latter suggest large variations in the thickness of the lithosphere (100–200 km) supporting our interpretation of a significant variation in heat flow from the mantle.

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