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Boundary conditions in thermal models: An application to the KTB site, Germany

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RESUMEN

Las estimaciones de la temperatura a profundidad son de suma importancia para los estudios reológicos de la corteza y también para una planeación correcta de la perforación de pozos profundos. La modelación térmica requiere de condiciones de frontera reales, para así poder obtener valores confiables de la temperatura a profundidad. Las condiciones de frontera necesarias para los modelos térmicos pueden ser inferidas a partir de parámetros geoquímicos y geofísicos medidos en el campo.

Se pueden obtener soluciones numéricas de las ecuaciones de transferencia de calor a través de modelos de elementos finitos o de diferencias finitas para una, dos o tres dimensiones, tomando en cuenta varias suposiciones en relación con los parámetros del medio, tales como homogeneidad y anisotropía de la conductividad térmica. En este trabajo presentamos modelos uni- y bi-dimensionales y resaltamos las diferencias entre ambos.

El área de Oberpfalz fue seleccionada para probar el modelo y las condiciones de frontera debido a que los datos geoquímicos y geofísicos se encontraban disponibles para estimar las condiciones de frontera, ya que las temperaturas a profundidad habían sido medidas en el pozo KTB. Se encontró concordancia entre las temperaturas medidas y las calculadas para el modelo bidimensional que incluía condiciones de frontera inferidas a partir de los datos geoquímicos y geofísicos.

PALABRAS CLAVE: Modelación, conductividad térmica, ecuaciones de transferencia de calor, condiciones de frontera.

ABSTRACT

Deep temperature estimations are important for rheological studies of the crust and also for the planning of deep drill holes. Thermal modeling requires the input of realistic boundary conditions in order to obtain reliable values for the temperature at depth. Boundary conditions necessary for thermal models may be inferred from geochemical and geophysical parameters measured in the field.

Numerical solutions of heat equations can be obtained through finite element and finite difference schemes in one, two or three dimensions, taking into account several assumptions regarding the medium parameters, such as homogeneity and anisotropy of thermal conductivity. In this paper we present one and two-D models and highlight the differences between them.

The Oberpfalz area was selected to test the model and boundary conditions, because geochemical and geophysical data were available to estimate the boundary conditions in our model and temperatures at depth have been actually measured at the KTB borehole. A good agreement between the calculated and measured temperatures is obtained for a 2-D model with appropriate boundary conditions from geochemical and geophysical data.

KEY WORDS: Thermal conductivity, modeling, heat equations, boundary conditions.

Functions and parameters	**Notations and units**
Heat production	A [μW.m^{-3}]
Unitary vectors of the coordinate system	e_i [m] (x,z)
Domain dimension	L ,H [m]
Time	t, [s]
Temperature	T [°C]
Temperature at the reference level	T_r [°C]
Temperature difference	ΔT [°C]
Thermal conductivity	$K(T)$ [W.m^{-1}.°K^{-1}]
Heat flow density	$\phi = (\phi_h, \phi_v)$ [mW.m^{-2}]
Depth scale factor	D [km]
Nabla operator	$\nabla = \left(\frac{\partial}{\partial x}, \frac{\partial}{\partial y}\right)$

INTRODUCTION

The geothermal field reflects the long and complex geological evolution of the Earth. Heat flow at the Earth's surface provides valuable information about the thermal conditions and processes at depth. Methods of heat flow analysis and construction of geothermal models are complicated. Several assumptions are needed in order to solve the heat equation, especially those related with boundary conditions. Thermal models allow us to study complex processes such as asthenosphere uplift, magma intrusions, and accumulation of sediments. Other important features in thermal modeling are thermal parameters such as radiogenic heat production and heat conductivity of the medium. Geothermal models of the crust are also used in rheological and stress field calculations (Ranalli, 1991).

Drilling of deep wells requires approximate calculations of the expected temperature at depth. The Continental Deep Drilling Project is a good example of the importance of a reliable estimation of deep temperature (Clauser and Huenges, 1993). In this paper we present an alternative method of fixing the deep boundary condition for the differential heat equation. We include a numerical approach for the solution of the conductive heat equation in 2-D. Our results agree with deep temperature data from the Oberpfalz deep borehole. It is shown that the good agreement between measurement and modeling is due to appropriate selection of boundary conditions.

BOUNDARY CONDITIONS OF HEAT EQUATION SOLUTIONS

Why choose a two or a three-dimensional model? Why use convective or conductive models? Why are one-dimensional solutions of differential equations useful? There are no strict rules on how to solve the heat transport problem. However, the fundamental problem is not related with dimensionality but with the boundary conditions assumed in the solution of the differential equations. The relative contributions of heat transport mechanisms is another parameter to fix. These mechanisms are conduction, convection and radiation. Which is the most important for each problem? These processes are dependent of each other, and their relationships can be expressed by different mathematical functions of conductivity, time, fluid velocity and heat production.

In most cases, the parameters in the heat equation solution are assumed to have no lateral variation, and the thermal conductivity tensor is considered homogeneous. These assumptions do not make a large difference between one-two-or three-dimensional analysis, as the boundary conditions determine the behavior of the model. In problems involving tectonics, the Mohorovicic discontinuity plays an important role in the evaluation of the boundary conditions. However, some authors (e.g., Cermak, 1984) consider neither the Moho discontinuity to be an isothermal surface, nor the flux of heat from the upper mantle to be uniform. The parameters of the lower crust depend on the composition of the mafic rocks within it which affects estimation of the lower crust-mantle boundary (CMB) depth. Griffin and O'Reilly (1987) suggest that the seismic Moho may differ from the CMB depending on the composition of mafic rocks and the regional tectonic evolution. Thus, the seismic Moho may overestimate the depth of the CMB in high heat-flow areas (Klemperer, 1987).

The estimation of temperature at depth is based on the assumption of equilibrium among different components in the rocks and the fluids that interact with them. Geochemical and isotopic geothermometers have been found accurate for temperature calculations in geothermal fields and volcanic environments (Henley *et al.*, 1984). Chemical geothermometers are constrained to relatively shallow depths because they indicate the last equilibrium stage, which is usually attained not far from the surface. However, good estimations of surface heat flow have been obtained using the correlation between equilibrium of temperatures calculated with quartz geothermometers and surface heat flow (Swanberg and Morgan, 1979). Isotopic geothermometers re-equilibrate slowly, and are more useful to determine deep temperatures. Helium isotopic composition shows anomalous values in areas of recent tectonic activity (Mamiryn and Tolstikhin, 1984; Oxburg and O'Nions, 1987). An exponential correlation ($^3He/^4He = \exp(6\,\phi\,[HFU] - 5.3)$) has been proposed between heat flow and $^3He/^4He$ ratio (Polak *et al.*, 1979; Mamiryn and Tolstikhin, 1984). Indirect methods can be applied to estimate the Moho temperature on the basis of seismic velocities (Cermak, 1984).

METHODS IN THERMAL MODELING

The traditional method (Kappelmeyer and Haenel, 1974; Cermak 1979; Werner and Kahle, 1980; Weber and Vollbrecht, 1986) for obtaining an approximation of the geothermal model solves the Fourier heat equation in one dimension for steady state conductive conditions assuming isotropic physical properties of the underlying rocks. However, in recent years numerical models are used to solve the three-dimensional heat equation

$$\nabla(-\mathbf{K}\nabla T) = \mathrm{v}' \cdot \nabla T + \frac{\partial T}{\partial t} - A(z) \ . \tag{1}$$

When no convective processes are involved and a stationary transport is assumed the equation takes the form

$$div(-\mathbf{K}\ gradT) = A(z) \ . \tag{2}$$

Assuming constant temperature T_o at the surface, constant heat flow Φ_b at depth computed from the surface heat flow Φ_s and from the heat produced within the layer of thickness L we find

$$\Phi_b = \Phi_s - \int_0^L A(z)dz \ .$$

The heat flow and thermal conductivity at the surface are measured experimentally. However, the underlying rocks and their heat production, their conductivity and the heat flow at the base are unknown. These parameters must be inferred from other geophysical studies and from the geological environment.

An alternative method for solving this equation has been proposed by Cermak (1984). It consists in decomposing the medium into a series of homogeneous layers, assuming each of them to be homogeneous. Solving the Fourier heat equation we find the heat flow at the bottom of each layer. The bottom temperature and heat flow are taken as boundary conditions for the next layer; the errors are cumulative.

An improved estimation of heat flow at the bottom of a geothermal model has been proposed (Royer and Danis, 1988). Poisson's equation is solved in two dimensions assuming an initial heat flow solution at the bottom and a zero-flow condition across the two vertical lateral boundaries of the domain:

$$\nabla\left(-\mathbf{K}\ (T)\nabla T\right) = A(z) \quad . \tag{3}$$

In this model, the thermal conductivity strongly depends on the geological features. Although pressure has some effect on K, it may be neglected (Dubois *et al.*, 1995). Thus

$$K(T) = \frac{K_o}{(1+\alpha T)} \ , \tag{4}$$

where K_o is the thermal conductivity at 25°C, and α usually ranges from 5 x 10^{-4} °K^{-1} to 10^{-3} °K^{-1} depending on the rock.

Typical heat production values are found in the literature (Table 1). We assume an exponential decrease of heat production with depth z in km, according to

$$A(z) = A_o \exp\,(-z/D) \ , \tag{5}$$

where the depth-scale factor D for radioactive enrichment varies for each region (Chapman, 1986; Cermak *et al*., 1991). This parameter can be assumed to lie between 8 and 16 km, depending on the area (Weber and Vollbrecht, 1986; Royer and Danis 1988; Werner and Kahle, 1980).

The temperature at the surface is the mean regional temperature in the specific region. The zero heat flow condition on lateral boundaries is supported by the absence of major conductive heterogeneities in the geological facies. The mantle heat flow condition has been calculated by iteratively solving the heat equation, and comparing the calculated and the measured heat flow at the surface, interpolated at each node of the grid by a krigging method. As the mantle heat flow boundary condition depends on the interpolation method used, no unique solution is available for equation (2). A Monte Carlo simulation method has been used in order to obtain a confidence interval for the calculated heat flow field. This technique was used for the southern Rhinegraben, and a good agreement was obtained between the modeling results and available geoscientific data (Royer and Danis, 1988).

Another approach considers advection processes in porous layers. The equations governing the advection of incompressible fluids in a porous domain are the heat transfer equation, the Darcy equation (motion equation), the conservation equation the variation of the fluid characteristics with temperature. Royer and Flores (1994) provide a dimensionless formulation that include anisotropy and heterogeneity. We use dimensionless parameters that are functions of the conventional dimensional parameters, and functions of heat transfer and Darcy equations. Thus, the problem of modeling advection in porous media is simplified as a set of two

Table 1

Parameters from literature used in thermal modelling

Ref. *	Facies	D Km	Conductivity K_0 W/m°K	Heat Production A_0 μW/m^3
1,4	Fanglomerate	-	1.3 - 2.2	0.5-1.0
1,4	Sandstone	-	2.20 - 2.46	0.5-1.0
2,4	Amphibolite	-	2.46 - 2.9	1.0
2,4	Schist	8	2.1 - 2.44	1.0
2,4	Metabasite	8,16	1.00 - 2.53	1.0
2, 1,4	Granulite	-	2.50 - 2.65	2-4
1, 3,4	Granite	8,10,12	2.90 - 4.60	5.0-7.5
2,3,4	Mantle	-	2.50 - 3.00	0.02

References: [1]Weber and Vollbrecht (1986); [2]Chapman (1986); [3]Royer and Danis (1988); [4]Kappelmeyer and Haenel (1974).

differential equations (heat transfer and Darcy), that include two unknown functions (temperature T and stream ψ). The conservation equation is automatically verified using the stream function formulation.

The solution of the coupled conduction-advection heat transfer is performed with a finite-differences scheme, using a doubly iterative approach that allows mutual coupling of the heat and Darcy equations (Flores, 1992). An iterative procedure is applied to the flip-flop process until the solution converges. This formulation can take into account the heterogeneous medium, the conductive structure of the basement, the anisotropy of the petrophysical properties and the geometry of the geological layers.

The preceding models were applied to Oberpfalz because this area has been well researched in recent years. Geophysical and geochemical studies and deep temperature measurements are available from the Deep Drilling Continental Program of Germany (KTB reports). The temperature data can be compared with the temperatures obtained from different kinds of models.

THE OBERPFALZ TEST AREA

Oberpfalz is the pilot zone for the German Deep Drilling Project (KTB, Deutsche Kontinentale Tiefbohrprogramme), whose purpose is to understand the processes, dynamics and evolution of the continental crust (Figure 1). Extensive geophysical and geological studies started in 1986 (Weber and Vollbrecht, 1986). In 1987, a 4000 m deep pilot hole (VB) was drilled accompanied by large-scale data acquisition. The main hole (HB) reached its final depth of 9101 m in 1994 (Kohl and Rybach, 1996). Initially the data obtained from thermal conductivity, magnetic, magnetotelluric and seismic reflection studies showed low heat flow. The early geothermal models at Oberpfalz showed similar patterns to those reported by Cermak (1984) for the northern and eastern European region. Based on those results, the drilling site was selected at Oberpfalz. In 1987 the KTB project started to obtain important geoscientific data through well logging. The logging program was thoroughly revised. One of the objectives was to gain insight into the extent of fluid movement in crustal processes. The temperatures encountered during drilling of KTB-HB reached 172°C at depths of 6024 m (Zoth, 1993) while the predicted temperatures from thermal models had reached only 75°C at depths of 4000 m (Weber and Vollbrecht, 1986)). This questioned the validity of thermal modeling for the prediction of temperatures at depth. Recently 3-D models were developed by different authors (Clauser and Huenges, 1993; Clauser and Mareschal, 1995; Jobmann and Clauser, 1994; Kohl and Rybach, 1994 and 1996, among others), and the results were consistent with measurements. Here, we attempt to predict thermal patterns using a simple 2-D conductive model similar to those proposed by Royer and Danis (1988). We take into account different interpretations of the field studies to include realistic boundary conditions. Two models were used, one considering conduction only and the other assuming advection in porous formations (Flores, 1992).

GEOLOGICAL SETTING

The continental Deep Drilling Site KTB is located near Erbendorf-Vohenstraus, Oberpfalz, Germany (Figure 2). Outcrops in the area are predominantly granites 424 Ma old (Quadt and Gebauer, 1993), and metamorphic rocks. The tectonic evolution of central Europe has been described by Weber and Vollbrecht (1986). The compressive regime corresponds to the Variscan thrust tectonics followed by the post-Variscan granite emplacement and Mesozoic faulting. Cenozoic tectonic processes have resulted in extensional features in the form of several continental rifts: Rhine and Eger, and the Pannonian Basin, accompanied by the eruption of Tertiary to Quaternary volcanoes. The surface and mantle heat flow values and the crustal thickness distribution on those features show the effects of the rifting processes, including higher than average heat flow values (Haenel, 1971). A thinner crust is associated with those rifts (Cermak *et al*., 1991).

During the pre-site evaluation, a correlation of stratigraphy and tectonic history was used together with geophysical data to infer the temperature field at depth. Relatively low temperatures were supposed to prevail at depths greater than 7000 meters (Weber and Vollbrecht, 1986). Presently, more geophysical and geochemical data are available showing that the Eger rift in the vicinity of the KTB site is an indication of recent tectonic activity in Oberpfalz, that affects the thermal regime of the area (Trappe *et al*., 1990). Chemical studies of the gases dissolved in mineral springs and ground water samples show that an important magmatic component is included in the crustal fluids from the area (Griesshaber *et al*., 1992; Weinlich *et al*., 1993). Thus the regional mantle heat flow should be considerably higher that the northern European values.

The Eger rift is characterized by extensional tectonics. It includes a graben-type structure with a NE-SW direction. Surface heat flow ranges from 70 to more than 90 mW/m^2 (Cermak, 1979; Haenel, 1983), above the world average of approximately 60 mW/m^2. An increased mantle input is also inferred from the studies of noble gases in ground water, rocks and mineral springs (Griesshaber *et al*., 1992; Weinlich *et al*., 1993; Bach *et al*., 1999). This evidence correlates with the data obtained for samples from the KTB (Schäfer and Kirsten, 1993), thus showing that the extensional tectonic regime of the Eger rift affects the KTB site area. The KTB site is located about 40 km southwest of the Eger rift.

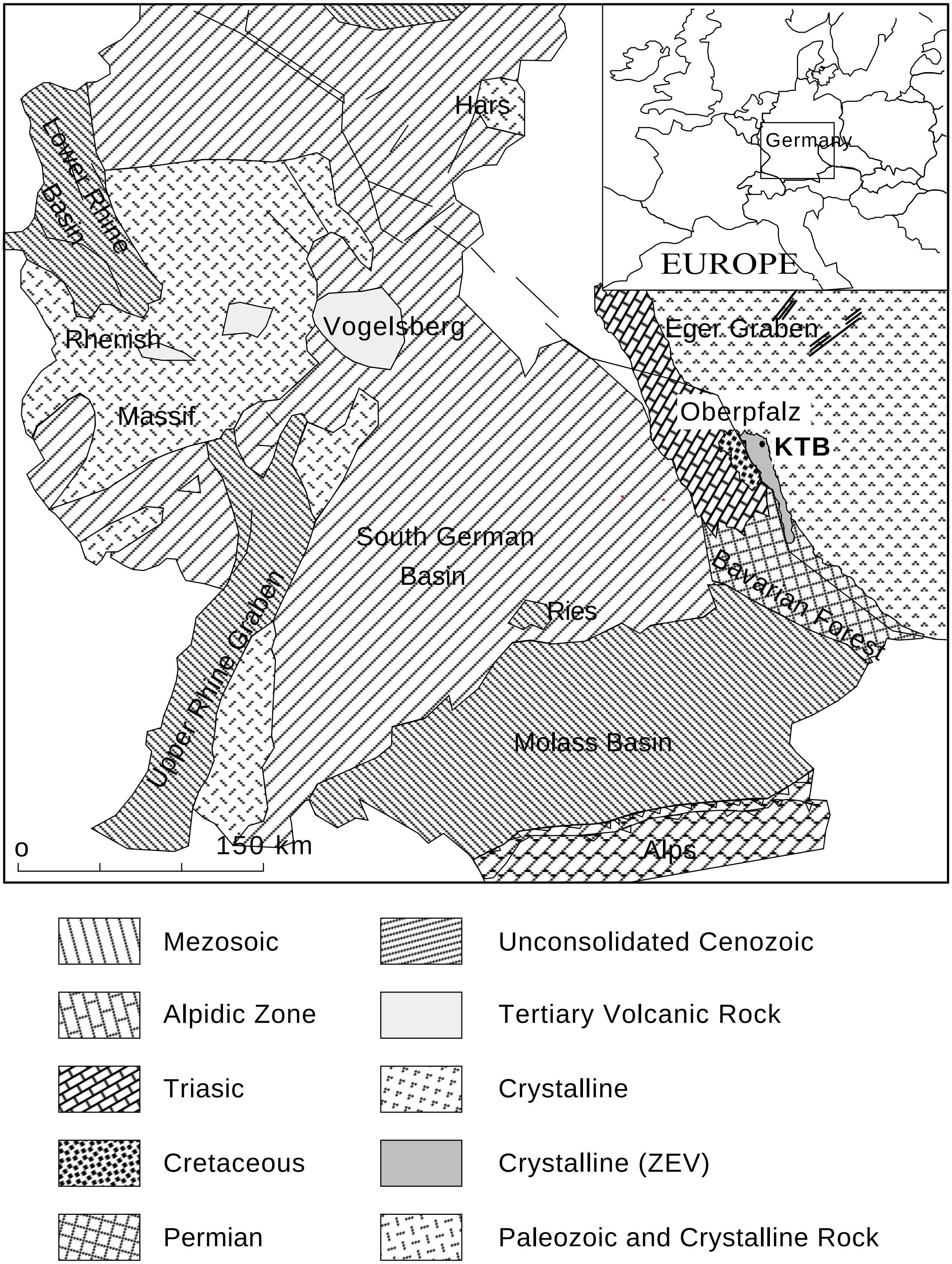

Fig. 1. Location and surficial geology in the Oberpfalz region.

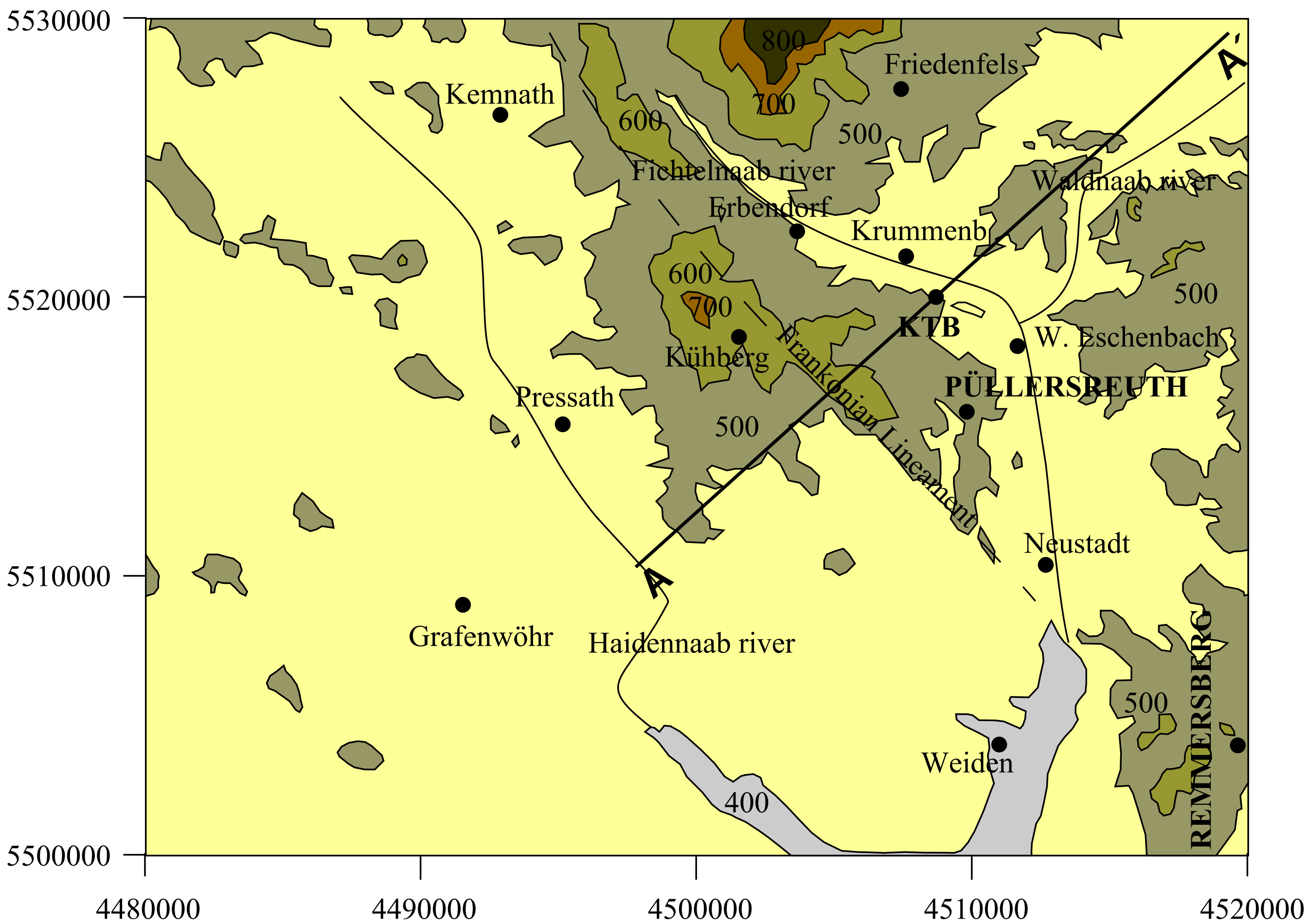

Fig. 2. Detail of surficial geology and location of the two-dimensional thermal profile (modified from Clauser and Huenges, 1993).

Geochemical and isotopic analyses of thermal springs in the Eger Rift near the KTB site, and data from KTB rock samples, indicate values of up to 6 for the corrected ratio (R/R_a) for helium isotopes (Schäfer and Kirsten, 1993; Griesshaber *et al*., 1992; Weinlich *et al*., 1993). According to available data the estimated surface heat flow for the area comprising the Eger Rift and the KTB site should be at least 40 mW/m^2.

Indirect methods to estimate the Moho temperature from seismic velocities (Cermak, 1984) yield an approximate temperature of 550°C for the KTB site (Cermak and Bodri, 1993), in close agreement with the actual thermal regime.

The estimated heat flow (>40 mW/m^2) and Moho temperature (approximately 550°C) are used as input parameters when modeling the temperature field at depth in the KTB site area.

2-D MODELING WITH TWO DIFFERENT BOTTOM BOUNDARY CONDITIONS

Equation (2) was solved numerically for the Oberpfalz zone using finite elements and finite differences in one and two dimensions (Figure 3).

In the first case we use 39 x 63 rectangular cells of 0.5 by 1 km. Surface temperature was taken as the mean annual temperature of the region (5°C); and the thickness parameter for radiogenic heat production D was assumed to be 8 km. Thermal parameters used in the model were those in Table 2. The solution was obtained by comparing iteratively the heat flow measured at the surface with the calculated heat flow. An iterative algorithm based on a Monte Carlo method modifies the heat flow at the bottom until the solution converges. This method yields also an estimate for the confidence interval of the temperature and heat flow fields.

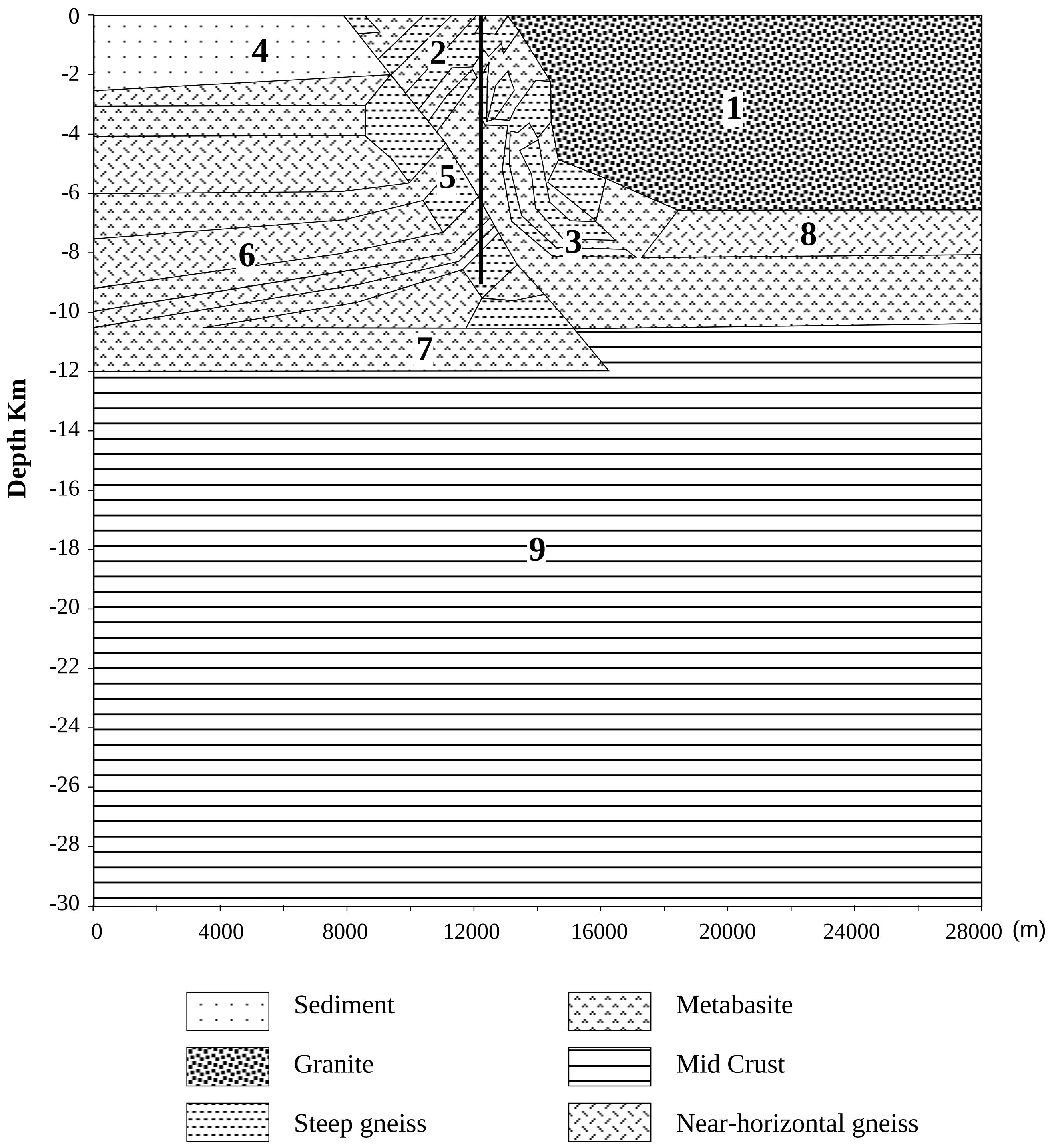

Fig. 3. Profile used for the 2D numerical model of heat transport using several boundary conditions (modified from Kohl and Rybach, 1996).

An initial constant heat flow up to 20 mW/m² was assumed at the bottom of the model (Figure 2b).

In the second case, a constant heat flow (18mW/m²) was assumed at the bottom of the section (Figure 4), on the assumption that the surface heat flow is similar to the heat flow calculated using the helium-isotope ratio (Mamiryn and Tolstikhin, 1984) and the surface heat flow measurements published for the surrounding area (Haenel, 1971; Cermak, 1984), which yields a minimum of 40 mW/m² at the surface. If the ratio of heat flow to He flux from the mantle is the same as in ocean basins, the He isotopic composition of base-

Table 2

Parameters used in the Oberpfalz thermal model

Ref. *	Facies	D km	Thickness	Conductivity K_0 W/m°K	Heat Production A_0 μW/m[3]
5	Fanglomerate	-	500	1.85	0.7
5	Sandstone	-	1300	2.20	0.7
2,4	Amphibolite	-	5200	2.46	1.0
2,4	Schist	8	6500	2.44	0.5
2,4	Metabasite	8	1500	2.53	1.0
1, 2, 4	Granulite	8	3000	2.65	0.2
1, 3,4	Granite	8	1200	2.44	5.0
2, 3,4	Mantle	-		2.50	0.02

* References: [1]Weber and Vollbrecht (1986); [2]Chapman (1986); [3]Royer and Danis (1988); [4]Kappelmeyer and Haenel (1974); [5]Burkhardt, Honarmand and Wägerle (1988).

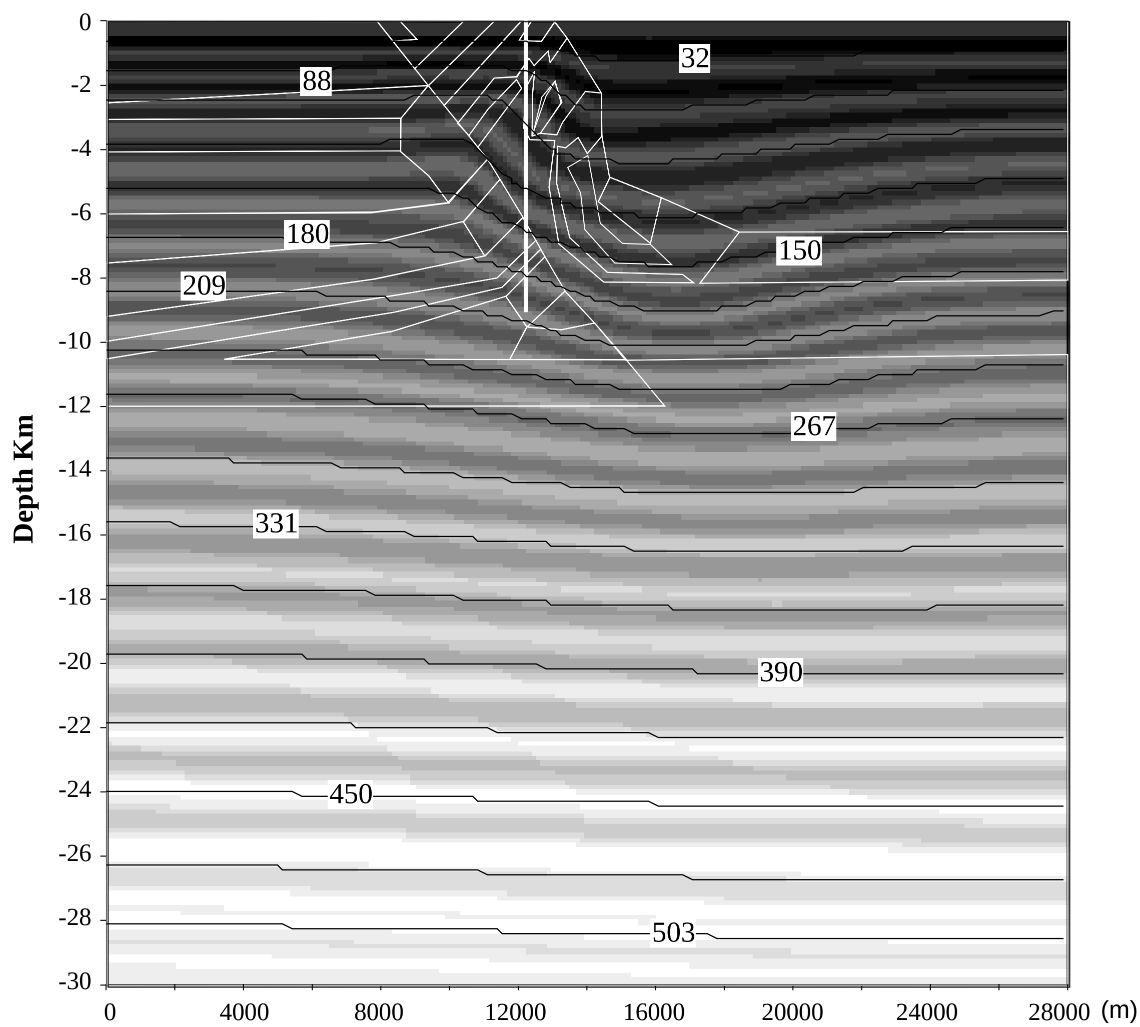

Fig. 4. Field temperatures obtained by the two-dimensional model of steady state conductive heat transfer (see also Tables 2 and 3).

ment fluids in the KTB boreholes is consistent with the inferred reduced heat flow at the KTB site (Wolfgang *et al.*, 1999). Calculation of the heat flow at the base includes the heat production data shown in Table 2.

In the third case, a constant temperature for the Moho was computed from the depth-temperature analysis proposed by Royer and Danis (1988). This proposes a linear relation between temperature and depth at Moho interface: $T_{Moho} = 12.7z+216+b$, with b=-91 and z in km. The depth is known, then the temperature at the Moho is estimated from the depth. In the Oberpfalz a minimum estimated temperature of 550°C seems consistent with the seismic data (Cermak and Bodri, 1993).

As a fourth case we consider a two-dimensional model (Figure 3), with a constant heat flow at the bottom as in the second case, and advection following Flores (1992). The grid is 60 by 60 cells, each cell being 0.5 x 0.5 km. Hydraulic parameters were taken from Jobmann and Clauser (1994) as in Table 3. The results are summarized in Figures 5 and 6.

RESULTS

The 1-D solution using fixed surface temperature and isotropic thermal conductivity (Table 2) is shown in Figure 6. This curve is nearly the same as the one published prior to the drilling results (Weber and Vollbrecht 1986). The temperature estimated at 4000m depth by the 1-D model is 83°C, using T_S=5°C and Φ_S=40 mWm^{-2}. This value is below the temperatures encountered during drilling, of 105° and 170°C at depths of 4000 and 6024 m, respectively (Zoth, 1993). The 1-D model yields a temperature of 120°C at a depth of 6000 m from indirect estimation of heat flow and temperature at depth. This temperature is closer but still lower than the actual measurements.

The 2-D model discussed in previous sections (Figure 6) provides a better approximation of the temperature (155°C and 180°C at 6024m depth respectively). This model shows larger temperature differences at depths greater than 25 km, especially near the Moho. At 10 km depth, the 2-D model estimates a temperature range from 270°C to 290°C, which is higher than the 171°C calculated with the 1-D model.

The last case considers an advective process (Figure 5). It shows a better fit to the observed temperatures, as shown in Figure 6. This figure shows the balance between temperatures resulted from the models specified in the previous section in one and two dimensions, including constant gradient, which is nearly the same as the equilibrium temperature reported by Zoth (1993).

DISCUSSION AND CONCLUSIONS

We show that the assumptions normally made for the Moho are acceptable in the case of the Oberpfalz and the Rhine-graben (Royer and Danis, 1988), despite the fact that the nature of the Moho is still controversial (Wolfgang *et al.*, 1999). Some thermal assumptions for the Moho and surface heat flow estimations can yield good results in calculating the temperatures at depth.

The low estimates of heat flow at the surface are due to the fact that we imposed a mantle heat flow value of 18 mW/

Table 3

Parameters used in the advective – conductive thermal model

Layer	Conductivity W/m°C	Permeability m^2	Heat Production mW/m^3
(1) Granite	3.5-3.7	1.10^{-15}-1.10^{-17}	6.0
(2) Sediments	2.2-3.0	1.10^{-15}-1.10^{-18}	1.2
(3) Steep gneiss	2.5-2.8	1.10^{-15}-1.10^{-17}	1.2
(4) Sediments Keuper	2.2-2.9	1.10^{-15}-1.10^{-18}	0.8
(5) Metabasite & horizontal Gneiss	2.9-3.3	1.10^{-15}-1.10^{-17}	3.0
(6) Near-horizontal gneiss	3.0-3.3	1.10^{-16}- 1.10^{-18}	1.5
(7) Gneiss & metabasite	2.7-3.3	1 – $5. 10^{-18}$	0.8
(8) Horizontal gneiss & metabasite	3.0-3.5	1 - 4.10^{-17}	3.0
(9) Mid crustal	3.3-3.4	1.10^{-25}	0.6-1.0

Taken from Kohland Rybach (1996) and Jobmann and Clauser (1994).

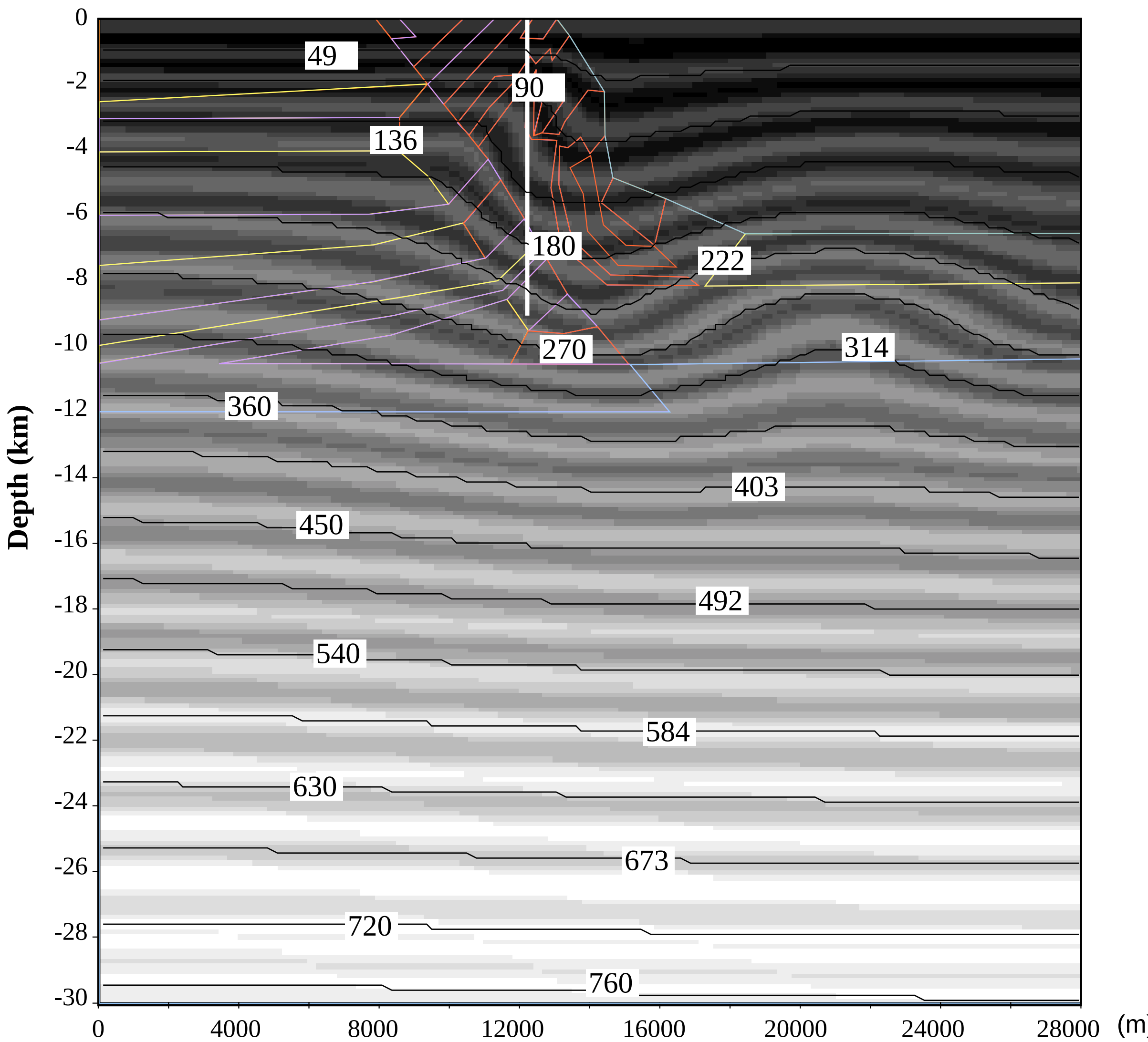

Fig. 5. Field temperatures obtained by the two-dimensional model of steady state advective-conductive heat transfer (see also Table 3).

m^2 in the first case. In the second case, the temperature imposed at the Moho produces the same effect at the surface. The depth-scale parameter D for radioactive heat production is 8 km for both cases, which is the minimum estimated value (Lachenbruch, 1970). Should this value be increased, the estimated surface heat flow would be much higher.

Temperatures obtained from 2-D modeling of the geothermal field are close to those encountered during drilling. If we assume a larger heat flow than inferred from geoscientific data, the calculated temperatures would be much higher than those measured in the borehole.

The use of 1-D models to determine deep temperatures is not validated from our results, because of the inherent restrictions of the boundary conditions in its solution. The 1-D model allows only one boundary condition on the independent variable, namely surface temperature. This yields erroneous values for deep temperatures, in our case underestimating the temperature at depth.

Griesshaber *et al.* (1992) found a $^3He/^4He$ ratio of 10^{-6} at Oberpfalz which predicts a surface heat flow of at least 40mW/m^2 according to Polak *et al.* (1979) and Mamiryn and Tolstikhin (1984). This value is in good agreement with the heat flow measurements (Haenel, 1971; Cermak, 1984). It was used to obtain a heat flow value at the approximate Moho depth, and it was also used in our 2-D model, yielding temperatures of 500°C to 598°C at 30 km, in agreement with the estimated temperature from seismic data (Cermak and Bodri, 1993). This model provides temperatures for shallow depths closer to the measured temperatures in the KTB borehole. Thus, when the model takes advective processes into account,

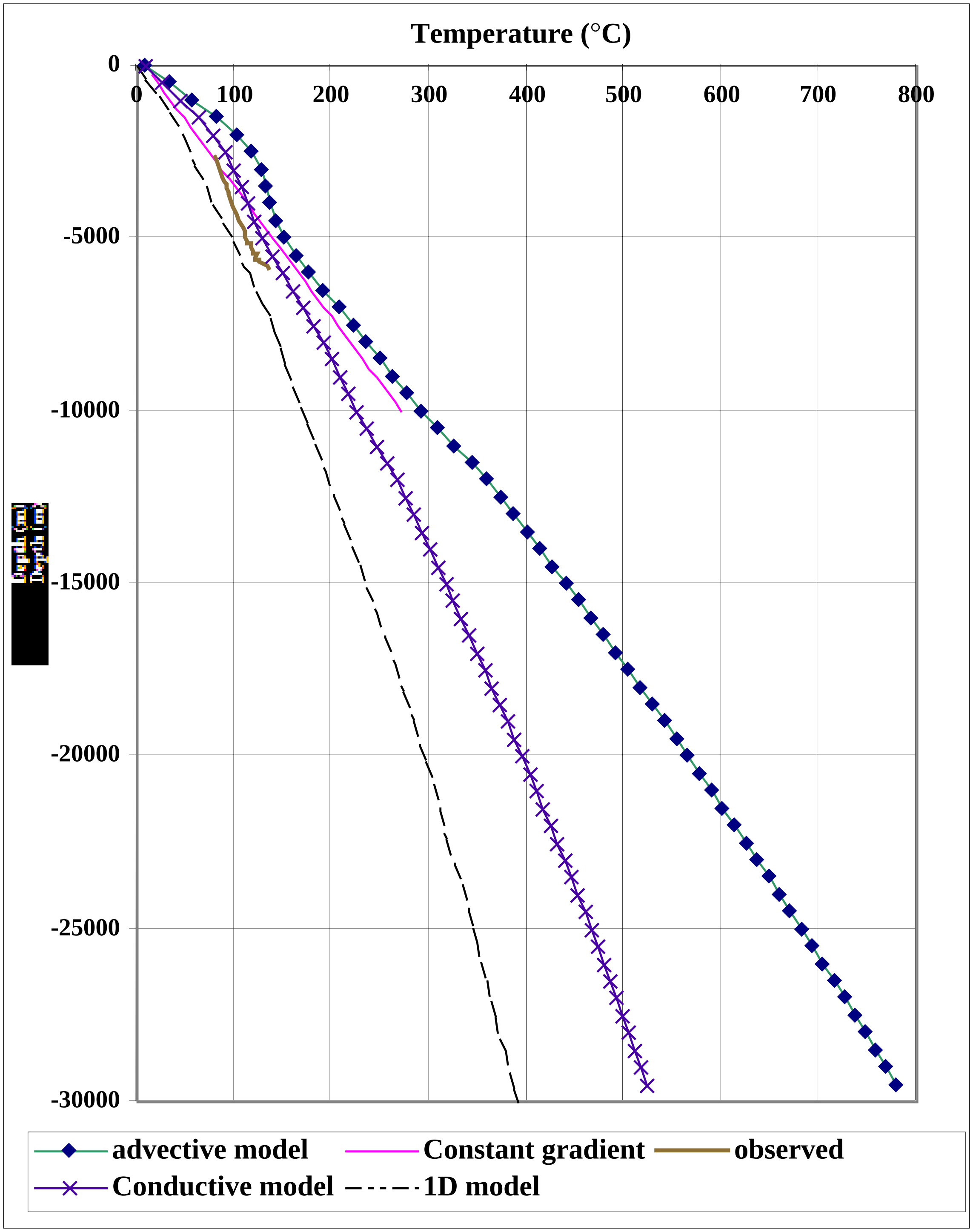

Fig. 6. Comparison between observed temperatures and model temperatures resulting from the proposed models in this work.

the computed temperatures agree with observations, but are slightly overestimated when compared with equilibrium temperature. However, temperatures at the Moho are clearly higher than those inferred by seismic data. In conclusion, our results show that thermal modeling can yield reliable temperature values at depth when the input parameters are inferred from geoscientific data.

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