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Heat flow models across the Trans-European Suture Zone in the area of the POLONAISE'97 seismic experiment

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Abstract

Heat flow data from the Polish basin show a sharp change in the transition from the East European Craton (EEC) and Teisseyre–Tornquist Zone (TTZ) in the north-east to the accreted terranes in the south west (Paleozoic Platform). The analysis of this data and numerical modelling of the crustal temperatures show evidence of extensive crustal–mantle warming in the area between the Sudetes to the south and the Trans-European Suture Zone to the north. The change in heat flow is 100% when compared with values for the EEC. Heat flow in the anomalous zone is also higher than in the Sudetes. The axis of the anomaly is aligned with the Dolsk Fault and Variscan deformation front. Low crustal/mantle temperatures derived from the relationship between temperature and P_n velocities (more than 8.2 and as high as 8.4 km/s) are at odds with high crustal temperatures calculated from surface heat flow, seismic velocity based heat generation models and thermal conductivity. High heat flow (Variscan platform) and related high temperatures of the crust coincide with small crustal thickness (30–35 km). The opposite is the case for the low heat flow EEC (45–50 km). High heat flow above thin crust and low heat flow above thick crust with no major variation in elevation is supported by a simple isostatic balance model. Crustal heat generation explains part of the high heat flow within the zone with thick meta-sediments reaching down to 20 km depth, however, it is far from explaining high heat flow in Variscan crust and in the transition zone into a cold EEC. 2D numerical models of heat flow based on new seismic data require a contrast of 15 mW/m^2 in mantle heat flow. High mantle heat flow (35–40 mW/m^2) is likely to occur in the high heat flow zone while cold crust and cold and high-density mantle (mantle heat flow of 20–30 mW/m^2) is typical of the EEC. Thermal lithosphere thickness for the craton is 200 km while it is only 100 km in the accreted terranes to the southwest of the TTZ. The TTZ in Poland appears as a relatively cold area.

Keywords: Heat flow; Geothermal modelling; Crustal structure; Seismic experiment; Poland; Teisseyre–Tornquist suture zone

1. Introduction

The crustal structure and thermal regime of the transition zone between the Precambrian and Paleozoic Platforms in Poland was one of the main aims of geophysical studies during the last 30 years. The tectonic position of Poland in Europe is unique and complex. Several large geological units of different age meet there: the East European Craton (EEC), Paleozoic mountain belts (the Caledonides and Variscides), the Paleozoic Platform of Central and Western Europe, and the South European Alpine orogene (Carpathian Mts). The SW margin of the EEC, between the North Sea and the Carpathians-Black Sea area, termed the Teisseyre–Tornquist Zone (TZZ), consists of two segments. The first one, the Sorgenfrei–Tornquist Zone, extends NW of Bornholm Island and the other, the TTZ, 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 feature in Europe separating the EEC with its Paleozoic cover from the Phanerozoic mobile belts of central and western Europe.

The deep crustal structure of that transition zone has been addressed in detail by a number of seismic refraction

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and wide-angle reflection experiments (Guterch et al., 1999; Jensen et al., 1999; Guterch and Grad, 2000). A large seismic experiment, the POLONAISE'97, was conducted in Poland during May of 1997 and targeted the deep structure of the Trans-European Suture Zone (TESZ). In addition to five POLONAISE'97 profiles (P1–P5) two deep seismic sounding (DSS) profiles (LT-7 and TTZ) pass through the Polish Basin. The quality of the data surpasses any of the previous DSSs for which heat flow modelling was completed (Majorowicz, 1978; Milanowsky, 1984 and Čermak et al., 1989). About 3000 km of high quality deep seismic profiles has shown that the Earth's crust in the TTZ has highly anomalous properties. The boundary velocities, mean velocities and stratification of the crust vary distinctly along the TTZ. This new seismic database allows a new look at the thermal state of the crust and comparisons with previous models from previous seismic experiments.

The heat flow at Moho and the temperature distribution in the crust and upper mantle were studied for the northern part of the TESZ (Balling, 1995). Mantle heat flow of 20–30 mW/m^2 was estimated for the northern and central Baltic shield transects, increasing to 30–40 mW/m^2 beneath the shield margin, deep basins (Danish and North German), and Tornquist zone (Balling, 1995). Föerster and Föerster (2000) estimated mantle heat flow of 20–30 mW/m^2 for the Variscan Erzgebrige in Germany.

In this paper we construct two-dimensional (2D) numerical models of the crustal and upper mantle temperatures in the transitional zone between the craton (Baltica) and the areas south west of the TTZ.

2. Geological background

The area of investigation in the Polish lowland is located at the border between Precambrian crustal terranes of the EEC and the younger Phanerozoic terranes in the southwest (Berthelsen, 1992a,b, 1998). Much of Poland and northern Germany is covered by the deep (>10 km) sedimentary basin, filled with Permian and Mesozoic sedimentary rocks. The Polish Basin forms the easternmost part of this Central European Basin. It is situated on the contact between and on the edges of the Paleozoic and Precambrian platforms. Its southwest border is formed by the Bohemian Massif (Dadlez, 1989; Ziegler, 1990; Pharaoh et al., 1997). The axis of the basin, the Mid-Polish Trough (MPT), parallels the edge of the EEC. The MPT is a part of the TESZ, a first-order geotectonic unit, stretching from the British Islands, through Poland to the Black Sea. In Poland, this axial zone has its own specific crustal structure of intermediary type.

During the late Cretaceous and Tertiary, the Alpine orogeny caused inversion of the European Mesozoic epicontinental sedimentary basins extending to the NW into Denmark and Sweden (Berthelsen, 1992a,b). During this inversion, the axial part of the MPT was significantly uplifted and eroded (see Krzywiec, 2002a for recent summary and further references). The Holy Cross Mountains in south central Poland exposes Palaeozoic structures and are the best known evidence of this uplift. The axial area of the Polish Basin has a long history of subsidence, possibly from the Early Palaeozoic (and probably from the Late Precambrian), the exact nature of this subsidence remains still 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 obeying a simple-shear model (Wernicke, 1981, 1985), with associated volcanism at 290–270 Ma (Early/Late Permian). Other authors stress rather the symmetrical shape of the Mesozoic infilling and attribute observed small-scale irregularities within the Triassic–Jurassic–Cretaceous sedimentary sequences to inherited highly asymmetrical TTZ crustal structure instead of asymmetrical extension mechanism (Dadlez et al., 1995). To some degree, local and regional irregularities observed within the Mesozoic sedimentary infilling could have also developed because of basin-scale decoupling caused by thick Zechstein (Upper Permian) salt (Krzywiec, 2002a,b) instead of basin-scale asymmetrical extension.

3. Geophysical investigations of the study area

The TESZ is clearly visible as a tectonic unit of distinct crustal structure (Guterch et al., 1986), 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) as well as the heat-flow data (Majorowicz and Plewa, 1979; Čermak et al., 1989), contrasting with neighbouring lithosphere blocks. In general, the EEC in NE Poland is characterised by a 0.5–5 km thick sedimentary cover, 42–47 km thick crust, cold lithosphere with relatively low heat flow of 40–50 mW/m^2, and an age of 2000–800 Ma. In the Paleozoic platform in SW Poland the sedimentary cover is up to 8 km thick, the thinner crust is 28–34 km thick, hot lithosphere has much higher heat flow of 60–90 mW/m^2 and an age of about 450–290 Ma (Majorowicz and Plewa, 1979; Majorowicz, 1984; Plewa, 1994; Gordienko and Zavgorodnaya, 1996; Karwasiecka and Bruszewska, 1997; Guterch et al., 1986; Čermak et al., 1989; Guterch and Grad, 1996; Pharaoh et al., 1997).

The area of TESZ is devoid of magnetic anomalies. Within the EEC strong and 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 Paleozoic Platform in south-western Poland exhibits numerous local, small-scale (10 km) magnetic anomalies (Karaczun et al., 1978; Petecki, 2001; Królikowski and Wybraniec, 1996).

The morphology of the gravity field coincides with tectonic structures clearly aligned in the SE-NW direction. The basic feature of Bouguer anomalies in the Polish Basin is the presence of an extensive depression down to −60 mGal; on this background, an increase of up to +15 mGal is noted over the Mid-Polish anticlinorium. The Fore-Sudetic monocline in the south-west and the East European Platform in the north-east are characterised by positive gravimetric anomalies of up to +20 and +10 mGal, respectively (Królikowski and Petecki, 1995).

The crust underneath the MPT thickens beneath the central part (Guterch et al., 1986) while depth to the Moho becomes shallow beneath the north-western part of the basin (Guterch et al., 1991, 1992, 1994). The seismic experiment POLONAISE'97 conducted in May 1997 targeted the deep structure of the TESZ in north-western Poland (Guterch et al., 1997–1999). Preliminary results were presented in the special issue of Tectonophysics (Guterch et al., 1999; Jensen et al., 1999; Środa and POLONAISE Working Group, 1999; Wilde-Piórko et al., 1999). In addition to the POLONAISE'97 profiles two DSS profiles LT-7 and TTZ across the Polish Basin are discussed (Guterch et al., 1994; Wilde-Piórko et al., 1999); (Fig. 1).

Profiles P3, P5 and the north-eastern part of profiles P4 and LT-7 (Fig. 1(b)) reveal the crustal structure of the EEC. 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 strongly to SW, almost perpendicularly to the edge of the craton. In the north-western part of profile P5, a body with high seismic velocities of 6.6 km/s was found at 2–10 km deep, coinciding with the rapakivi-like and anorthosite Mazurian complex, well known from geology. The depth of Moho ranges from 39 to 45 km in north-eastern Poland, reaching 50 km beneath Lithuania. Below the Moho P-wave velocity is 8.05–8.1 km/s.

The crustal structure of the Polish Basin and Paleozoic Platform is described by profile P1, south-western part of profiles P2, P4 and LT-7, as well as profile TTZ. Profiles P1 and TTZ on the Polish Basin are parallel to the general lineament; profile P1 is in the zone of Variscan influence. Profiles P4, P2 and LT-7 cross the above mentioned profiles and the edge of the EEC 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 of depth. It can be interpreted as evidence for meta-sediment of 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 (>8.2–8.3 km/s).

Using the geometry of seismic boundaries and seismic velocities of P-waves, the successive layers in the models were classified as sediments (Cenozoic and Mesozoic), compact sediments (meta-sediments, lower Palaeozoic–Neo-Proterozoic), granitoid (upper crust), basalt (middle crust), gabbro (lower crust) and peridotite (uppermost mantle). All cross-sections clearly show differentiation of the seismic structure, strong asymmetry of the Polish Basin both in the basement and in the Moho shape, variation in crustal layering and thickness, as well as sub-Moho velocity changes beneath both platforms (Krysiński et al., 2000).

4. Geothermal data and heat flow

A recent critical review of Polish well temperature logs (Plewa, 1994; Karwasiecka and Bruszewska, 1997; Majorowicz et al., 2002; Szewczyk, 2001) addressed their limitations due to well conditions (equilibrium state) and quality of industrial continuous logging. Continuous temperature logs obtained in 231 wells (Fig. 1(a)) with commercial logging equipment is of low accuracy (0.1–0.5 K). Ten precise (better than 0.03 K) temperature-depth measurements with thermistor probe were recorded in wells not deeper than 0.5 km in south-western Poland (Majorowicz et al., 2001). While accuracy of commercial equipment is a concern, the main uncertainty is due to perturbed borehole temperatures measured 200–700 h after drilling in deep wells (2000–6000 m) (Popov et al., 1998). In some cases logs were taken only a few days after drilling ceased. Location of wells with logs taken at least 10 days after the drilling stopped and the ones logged earlier are shown in Fig. 1(a). Temperatures especially in the upper section differ from equilibrium temperatures by as much as 10–15 K. Temperatures extrapolated to the surface are higher than those expected from the ground surface mean annual temperatures (9 ± 1 °C; Plewa, 1994). Corrections applied to the geothermal gradient for non-equilibrium due to drilling (Karwasiecka and Bruszewska, 1997) and constraints by surface temperature are 10–15%. Another source of uncertainty is limited core sampling for thermal conductivity and accuracy of their measurements. Heat flow values are accurate within $20 \pm 10\%$ (mW/m^2) on the average. We assume that the large variations in heat flow (Figs. 1(b) and 2) from <40 to >80 mW/m^2 are significant.

Significant uncertainty in heat flow determined from shallow temperature data (upper few hundred to a thousand meters) are due to poor knowledge of the

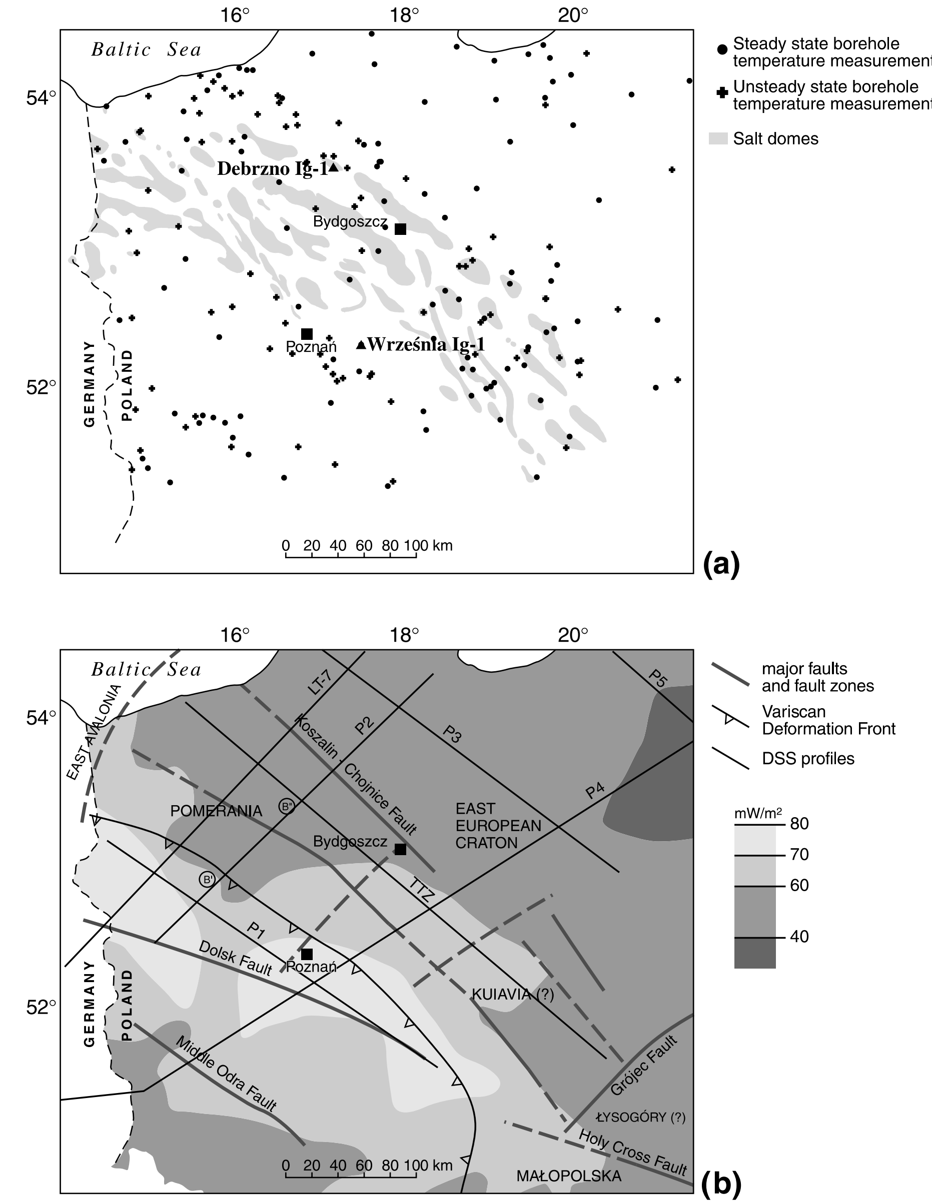

Fig. 1. (a) Location of wells with temperature logs in the study area of the POLONAISE'97 seismic experiment and salt tectonic structures (grey). (b) Main tectonic: structural elements of the Polish lowland according to Dadlez (2000) and locations of the new generation seismic profiles (Guterch and Grad, 2000) shown on the heat flow pattern (Majorowicz, 1984; Majorowicz et al., 2002).

temperature amplitude of the Pleistocene-Holocene paleo-climatic event in Poland. This may explain the lowest heat flow values (<40 mW/m^2) observed in north-eastern part of Poland (Figs. 1(b), 2 and 3) where wells are shallower than 1.5 km. The disappearance of the last ice sheet (12–15 kyr before present) from the Polish lowland probably affected the subsurface thermal regime 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 °C in the northern wells. Evidence of such a disturbance is known only from a handful of wells in north-eastern Poland (Majorowicz, 1976) and is well documented in wells north of Poland (Kukkonen et al., 1998). The most prominent case of disturbance due to paleoclimatic conditions is observed in a group of wells in north-eastern Poland in the low heat flow area with strong inversion of the temperature gradient with depth (Fig. 4 and Majorowicz, 1976). Application of functional space inversion (FSI) procedure (Shen and Beck, 1991) on the Udryń-4 log reveals surface changes of 8 °C (Fig. 5). This magnitude of change does not occur in many of the other Polish Lowland wells as shown for well Prabuty 1 (Figs. 4 and 5).

We also notice that heat flow as a function of depth does not show a statistically significant trend for the

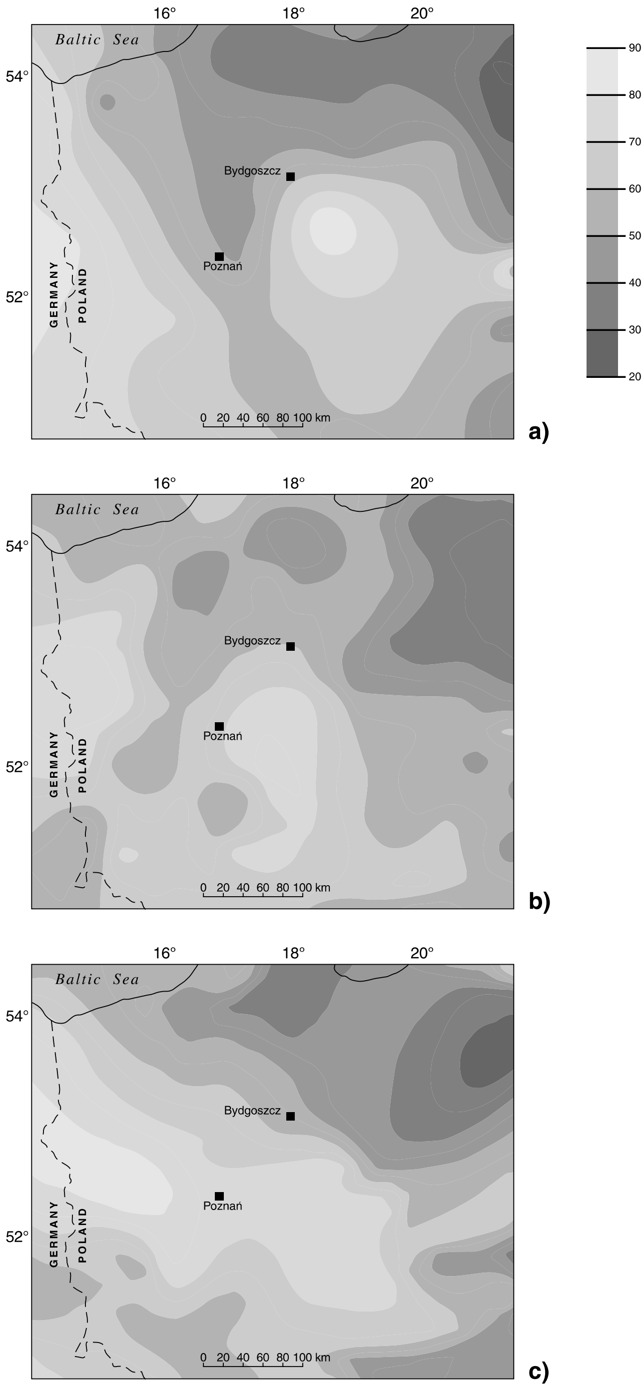

Fig. 2. Heat flow contour maps based on heat flow data compiled by: (a) Plewa (1994), (b) Gordienko and Zavgorodnaya (1996) and (c) Bruszewska (2001). Heat flow contours are in mW/m^2.

Paleozoic Platform, however, the six lowest heat flow values in north-eastern Poland (Precambrian Platform) are from shallow depth wells (<1.5 km), (Fig. 6). This may be a result of the climatic reduction in heat flow as described for the Udryń-4 case above.

We have constructed heat flow contour maps of the study areas based on data bases from Plewa (1994), Gordienko and Zavgorodnaya (1996) and Karwasiecka and Bruszewska (1997) (Fig. 2(a)–(c) respectively). All of these maps show variations between each other mostly within a range of the estimated error of heat flow determination (15–20%); We use heat flow values shown in Fig. 2(c) (based on the latest and largest Polish heat flow data base from Karwasiecka and Bruszewska, 1997) in our modelling. Karwasiecka and Bruszewska (1997) used corrected geothermal gradients for thermal equilibrium. Thermal conductivity values were measured for hundreds of cores using the divided-bar apparatus. However, these are point values and use of averages and net rock description improved knowledge of the conductivity variations with depth.

The analysis of the heat flow pattern across the area in which about 3000 km of new DSS profiles exist shows significant changes from low heat flow (<40 mW/m^2) in the north-eastern part of the TESZ to high values (60–80 mW/m^2) in the western and southern section of the TESZ (Fig. 3). The analysis of the main heat flow features (Figs. 1(b) and 2(c)) shows evidence of extensive crustal–mantle warming in the area between the Sudetes to the south and the TESZ to the north. The change in heat flow is more than 100% when compared with values on the craton and is well above the uncertainty of heat flow estimates (15–20%). Heat flow in the anomalous zone is 25% higher than in the Sudetes. The axis of the anomaly is aligned with the Dolsk Fault and Variscan deformation front. Large crustal/mantle temperatures seem to contradict to high P_n velocities (as high as 8.4 km/s) revealed by the POLONAISE'97 data. High heat flow (>80 mW/m^2) coincides with lower crustal thickness and the opposite is the case for the low heat flow zone (no major changes in elevation).

5. Crustal temperature modelling

Assumption of constant mantle heat flow of 20–30 mW/m^2 (25 mW/m^2) derived from previous models of heat generation in the crust based on the relationships between heat generation and seismic velocities for the older DSS profiles and the surface heat flow (Čermak et al., 1989) results in large lateral variations of crustal heat flow from as high as 65 mW/m^2 in south-western Poland to 5 mW/m^2 in north-eastern Poland (Fig. 7). The analysis of derived crustal heat flow variations shows little correlation between the thickness of the upper crustal high heat generating zones (sediments, meta-sediments and granitoids) and heat flow for the area east of the EEC. Profile LT-7 (Fig. 7) is the best example. The obvious discrepancy between high and low crustal heat flow between the EEC and the Paleozoic Platform is due to a large contrast in the surface heat flow. Allowing larger contrast in the mantle heat contribution (higher in the high heat flow areas) would solve

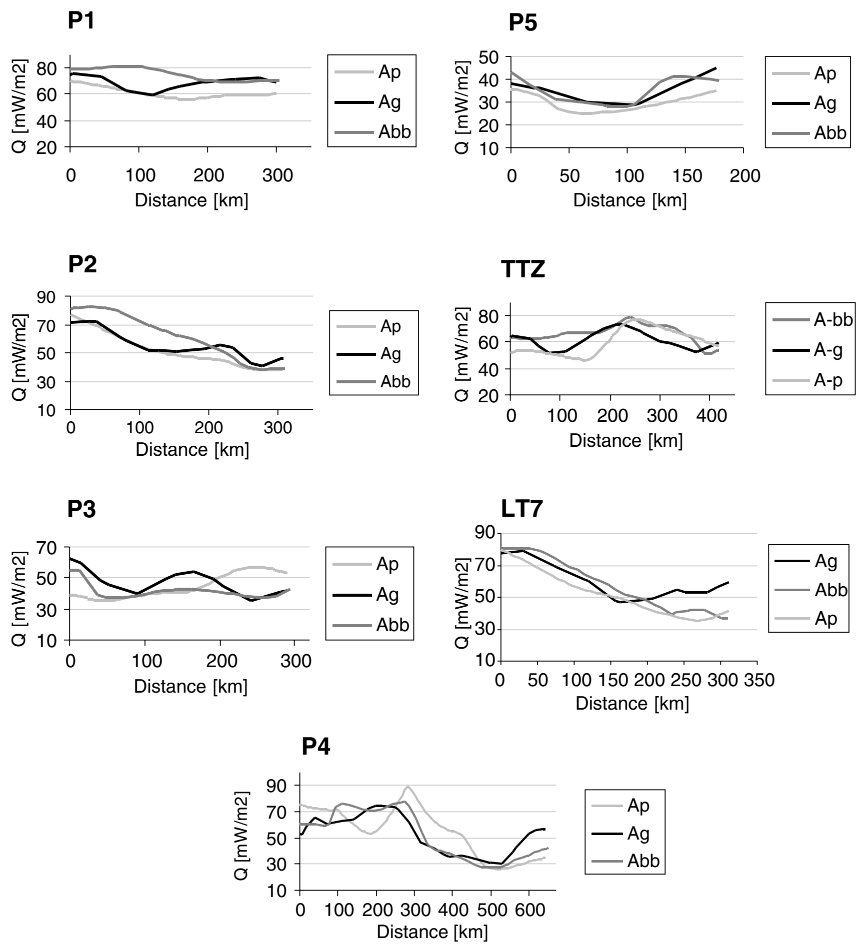

Fig. 3. Variations of heat flow along the new generation seismic profiles: P1, P2, P3, P4, LT-7 and TTZ. Ap, Ag, Abb are heat flow profiles according to Plewa (1994), Gordienko and Zavgorodnaya (1996) and Karwasiecka and Bruszewska (1997), respectively.

these discrepancies. Therefore, we applied 2D modelling to address mantle and crustal heat flow variability and its dependence on surface heat flow and heat generation variations along three profiles running through the characteristic low and high heat flow anomalies. We have chosen three profiles with seismic velocity models which characterise the high heat flow zone in the Variscan platform (P1), the cold area of the craton (P3) and, perpendicular to the main boundaries, the profile crossing through the Paleozoic platform, TTZ and the craton (Fig. 1(b)). The models are shown in Figs. 8–10.

Numerical 2D modelling has been carried out for profiles P1, P3 and LT-7 based on the numerical solution of the steady-state heat conduction equation in vertical geothermal models along the profiles. The calculated heat flow corresponds to the modelled heat flow resulting from the numerical solution of the heat conduction equation within the geothermal model described by heat production, thermal conductivity and mantle heat flow.

The length of each profile was 300 km, the depth 50 km. The surface temperature was fixed at 8 °C. The observed surface heat flow along the profiles (Karwasiecka and Bruszewska, 1997; Fig. 2) was used to optimise heat production within the model (profile P3) or to optimise the mantle heat flow (profiles P1 and LT-7), which was used as a boundary condition at the bottom. A constant mantle heat flow of 25 mW/m^2 was considered along profile P3 (Figs. 8–10). The optimised, variable mantle heat flow along profiles P1 and LT-7 was determined by a regularisation algorithm described by Stromeyer (1984) and Šafanda (1985). The algorithm seeks the mantle heat flow, which minimises the difference between the observed and modelled surface heat flows on a set of functions with a fixed upper limit of the horizontal variation in order to avoid unrealistic differences in the heat flux from the mantle. The same algorithm was used to optimise the original heat generation model (see the text below) in the uppermost 6 km of the crust in profile P3. The algorithm seeks an additional heat source $f(x)$ constant in the vertical direction, which, when added to the original heat production, minimises the differences between the observed and calculated surface heat flows.

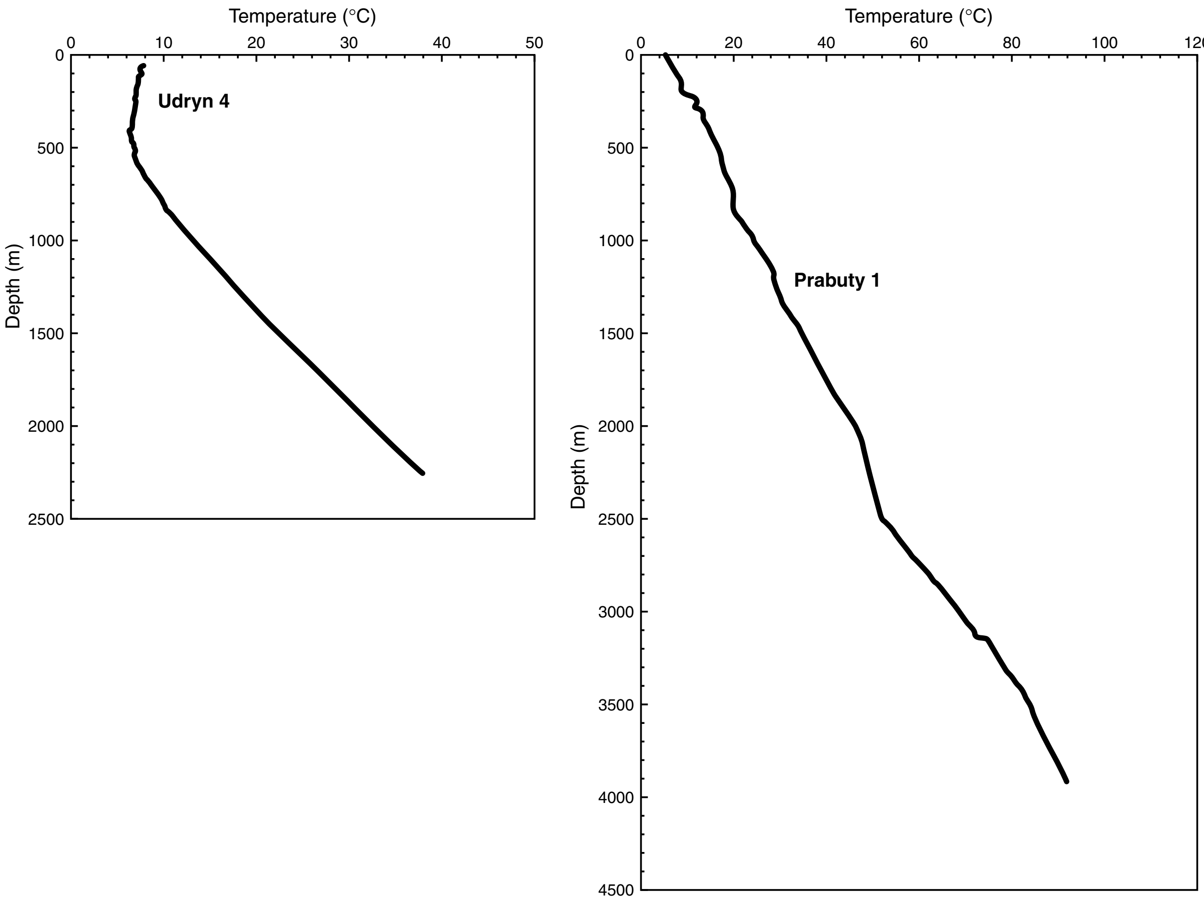

Fig. 4. Temperature profiles in Udryń-4 (northeastern corner of Poland) and Prabuty-1 (northern Poland) wells.

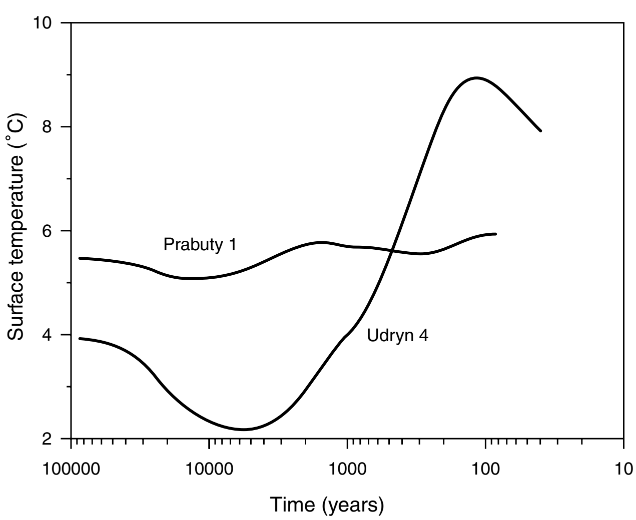

Fig. 5. Ground surface temperature history derived from FSI of the temperature profiles in well Udryń-4 and Prabuty-1. Note that the sharp change observed in Udryń-4 well is not observed in Prabuty-1 well.

The model of heat production is based on the empirical relationship between radiogenic heat production and compressional 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) distinguishing between the Phanerozoic and Precambrian geological units. The heat production of the uppermost sedimentary layer was not calculated from the heat generation–seismic velocity relationship. It was estimated at 1×10^{-6}–1.2×10^{-6} and 1×10^{-6}–2×10^{-6} W m^{-3} for the thick layers of metamorphosed sediments reaching down to 20 km. Heat generation of the mantle was estimated at 0.01×10^{-6} W m^{-3} (Precambrian) and 0.015×10^{-6} W m^{-3} (Phanerozoic). Heat production flow variations in the sedimentary layer influence significantly our estimate of the mantle heat flow and mantle temperature. Trials with lower heat production values than the ones above increase mantle heat flow and mantle temperatures unreasonably.

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

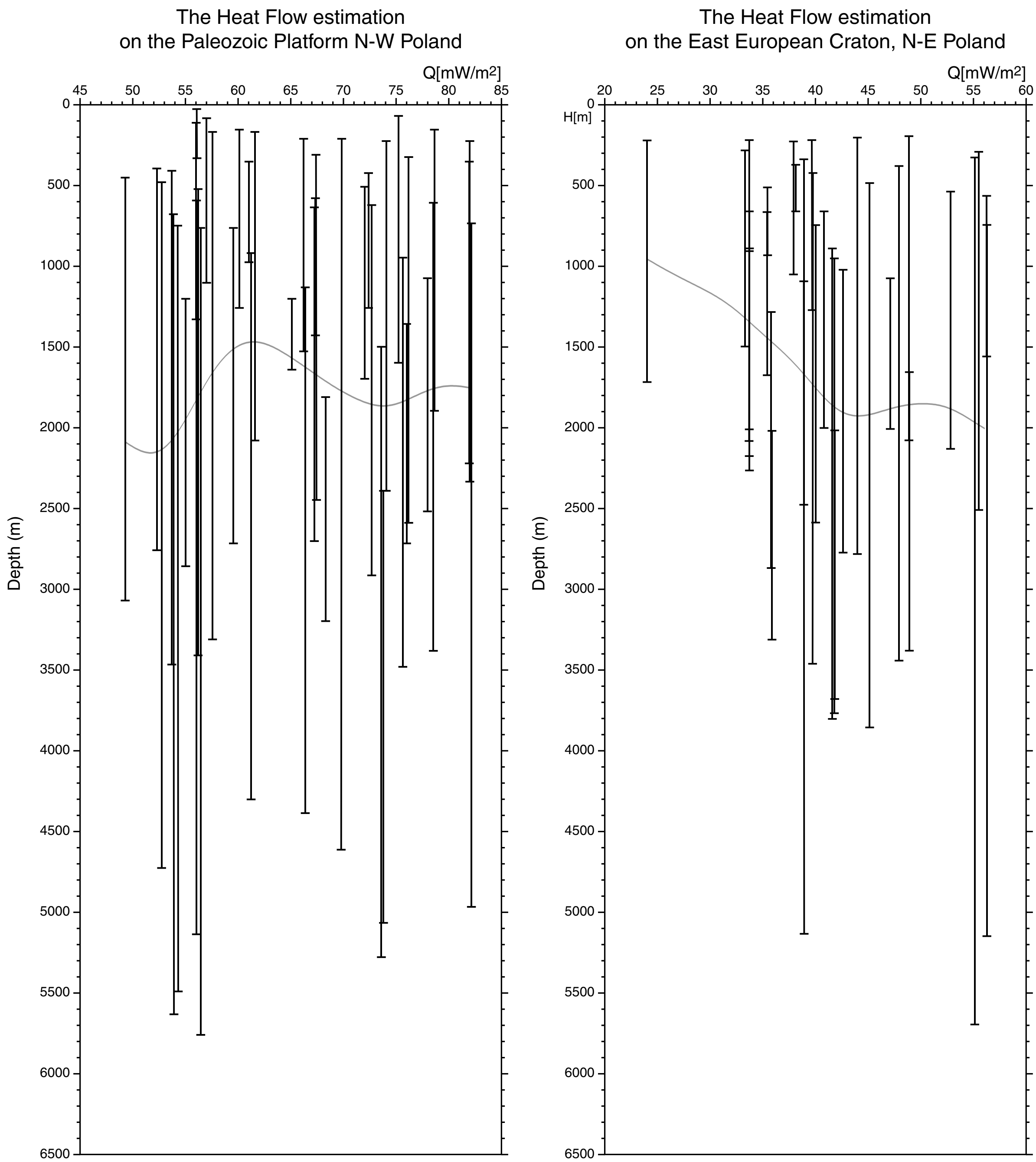

Fig. 6. Heat flow versus depth in Precambrian (EEC) and in the Paleozoic platforms.

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

where T is temperature in °C, z is depth, b and c are constants, and k_0 is the thermal 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 measured 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 are 2.2 $W\,m^{-1}\,K^{-1}$ for sediments, 3.0 $W\,m^{-1}\,K^{-1}$ for the upper (granitoids) and 2.6 $W\,m^{-1}\,K^{-1}$ for the middle and lower (basalts, gabbros) crust, respectively. The mantle in the models between the Mohorovicic discontinuity and the depth of 50 km, probably of peridotite composition, was assigned values $k_0 = 4$ $W\,m^{-1}\,K^{-1}$ and $b = 0$.

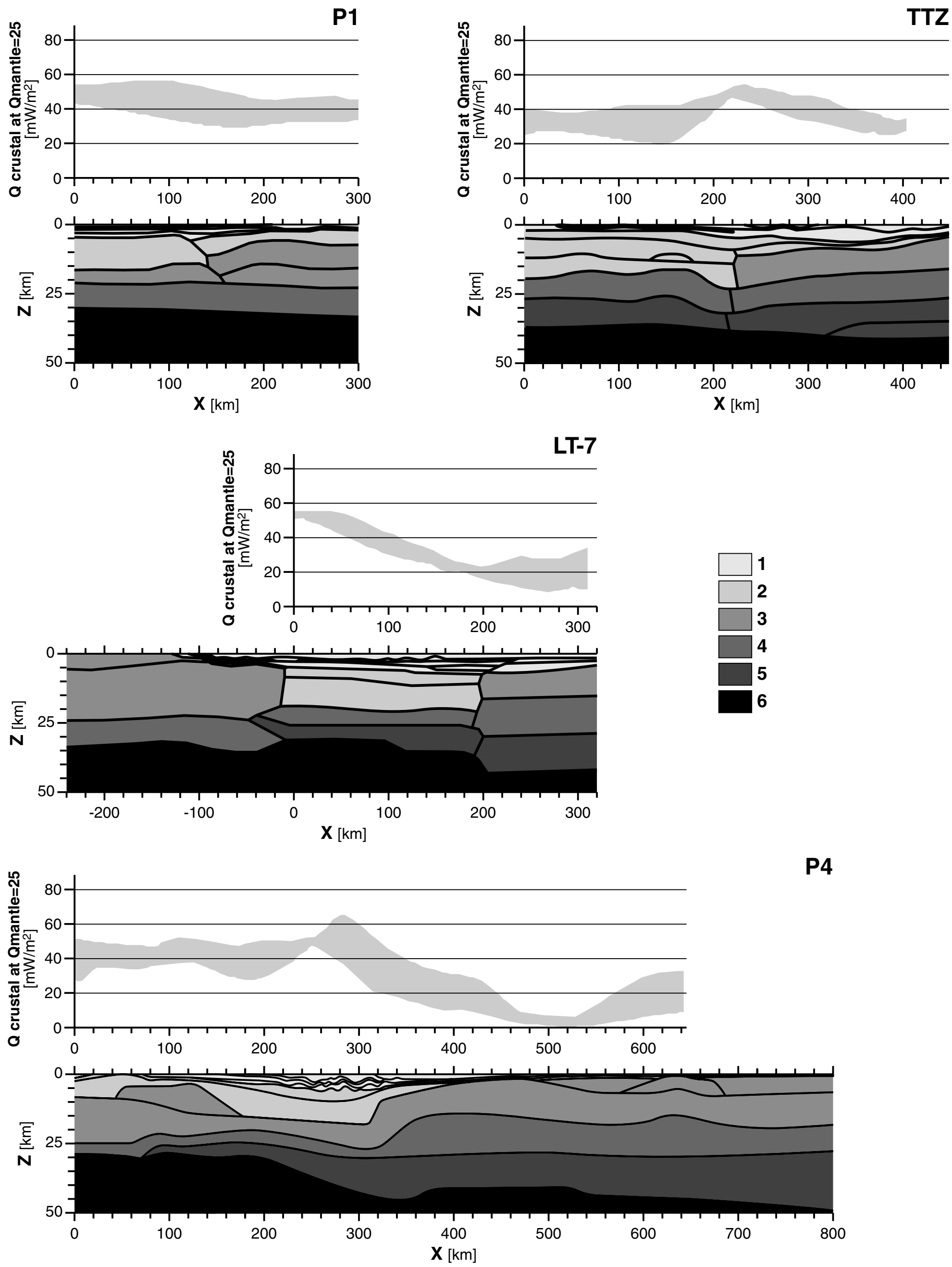

Fig. 7. Crustal heat flow at the surface for the LT-7, TTL and Polonaise profiles assuming mantle heat flow of 25 mW/m^2. Composition of the individual structural elements of the crust are from Krysiński et al. (2000): 1–sediments, 2–compacted sediments, 3–granitoids, 4–"basaltic" layer, 5–gabbroes, 6–subcrustal mantle (peridotites).

Due to the temperature-dependent thermal conductivity, the system of algebraic equations for temperature to be solved during the finite-difference solution of the heat conduction equation is non-linear and the numerical technique used for the temperature-independent model cannot be applied. Therefore, we used the approach described in Šafanda (1988), where the conductivity is adjusted to temperature by an iterative process. The output of the numerical models (temperature and heat flow) is shown in Figs. 8–10.

6. Discussion

The new results of modelling show that mantle heat flow increases after crossing the TTZ zone southward by 15–20 mW/m^2. The small-scale variations of mantle heat flow, e.g. along profile P1 (Fig. 8), are one possible cause of variations of the observed surface heat flow. Reliability of the observed surface heat flow is not high so small variations in mantle heat flow cannot be resolved. Surface heat flow variations can also be caused by

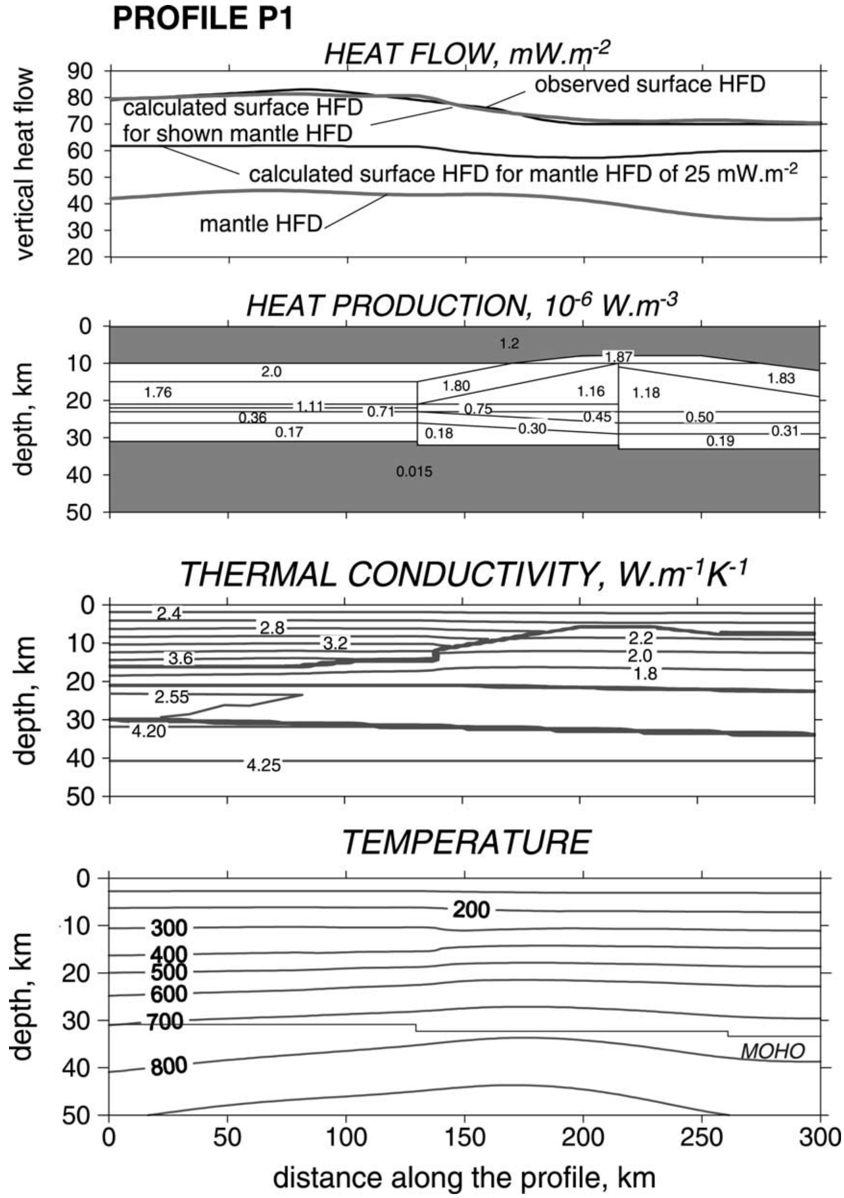

Fig. 8. 2D models of surface and optimised mantle heat flow densities, heat production (μW m^{-3}), temperature corrected thermal conductivity (W m^{-1} K^{-1}) and crustal temperature (°C) for the profile P1.

lateral changes of heat production or thermal conductivity. The variation of heat production in the sedimentary layer used to improve the fit between the observed and calculated surface heat flow in case of profile P3 (Fig. 9) shows that horizontal changes in heat production produce a good fit to surface data assuming a uniform mantle heat flow.

A contrast between the low to moderate temperature values in the TTZ zone and the EEC (450–550 °C) and the high temperatures at the Moho in the Paleozoic platform (600–750 °C), results from our modelling (Figs. 8–10). This temperature range is well within the bounds of 400 and 1000 °C (Baltic shield and Alpine–Mediterranean areas, respectively) from 3D modelling along the European Geotraverse (Čermak and Bodri, 1986). It is expected that Moho temperatures for the low heat flow zone associated with the Precambrian platform will be higher (by 100 °C) than for the older exposed shield because of heat sources in the sediments and the upper part of the Precambrian basement underneath the platform. The Moho temperature in the high heat flow zone typical of the Paleozoic platform (Variscan) is significantly lower (by 300 °C) than in the younger Alpine–Mediterranean system. Mantle heat flow values modelled for the Polish area (25–40 mW/m^2) are smaller than for the oldest and youngest tectonic areas along the European Geotraverse (20–60 mW/m^2). Our model temperatures and heat flow at Moho agree with models

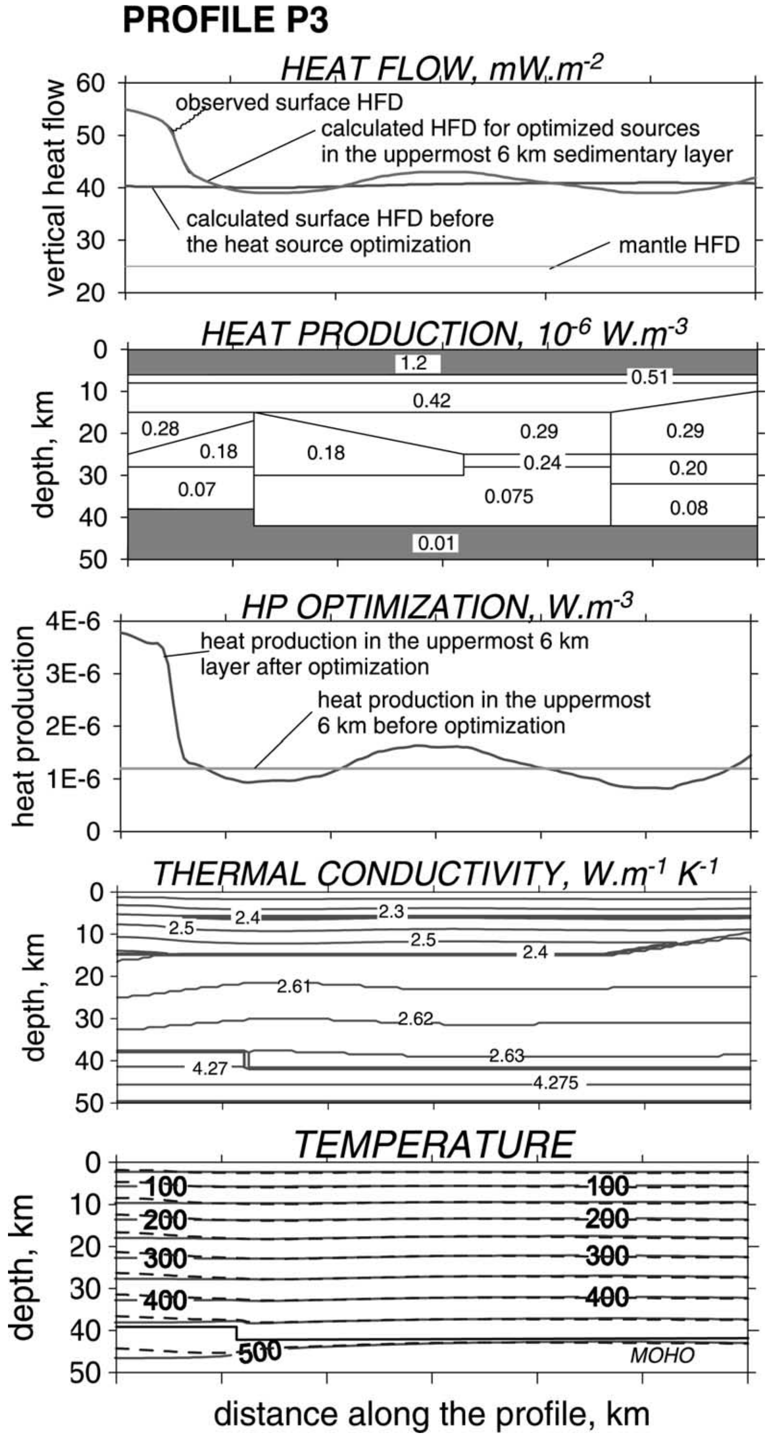

Fig. 9. Same as in Fig. 8 for the profile P3 for which mantle heat flow of 25 mW/m^2 is assumed (based on the modelling of the optimised mantle heat flow for the LT-7-old craton part–north-eastern part of the profile). Original heat generation model based on the heat generation-seismic velocity relationship was optimised in the uppermost 6 km (the middle panel).

across the Baltic Shield and the Tornquist zone north of Poland (Balling, 1995). Balling (1995) estimated that the mantle heat flow varies from 20 to 25 mW/m^2 along the northern and central shield transect increasing to 30–40 mW/m^2 beneath the basins and northern Tornquist zone. In Poland mantle heat flow along the TTZ is about 30 mW/m^2 increasing to 40 mW/m^2 in the area of the Variscan Deformation front. Föerster and Föerster (2000) estimate lower values of mantle heat flow (20–30 mW/m^2) for the Bohemian Massif (Variscan Erzgebrige in Germany), to the south west of Polish high heat flow zone in the Variscan Frontal area. These are less by

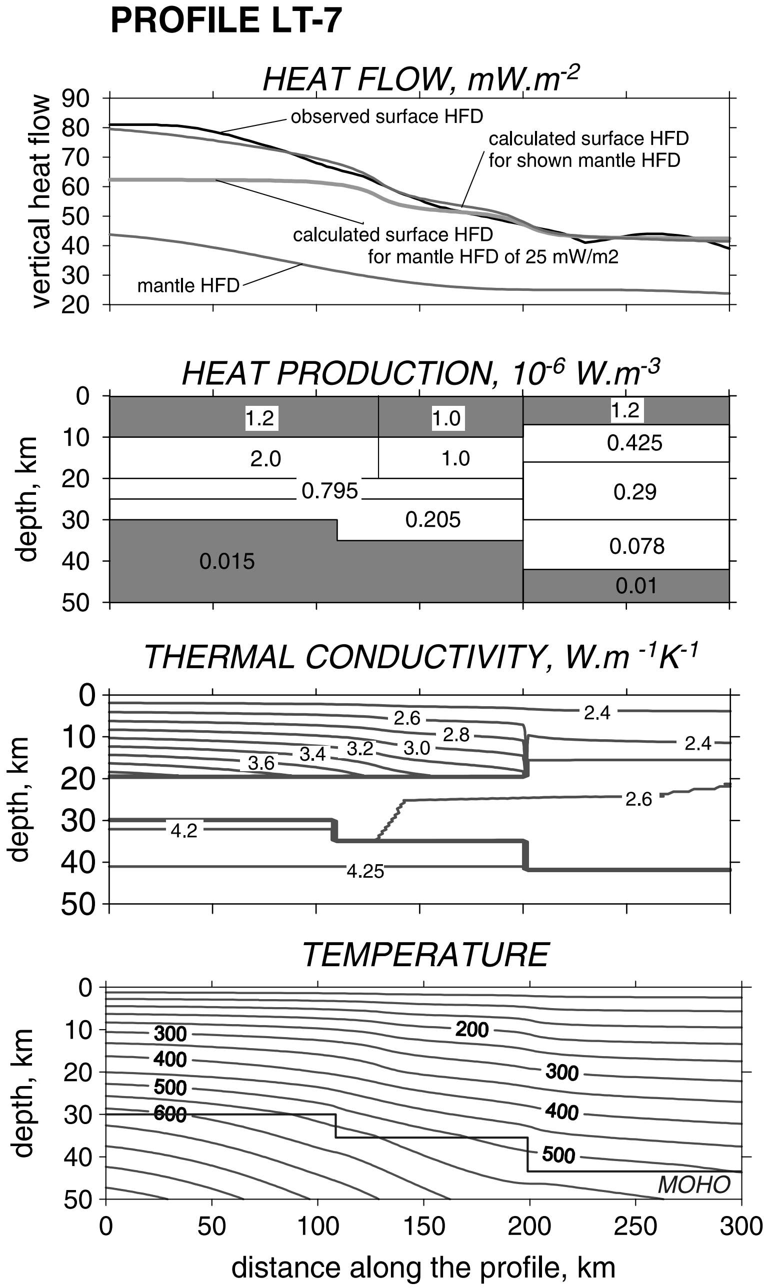

Fig. 10. Same as in Fig. 8 for the profile LT-7.

10–20 mW/m^2 than in Variscan area of Poland. Mantle heat flow of 20–30 mW/m^2 proposed by Föerster and Föerster (2000) agree better with TTZ values than with the Variscan Deformation Front areas of the Polish basin (this work) and German and Danish basins to the north (Balling, 1995). Lowering the mantle heat flow and temperatures at Moho to reach agreement with Föerster and Föerster (2000) requires much stronger heat flow sources in the thick low velocity layer (<6 km/s) interpreted as meta-sediments. Values as high as 3×10^{-6} W m^{-3} would be need for the meta-sediments. This however would result in unreasonably low mantle heat flow in the areas with lower heat flow and thick meta-sediments like in the TTZ. It is interpreted here that a significant difference in the mantle heat flow exists between the Bohemian Massif and the Variscan Deformation Front.

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 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 processes. These results are in good agreement with seismological information from surface wave dispersion analysis and the refraction/wide-angle seismic interpretation of FENNOLORA data (Balling, 1995). Geotherms calculated for the crust and upper mantle models along the profiles P1 and P3 (Variscan Platform and Precambrian craton respectively; Fig. 11) reveal similar variations of Moho and upper mantle temperature as those of Balling (1995). The depth of the thermal lithosphere (peridotite solidus) amounts to 100 km under the Variscan Platform, and is more than 200 km below the craton.

Relatively high temperatures at the Moho in the high heat flow area in the south-west Poland (600–750 °C) are not supported by high P_n seismic velocities at the top of the mantle (8.25–8.4 km/s; Jensen et al., 1999 and 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 (Sudetic Monocline) with 30–40 km thick crust and the areas with the 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 worldwide (Black and Braile, 1982) and for Europe (Kubik, 1986). This allows the use of P_n velocities as a proxy for deep temperatures. High P_n velocity of 8.1–8.4 km/s interpreted from seismic experiments for the Paleozoic and Precambrian platforms in Poland (Jensen et al., 1999; Guterch and Grad, 2000) result in a range of proxy temperatures between 600 and 350 °C respectively (Fig. 12). While the first number is in fairly close agreement with modelled Moho temperature in the area of the thick Precambrian crust, the second number is too low compared with modelling (600–750°) of the Paleozoic platform (Figs. 8 and 10), where P_n is 8.25–8.4 km/s (profile P1, Jensen et al., 1999). This is a factor of two times higher than proxy temperatures based on P_n velocities. High P_n velocities also suggest low heat flow in this area in disagreement with the observed high heat flow (Figs. 1(b) and 2).

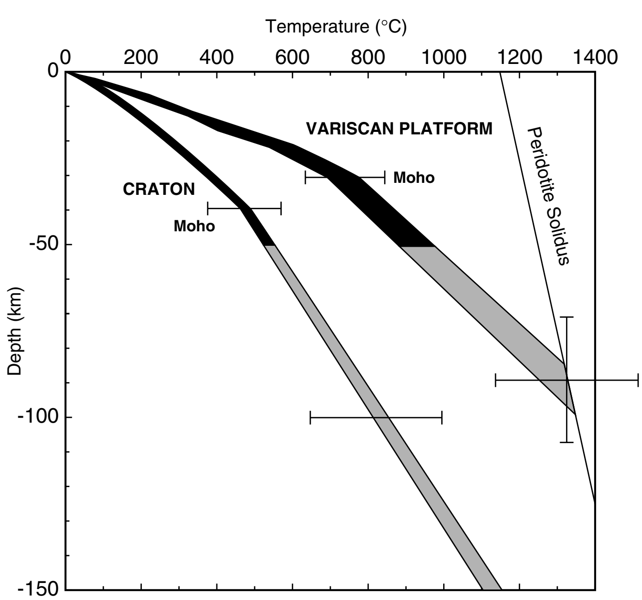

Fig. 11. Lithospheric geotherms for the profile P1 (Variscan foreland basin) and profile P3 (craton). Conservative estimates of the uncertainties in the temperature prediction are shown.

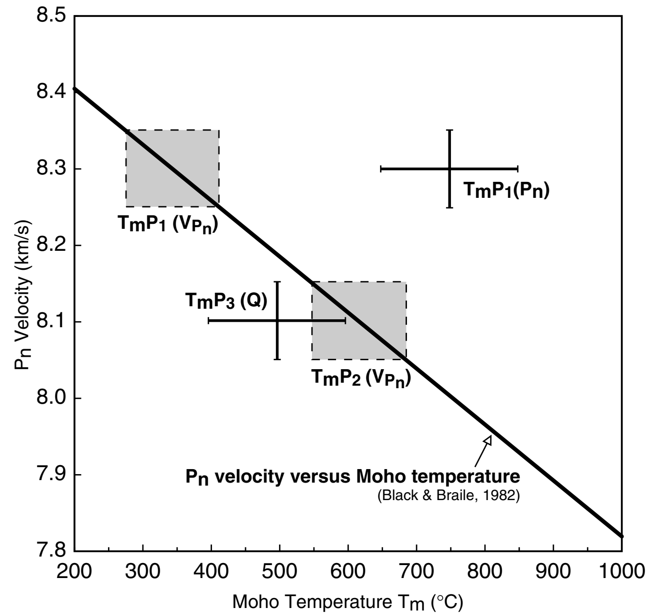

Fig. 12. Estimates of temperature from the relationship between P_n velocity and temperature at Moho depth (Black and Braile, 1982, Hyndman and Lewis, 1999).

The question can be raised if heat flow determined in the Polish basin (Paleozoic Platform) represents deep crustal heat flow below the highly porous sediments or is the result of upward hydrologic flow. Heat flow in the potential recharge zone of the high elevation area in the Sudetic Mountains and the Sudetic block is quite high (60 mW/m^2). Since heat flow in the potential discharge zone along the Middle Odra Fault is 80 mW/m^2 the reduction or increase of heat flow cannot be less than −10 mW/m^2 in the recharge area or more than +10 mW/m^2 in the discharge area. A difference of 10 mW/m^2 can be also explained by differences in heat generation between the exposed Sudetes and buried sediments on basinal crust. A difference of 20 mW/m^2 due to hydrogeology cannot explain high heat flow in the Paleozoic platform and a difference of almost 100% between this high heat flow and the low heat flow in the EEC (Figs. 1(b) and 2(c)). Analysis of heat flow variation with depth in wells of the potential discharge area does not show any significant variations, which could be caused by the regional hydrodynamic discharge. However, this applies only to the upper 4 km of sediments and there is no information from below. It is assumed that due to decreasing porosity and permeability this effect would be even less at greater depth.

Another explanation of high heat flow in the central part of TESZ in the Polish basin is salt tectonics (see Fig. 1(a)). However this affects mainly the southern part of the TTZ as shown by heat flow variation along the TTZ profile (Fig. 3). The northern zone of the Pomeranian block (Fig. 1(b)) is characterised by low heat flow, which likely expresses the deep undisturbed heat flow. As an example, heat flow in the well Debrzno Ig-1 located close to a salt pillow structure (Fig. 1(a)) is 54 mW/m^2 while heat flow in the well Wrzesnia Ig-1 (located far from salt structures, Fig. 1(a)) is 73 mW/m^2. Heat flow for eleven wells in the salt structures varies between 54–82 mW/m^2 (with mean of 69 ± 9 mW/m^2). Heat flow in 36 wells outside the structures varies between 49 and 82 mW/m^2 with a slightly lower mean of 65 ± 10 mW/m^2. Therefore salt tectonics does 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. Along the TTZ the northern low heat flow block is characteristic by low level magnetic anomalies east of the TTZ boundary not seen in the south (Wybraniec, 1999). Therefore the north to south increase of heat flow along the TTZ may indicate deeper changes of heat flow at the crystalline upper crustal levels caused by differences in heat production between these blocks.

Isostasy furnishes another indication of heat flow variations and the thermal state of the crust and upper mantle. Thick crust and low elevation are indicative of low temperatures (thick lithosphere). Thin crust and high elevation are indicative of high temperatures and thin lithosphere (Bodri and Bodri, 1985; Hyndman and Lewis, 1999). High heat flow and hot crust and mantle in the Paleozoic platform characterised by thin crust versus low heat flow in the areas of thick crust of the EEC and TTZ is confirmed by a first order model of crustal thickness variation required for isostatic balance as a function of heat flow (Bodri and Bodri, 1985 and Hyndman and Lewis, 1999). Small elevation changes observed along the low elevation area of LT-7 (<200 m) and high variations in crustal thickness from 30 to 45 km require heat flow variations between 35 and 75 mW/m^2 respectively. This is of the same magnitude as variations in the heat flow observed between profiles P1 and P3 (Paleozoic and Precambrian platforms, respectively) and along profile LT-7 (Figs. 2, 7 and 13).

We suggest that the proxy relationship between P_n velocity and temperature of the crust/mantle boundary as well as the relationship between heat flow and P_n distribution overestimates temperatures at the Moho and is not applicable in this case. Another possibility is

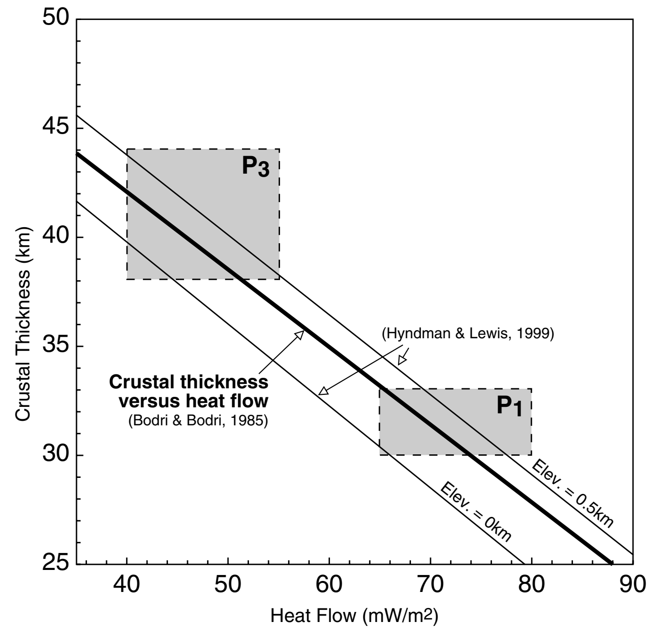

Fig. 13. Crustal thickness versus heat flow for P1 and P3 against the relationships by Bodri and Bodri (1985) and Hyndman and Lewis (1999).

that P_n velocities are much lower (~8 km/s): there is no evidence of that. There is evidence of mantle composition variation (Guterch and Grad, 2000). The profiles perpendicular to the strike of the TTZ zone show normal P_n velocity around 8.1–8.2 km/s while the profiles parallel to the strike show Moho velocities as high as 8.4 km/s. Other explanations would include decrease of heat flow with depth in the high heat flow zone due to the existence of a strong vertical hydrodynamic component of heat flow in porous sediments. This effect would diminish the discrepancy between our model of the thermal structure of the upper lithosphere based on observed heat flow and the one based on the interpretation of the high P_n velocities. The simple model of isostatic balance between high heat flow—low crustal thickness area and low heat flow—high crustal thickness area agrees with the observations and model of temperature structure of the crust and upper mantle as presented in this paper. It is also noticed that high contrast in heat flow between the old craton and the Palaeozoic platform existed in the past as showed by the analysis of vitrinite reflectance data (Majorowicz et al., 1984).

7. Conclusions

The analysis of heat flow data in the Polish basin and 2D numerical modelling of the crustal temperatures shows evidence of extensive crustal–mantle warming with the axis of the anomaly aligned with the Dolsk

Fault and Variscan deformation front. High heat flow above thin crust and low heat flow above thick crust with no major variation in elevation is supported by a simple isostatic balance model.

Crustal heat generation explains part of the high flow in the zone with thick meta-sediments reaching down to 20 km depth, however, it is far from explaining high heat flow in Variscan crust and in the transition zone into a cold craton. 2D numerical models of heat flow strongly based on new seismic data show that a contrast of 15 mW/m^2 in mantle heat flow is required. High mantle heat flow (35–40 mW/m^2) is likely to exist in a zone of high surface heat flow while cold crust and high-density mantle (mantle heat flow of 20–30 mW/m^2) is typical of the craton.

Thermal lithosphere thickness for the craton is 200 km while it is only 100 km in the accreted terranes to the south-west of the TTZ. The TTZ in Poland appears as a relatively cold area.

High modelled crustal/mantle temperatures in high heat flow zones are at odds with very low temperatures at Moho from the relationship between temperature and P_n velocities (more than 8.2 and as high as 8.4 km/s). High P_n velocities also suggest low heat flow in this area in disagreement with the observed high heat flow. Therefore we suggest that the proxy relationship between P_n velocity and temperature of the crustal/mantle boundary as well as the relationship between heat flow and P_n distribution overestimates temperatures at the Moho and it is not applicable here. Another possibility is that only a few km of the uppermost mantle is of high velocity and P_n velocities beneath are much lower (closer to 8 km/s) but there is no evidence of that.

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