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## Numerical model of heat flow in back-arc regions

Martin Kukačka\*, Ctirad Matyska

Department of Geophysics, Faculty of Mathematics and Physics, Charles University, V Holešovičkách 2, 180 00 Praha 8, Czech Republic

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## ABSTRACT

We numerically modeled mantle wedge flow driven by subducting plates and its consequences to surface heat flow in back-arc regions, where an increase in surface heat flow has been observed in almost all subduction zones. In order to calculate the steady state temperature and velocity field in a generic subduction zone model we utilized a new technique which does not require a priori assumptions about the position of the slab–wedge boundary. We demonstrated that calculated surface heat flow is very sensitive to the thermal boundary condition prescribed at the base of the model for a simple temperature-dependent viscosity. Since an increase of surface heat flow is a common feature of many back-arc regions, it should be rather independent on various thermal conditions applied at the bottom. We showed that such a robust behavior can be reproduced if, moreover, a strong pressure-dependence of viscosity is taken into account. The pattern of increased surface heat flow is then stable for a wide range of settings including various dip angles and age of the subducting slab. The temperature- and pressure-dependent viscosity resulted in a nearly isothermal mantle wedge, where temperatures exceeded 1200 °C even in relatively shallow depths, and, consequently, elevated surface heat flow is consistent with recent hot back-arcs evidence. Circulation in the wedge was obtained since the employed viscosity law resulted in creation of a zone of substantially reduced viscosity in shallow depths below the continental crust.

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## 1. Introduction

Subduction process is one of the most studied manifestations of the plate tectonics, however, it has not been fully understood yet. There are lots of data that may constrain subduction models but different data are of different importance and credibility.

First, seismic tomography provides evidence of the dip angle and the depth of slab penetration into the Earth's mantle together with the distribution of seismic wave velocities and attenuation factor  $Q$ . Tomographic inversions have clearly indicated that the decrease in seismic velocities in overlying mantle wedges is up to 6%, which may be caused by high temperatures, hydration state or both (Currie and Hyndman, 2006). It has also been shown that most mantle wedges exhibit a high  $Q$  factor in fore-arc regions and a low  $Q$  factor beneath the volcanic front and in back-arc regions (Schurr et al., 2003; Stachnik et al., 2004). This indicates that unlike the fore-arc, which is cold, the back-arc region is hot. The observed decrease in  $Q$  factor corresponds to a decrease in viscosity by about two orders, whereas the observed decrease in P-wave velocities indicates a viscosity weakening by at least three or four orders (Karato, 2003). The difference in the viscosity decrease evaluated from seismic velocities

and  $Q$  factor is usually explained by the fact that the seismic tomography of  $Q$  factor has lower spatial resolution.

Another important quantity is the surface heat flow which is used to constrain the temperature field in the subduction zone and particularly in the overlying mantle wedge. Measurements showed that surface heat flow is low in the fore-arc regions and high in the back-arc regions, which is, in general, in a good agreement with the  $Q$  factor distribution. The typical values of surface heat flow are about 30–40 mW m<sup>−2</sup> in fore-arc regions, more than 100 mW m<sup>−2</sup> in the volcanic front, and about 70–80 mW m<sup>−2</sup> in back-arc regions (Hyndman and Lewis, 1999), see also Fig. 1. Elevated surface heat flow is a common feature of most, if not all, subduction zones even in the cases where no recent extension is observed (Hyndman et al., 2005). Hot back-arcs also do not seem to depend on the age of the subducting lithosphere: zones of increased heat flow have been reported in the case of young subducting lithosphere (e.g. Cascadia, Ryukyu) as well as in the case of much older subduction zones (Java, NE Japan) (Currie and Hyndman, 2006). The extent of anomalously high surface heat flow zone may vary but usually is hundreds of kilometers. The transition between the cold fore-arc and the hot back-arc is sharp; a considerable increase in surface heat flow is observed in a narrow zone up to 50 km in width (Bostock et al., 2002).

A good example of this common pattern is the well studied Cascadia subduction, where the zone of high surface heat flow extends up to 550 km. The anomalously high surface heat flow is well noticeable here because the continental lithosphere adjacent to the

\* Corresponding author.

E-mail addresses: [Martin.Kuckacka@mff.cuni.cz](mailto:Martin.Kuckacka@mff.cuni.cz) (M. Kukačka), [Ctirad.Matyska@mff.cuni.cz](mailto:Ctirad.Matyska@mff.cuni.cz) (C. Matyska).

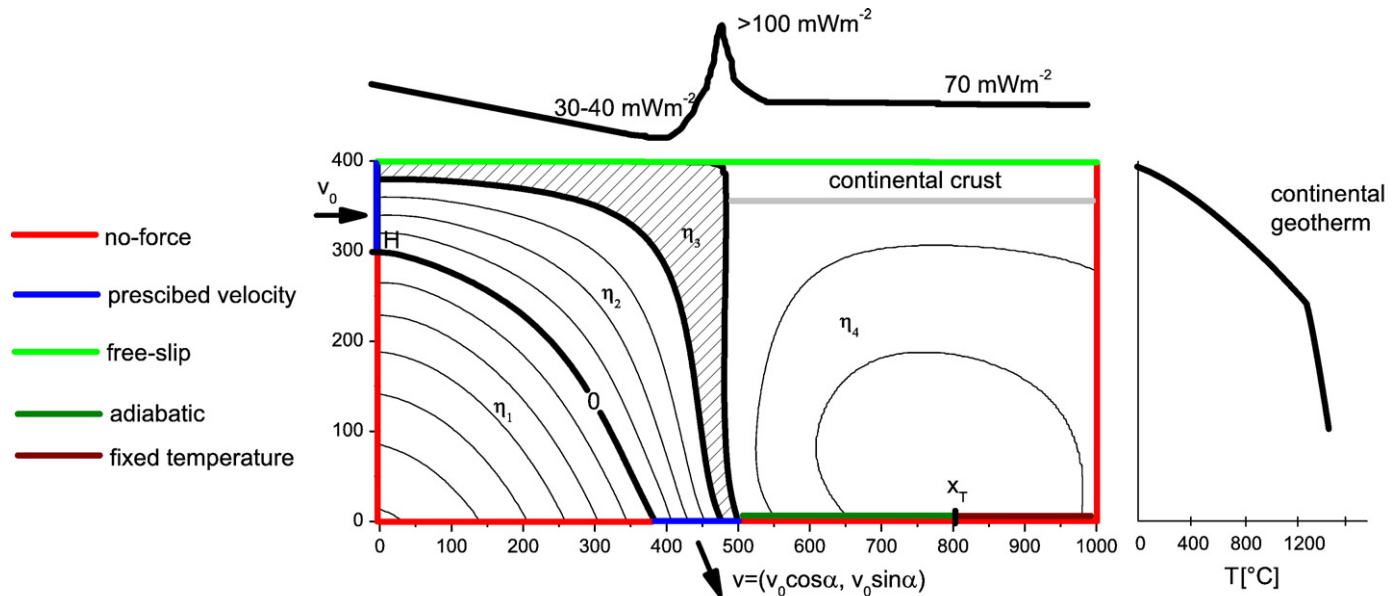


**Fig. 1.** Scheme of the computational rectangle domain  $1000 \times 400$  km with boundary conditions on velocity showing streamlines in the case of constant viscosity in the whole domain. The value of the stream function is scaled by the value  $\psi_s = H v_0$ . Streamlines labeled as 0 and 1 thus border the subducting plate. The weakening zone given by  $w=0.2$  is hatched and domains with different viscosity prefactors defined by Eq. (10) are indicated. Typical surface heat flow (based on Hyndman et al. (2005)) is schematically shown together with the cold continental geotherm prescribed at the landward boundary up to the depth of 300 km. The transition point  $x_T$  at the base of the back-arc, where the thermal boundary condition changes from adiabatic gradient to prescribed temperature, is marked.

Cascades is a cold craton, where average measured surface heat flow as low as  $42 \text{ mW m}^{-2}$  has been measured (Rudnick et al., 1998). Actually, a slight increase in surface heat flow with a decreasing distance from the cold craton has been reported, however, it has been attributed to the radiogenic heat production (Hyndman and Lewis, 1999).

Another independent evidence of hot back-arcs comes from xenoliths. In some back-arcs, xenoliths showing temperatures about  $900^\circ\text{C}$  at depths around 40 km and exceeding  $1200^\circ\text{C}$  at depths around 60 km were reported. These temperatures inferred from back-arc xenoliths are much higher than temperatures inferred from xenoliths in the cold cratonic lithosphere (Currie and Hyndman, 2006).

Finally, rheological properties of mantle materials have been studied extensively during the past decades. Various creep regimes may occur in the mantle wedge but it is believed that the most dominant are diffusion and dislocation creep (Karato and Wu, 1993). These can be generally described by the equation

$$\dot{\epsilon} = A \sigma^n d^{-m} \exp\left(-\frac{E + pV}{RT}\right), \quad (1)$$

see Table 1 for the explanation of the notation. This generalized form includes both Newtonian and non-Newtonian behavior. However, it has recently been shown that also the Peierls creep can play an important role, namely in the structure of cold fore-arc regions (Kneller et al., 2007). Generally, it seems that experimental data on creep properties do not give any strict constraints on the subduction-wedge models because the range of possible creep mechanisms is quite wide. Nevertheless, realistic rheology laws are necessary in numerical models, which should serve as a tool for studies of thermal state of subduction zones.

There are two kinds of numerical models used in modeling the mantle wedge: dynamic and kinematic. It has been shown that under certain conditions, the kinematic models, in which the velocity of the subducting plate is prescribed, give similar results as fully dynamic models (Han and Gurnis, 1999). Consequently, the number of kinematic models has increased in recent years (Peacock et al., 2005; Manea et al., 2005). The kinematic models can predict the temperature field in the subduction and both the temperature and the velocity fields in the mantle wedge, but need a contact between the

subducting slab and the overriding lithosphere to be prescribed a priori. Most of these kinematic models were primarily focused on the temperature of the slab-wedge boundary and therefore did not include a large portion of back-arc regions (van Keken et al., 2002) and thus the effect of the hot back-arcs has been usually overlooked. The study of a slab-driven flow in the overlying wedge, which focused on the elevated heat flow in the Cascadia back-arc (Currie et al., 2004), is the exception. However, it came to the conclusion that the traction from the subduction plate is not sufficient to produce upward flow in the wedge that would explain the heat flow pattern observed in back-

**Table 1**  
Notations and used values

| Symbol           | Meaning                                              | Value              | Dimension                         |
|------------------|------------------------------------------------------|--------------------|-----------------------------------|
| $A$              | Viscosity prefactor                                  | –                  | $\text{Pa}^{-n} \text{s}^{-1}$    |
| $\alpha$         | Thermal expansivity                                  | $3 \times 10^{-5}$ | $\text{K}^{-1}$                   |
| $c_p$            | Specific heat                                        | 1250               | $\text{J kg}^{-1} \text{K}^{-1}$  |
| $D$              | Deviatoric stress tensor                             | –                  | Pa                                |
| $\delta$         | Dip angle                                            | –                  | –                                 |
| $d$              | Grain size                                           | –                  | m                                 |
| $\Delta T$       | Difference of top and bottom temperatures            | 1450               | K                                 |
| $\Delta p$       | Lithostatic pressure at the bottom                   | $16 \times 10^9$   | Pa s                              |
| $\dot{\epsilon}$ | Second invariant of strainrate tensor                | –                  | $\text{s}^{-1}$                   |
| $\eta$           | Dynamic viscosity                                    | –                  | Pa s                              |
| $E$              | Rheological activation energy                        | –                  | $\text{J mol}^{-1}$               |
| $H$              | Thickness of subducting plate                        | –                  | m                                 |
| $g$              | Acceleration of gravity                              | 9.8                | $\text{ms}^{-2}$                  |
| $k$              | Thermal conductivity                                 | $3.1^a$            | $\text{W m}^{-1} \text{K}^{-1}$   |
| $K_T$            | Parameter scaling temperature-dependence sensitivity | –                  | –                                 |
| $K_p$            | Parameter scaling pressure-dependence sensitivity    | –                  | –                                 |
| $p$              | Pressure                                             | –                  | Pa                                |
| $R$              | Gas constant                                         | 8.314              | $\text{J K}^{-1} \text{mol}^{-1}$ |
| $\rho$           | Density                                              | 3300               | $\text{kg m}^{-3}$                |
| $\sigma$         | Second invariant of deviatoric stress tensor         | –                  | Pa                                |
| $T$              | Absolute temperature                                 | –                  | K                                 |
| $v$              | Velocity                                             | –                  | $\text{ms}^{-1}$                  |
| $V$              | Activation volume                                    | –                  | $\text{m}^3 \text{mol}^{-1}$      |
| $w$              | Weak fraction of subducting slab                     | –                  | –                                 |
| $:$              | total scalar product $x:y = x_{ij}y_{ij}$            | –                  | –                                 |

<sup>a</sup> 2.5 in the continental crust.

arc regions. The dynamic models also do not reproduce elevated heat flow in long distances from the trench (Eberle et al., 2002) and also need to prescribe a contact between the subducting lithosphere and the overlying wedge as the input parameter (Čížková et al., 2007).

In this study, we used a method, in which the position of the contact boundary between the subducting slab and the overlying wedge is not prescribed a priori (Kukačka and Matyska, 2004). Instead, the position of the plate-wedge interface is a part of the solution. The

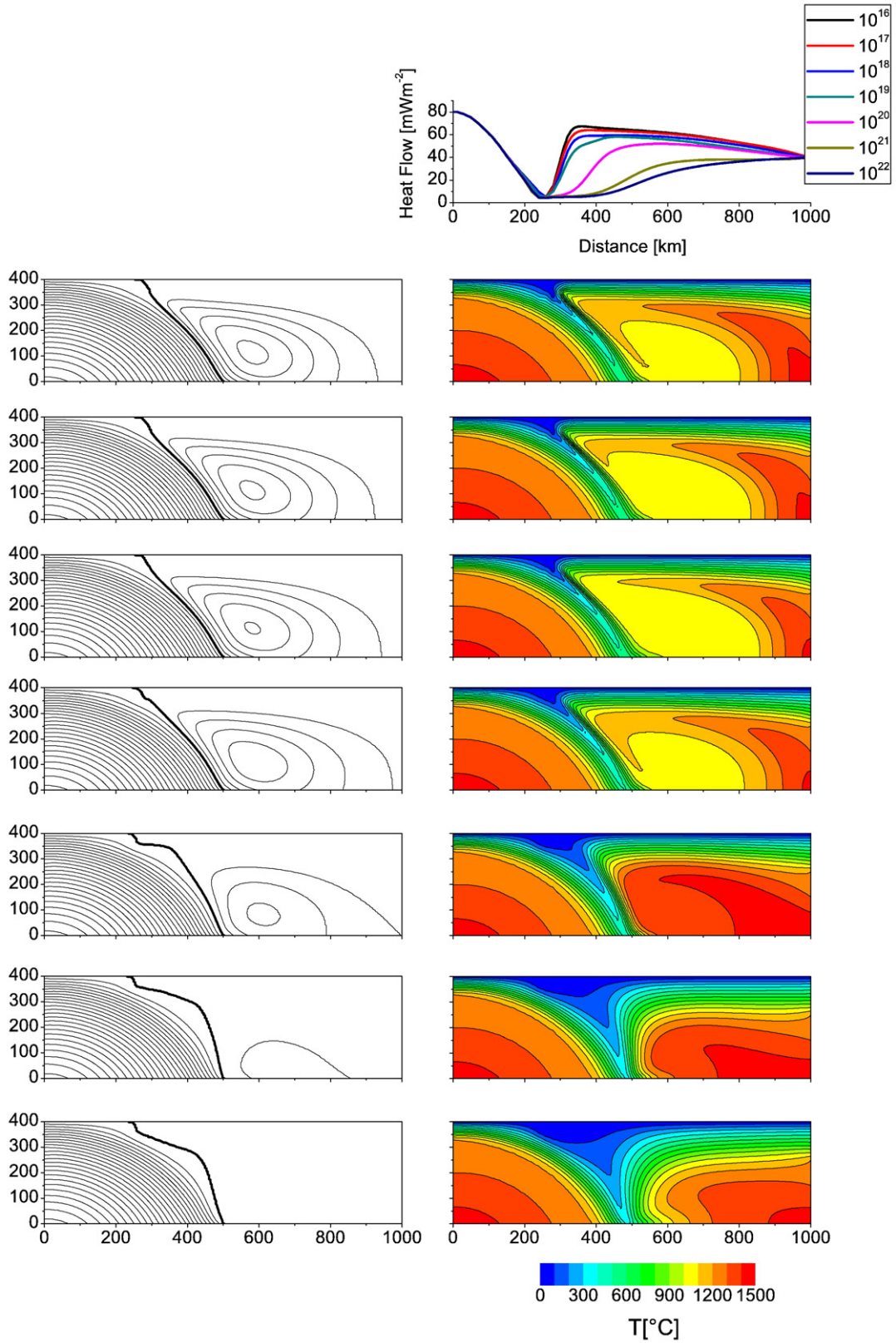


**Fig. 2.** The comparison of the flow patterns, temperature fields, and surface heat flow for different values of the constant mantle wedge viscosity  $\eta_{4W}$  changing from  $10^{16}$  (top) to  $10^{22}$  Pa s (bottom). Other viscosity values ( $\eta_1$ ,  $\eta_2$ ,  $\eta_{4C}$ ) are set to  $10^{23}$  Pa s except the weaker topmost layer in the subducting lithosphere ( $\eta_3$ ) which is set to  $10^{20}$  Pa s. The stream function is scaled by the value  $\psi_H$ ; the contour interval is  $\psi_H/10$ . The adiabatic temperature gradient is considered along the whole bottom boundary.

advantage of such an approach consists in the fact that the (de) coupling between the slab and the wedge does not require any special treatment. In the usual kinematic approach, decoupling in the shallow wedge must be somehow handled, e.g. by imposing a stagnant cold nose of the mantle wedge (Abers et al., 2006) or by applying a reduced velocity boundary condition at the slab–wedge boundary up to a certain depth (Kneller et al., 2007). Such model tuning seems to be a kind of arbitrariness. Moreover, it was shown that the way how decoupling of the subducting lithosphere from the overlying wedge is treated can substantially change the calculated temperature field (Conder, 2005).

This is the reason why we think it is important to include the slab–wedge boundary in a more self-consistent way. In the cases when the actual slab shapes are known well from local tomographic studies, they could be employed as additional information for evaluation of the models a posteriori.

The aim of this study is to find a rheology which is capable to reproduce the observed surface heat flow pattern in back-arcs under various temperature boundary conditions. This is important since the

increase in surface heat flow in back-arcs is reported from nearly all subduction zones, but one may assume that temperature conditions at the base of back-arc may differ from case to case. Therefore, a rheology successful in reproducing the observed heat flow pattern for a wide range of boundary conditions could be a good candidate for the actual rheology. Rheology of back-arc may be affected by a number of parameters; besides temperature and pressure, it depends on mineral composition, water content or grain size. The fact that the slab–wedge interface is a part of the solution in our approach makes the problem non-linear even in the case when different constant viscosity values are prescribed in the slab and in the wedge. To make the problem in this study more tractable, we thus deal with the influence of only temperature- and pressure-dependence of viscosity to the state of the mantle wedge.

## 2. Model description

We have formulated a 2-D model of the flow in the mantle wedge, which enables to calculate simultaneously the velocity field, the

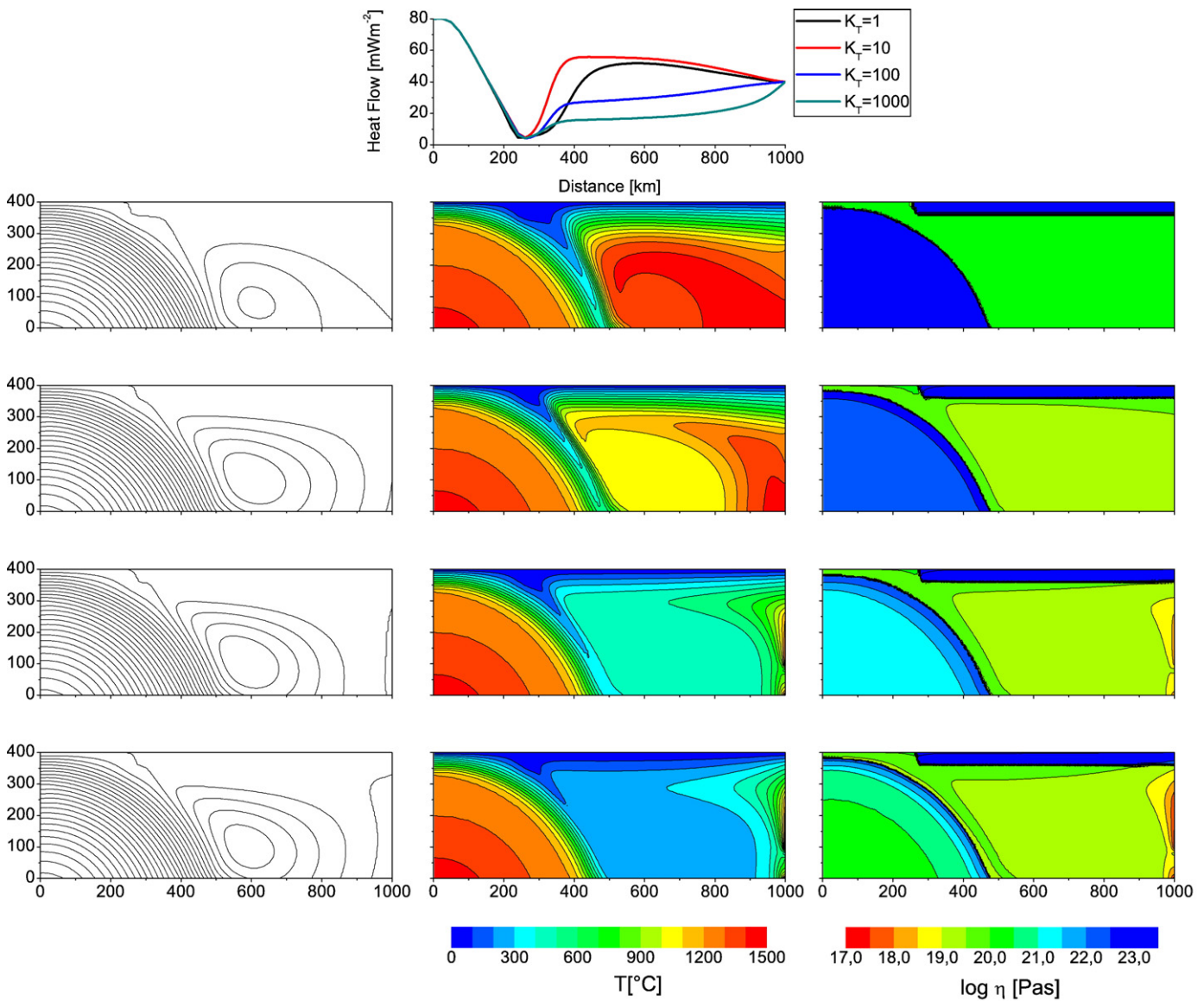


**Fig. 3.** The comparison of the flow and temperature fields and surface heat flow for a purely temperature-dependent viscosity ( $K_p=1$ ) with  $K_t$  changing from 1 (top) through 10 and 100 to  $10^3$  (bottom) in the case  $\eta_{4W}=10^{20}$  Pa s. Other viscosity values and representation of the stream function is the same as in Fig. 2.

temperature distribution, and particularly the boundary between the slab and the wedge. The steady state flow is described by the Stokes equation

$$-\nabla p + \nabla \cdot (\eta (\nabla v + (\nabla v)^T)) = 0 \quad (2)$$

and the continuity equation

$$\nabla \cdot v = 0. \quad (3)$$

In the Stokes equation the body forces are not taken into account as we are focused on the flow in the mantle wedge driven solely by the subduction traction. The temperature distribution is given by the steady state heat transfer equation

$$\rho c_p v \cdot \nabla T = \nabla \cdot (k \nabla T) + D : \nabla v + \rho \alpha T v \cdot g, \quad (4)$$

which includes also dissipation of heat. Radiogenic heating in the crust is not taken into account since we are focused purely to the effect of heat carried from the depth by convection in the wedge. Radiogenic heat sources localized in the continental crust are not the cause of hot back-arcs and their contribution to surface heat flow is additive.

Moreover, the dissipative terms are neglected in the continental crust, which is assumed to be a stagnant lid modeled as a high viscosity layer.

The above mentioned set of equations is solved in a rectangular domain  $1000 \times 400$  km. The boundary conditions for velocities, which are the only source of the flow in our model, are as follows: the free-slip condition

$$v \cdot n = 0, \quad (5)$$

$$D \cdot n - (n \cdot D \cdot n) n = 0, \quad (6)$$

(i.e., no flow through a boundary and zero tangential force acting on a boundary) is prescribed on the top boundary and the no-force condition

$$D \cdot n = 0 \quad (7)$$

at the rest of the boundary except the parts, where an inflow or an outflow is given; for details see Fig. 1. In principle, mass can freely flow through the boundary, where Eq.(7) is prescribed.

The lithospheric inflow is modeled by means of a prescribed velocity  $v_0$  at the topmost part of the left boundary. The depth interval

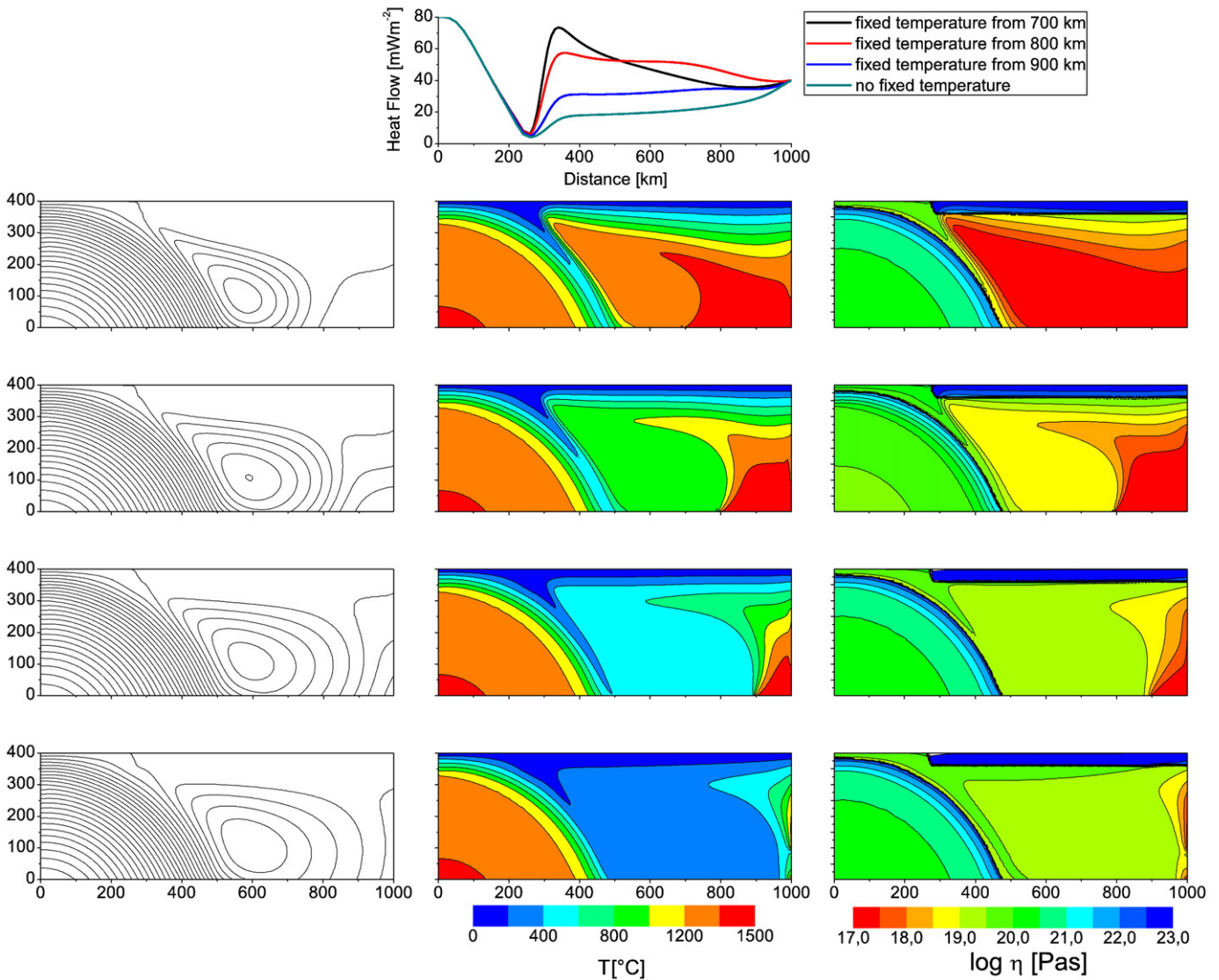


**Fig. 4.** The different flow patterns, the temperature distributions, and surface heat flow in the case of temperature-dependent viscosity  $K_T = 10^3$  ( $K_p = 1$ ) for different thermal boundary conditions at the bottom: The fixed temperature  $1450^\circ\text{C}$  is applied at the bottom boundary from the point  $x_T$  equal to 700 km (top), 800 km, 900 km, and 1000 km (bottom). Representation of the stream function and values  $\eta_1$ ,  $\eta_2$ ,  $\eta_3$ ,  $\eta_{AC}$ , and  $\eta_{AW}$  are the same as in Fig. 2.

of the inflow is denoted by  $H$ . The lithospheric outflow is modeled as a part of the bottom boundary at which both horizontal and vertical velocities are prescribed. The end point of the outflow of the slab is located in the middle of the bottom boundary and the outflow width is  $H/\sin \delta$ , where  $\delta$  means the given dip angle. The prescribed velocities are  $v_x = v_0 \cos \delta$  and  $v_y = v_0 \sin \delta$  and thus the mass flow through the outflow boundary is equal to the mass flow through the inflow boundary.

At the side parts of the rectangular domain the geotherms are prescribed: the oceanic geotherm on the left and the continental geotherm on the right up to a depth of 300 km, below which the zero heat flux through the boundary is assumed. The temperature at the top boundary is set to 0 °C, while at the bottom of the computational domain a variable boundary condition is used: an adiabatic temperature gradient 0.3 °C/km is prescribed from  $x=0$  to a certain distance  $x_T$ , which is considered to be a free parameter. In the rest of the boundary temperature is fixed to 1450 °C, i.e., a typical temperature for the depth of 400 km (Ito and Katsura, 1989). This is the crucial point of this study – the exact boundary condition is not known and it is supposed to be different for various subduction zones. For steeply subducting slabs the point  $x_T$  is presumably close to the prescribed outflow since

the area underlayed by the slab is relatively narrow. For shallower slabs the point  $x_T$  should be farther, however, an important source of uncertainties comes from the slab behavior, which may be affected by the phase transition at 660 km; flattening of the subducting slabs at 660 km discontinuity was both observed by seismic tomography (van der Hilst, 1995) and calculated in numerical models (Christensen, 1997). If a large part of the back-arc was underlayed with the cold slab, the point  $x_T$  might be relatively far from the prescribed outflow. Although the actual thermal boundary condition at the bottom of the back-arc may differ from case to case, elevated surface heat flow is observed throughout back-arcs, so it should be insensitive to prescribed thermal conditions.

We employ viscosity formally dependent on the stream function. The idea is that in the steady state streamlines separate different parts of the fluid since there is no flow through the streamlines. The stream function  $\psi$  is defined in terms of velocity derivatives

$$v_x = \frac{\partial \psi}{\partial y}, \quad v_y = -\frac{\partial \psi}{\partial x}, \quad (8)$$

and thus it is non-unique. We can achieve its uniqueness by defining  $\psi = \psi_H = v_0 H$  at the left upper point of the rectangular domain. Then the

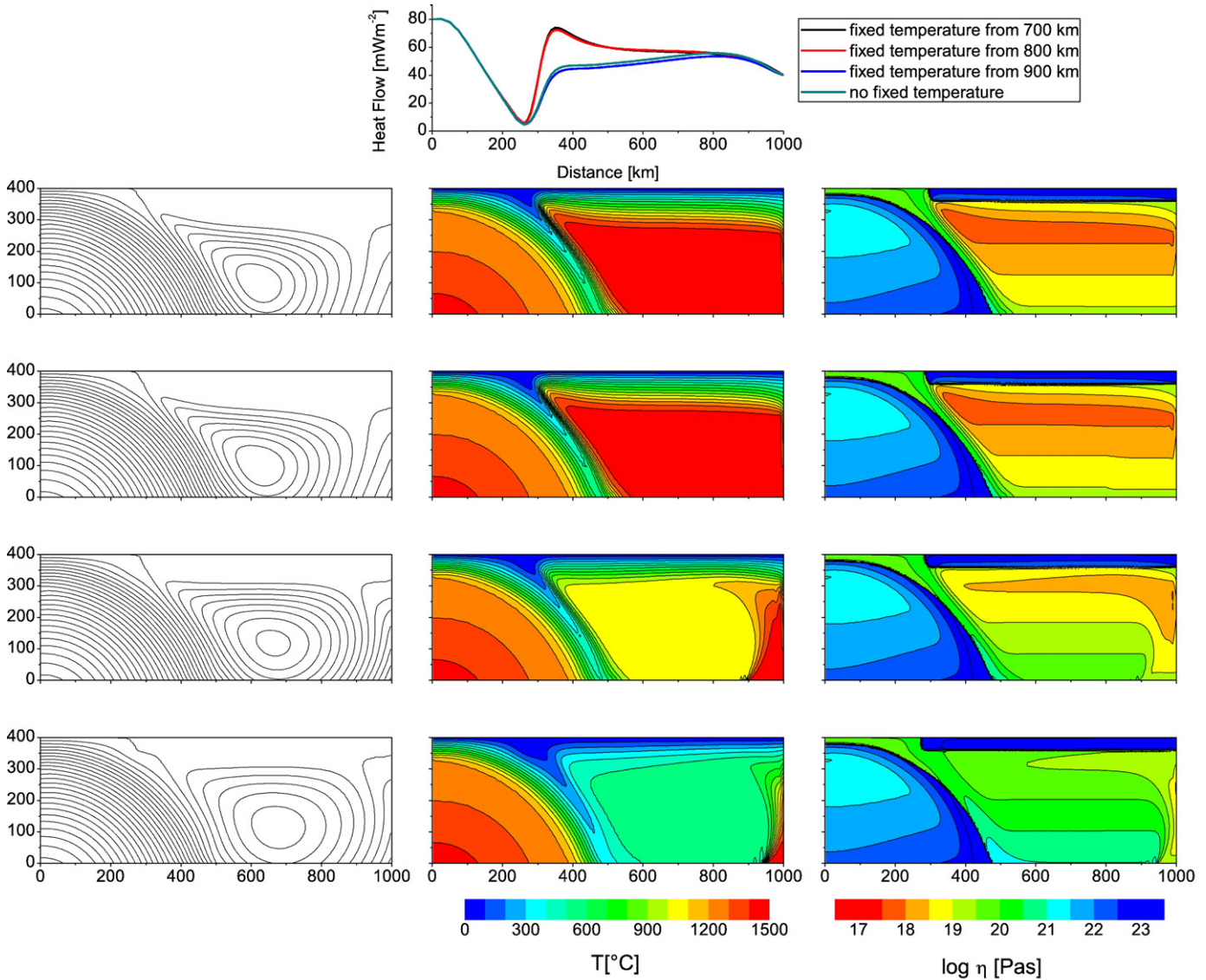


Fig. 5. The same as Fig. 4, but the pressure-dependence  $K_p = 10^2$  was used.

subducting slab is bordered by the streamline  $\psi=0$  on the bottom and  $\psi=\psi_H$  on the top. We then define the viscosity as

$$\eta = \eta_R(x, y, \psi) \exp\left(-\ln(K_T) \frac{T}{\Delta T} + \ln(K_p) \frac{p}{\Delta p}\right), \quad (9)$$

where  $\eta_R$  is the reference viscosity prefactor. The considered temperature- and pressure-dependence of viscosity is thus a simplified version of Eq.(1), which is broadly employed in convection studies to distinguish between both dependences in a more straightforward way (Christensen, 1989; Čadež and van den Berg, 1998; Trubitsyn et al., 1999). The different parts of the model are defined by

$$\eta_R = \begin{cases} \eta_1, & \psi < 0 \\ \eta_2, & \psi \in (0, (1-w)\psi_H), 0 < w < 1 \\ \eta_3, & \psi \in ((1-w)\psi_H, \psi_H) \\ \eta_4(y), & \psi > \psi_H \end{cases} \quad (10)$$

In this way we can separate four distinct parts as shown in Fig. 1: the mantle below the oceanic lithosphere with the viscosity prefactor  $\eta_1$ , the subducting slab except its top with the viscosity prefactor  $\eta_2$ ,

the topmost part of the slab with the viscosity prefactor  $\eta_3$ , and the mantle wedge and the overlying plate with the viscosity prefactor  $\eta_4$ . The prefactors  $\eta_1$ – $\eta_3$  are constant and  $\eta_4$  is a simple function of depth: the lithosphere overlying the subducted slab is divided into the continental crust (its prefactor is denoted as  $\eta_{4C}$ ) and the wedge ( $\eta_{4W}$ ) by specifying a crustal depth. The subduction top is treated separately as we suppose it is significantly weaker than the rest of the subducted plate. The ratio between the thickness of the weakening zone at the inflow part of the boundary and the thickness of the whole subducting lithospheric plate is denoted by  $w$ .

### 3. Results

The results were calculated using a finite-element code successfully tested against the steady state tests from the classical benchmark commonly used in mantle simulations (Blankenbach et al., 1989). The mathematical background of the used method was in detail described in Kukačka and Matyska (2004). Briefly, the problem is non-linear since the viscosity depends on the velocity through the stream function and thus the Picard iterations are employed (Cuvelier et al., 1986): first the energy Eq.(4) is solved using some starting velocity

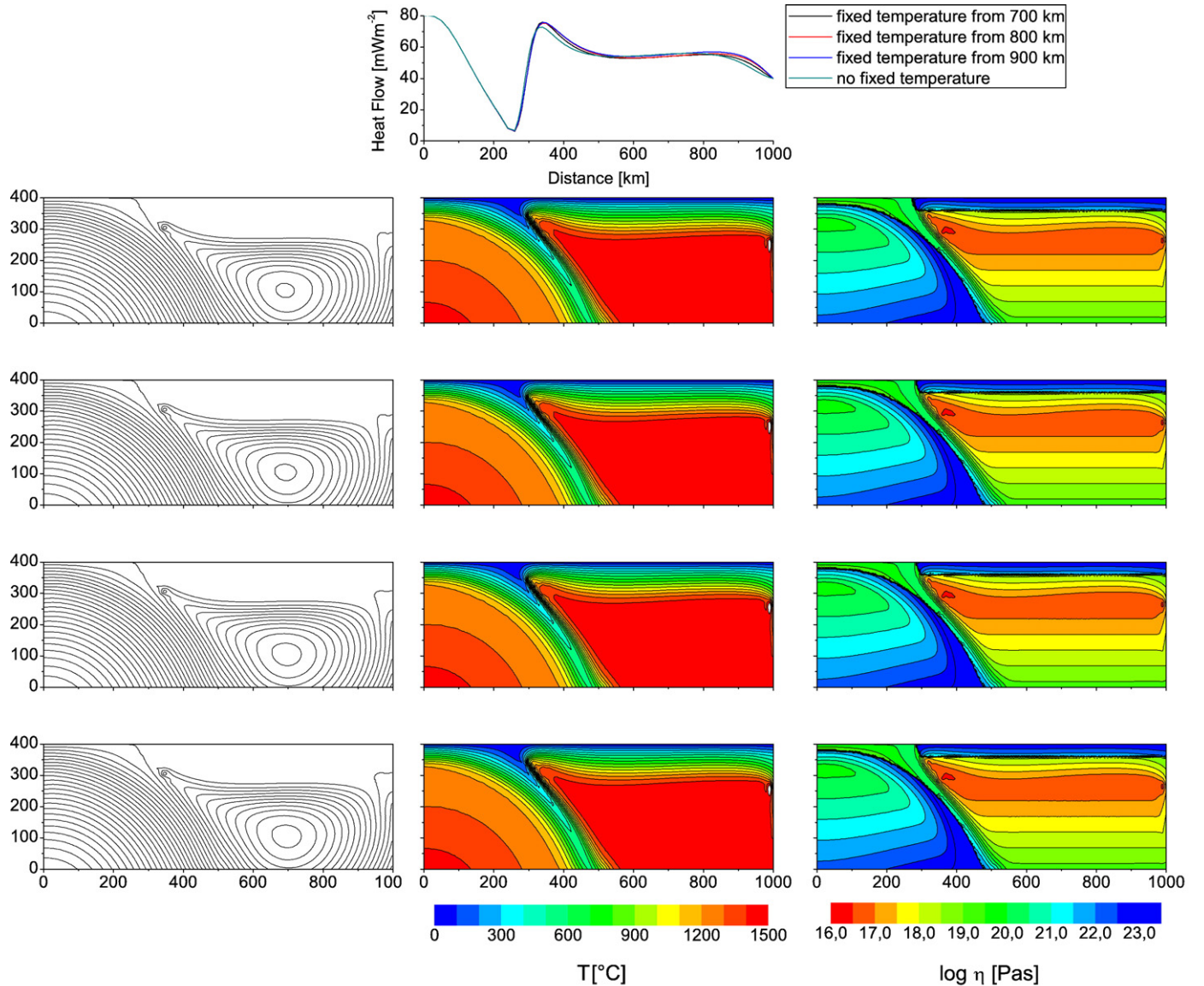


Fig. 6. The same as Fig. 4, but stronger temperature and pressure-dependence  $K_T=10^5$  and  $K_p=10^4$  was used.

field and then the obtained temperature field is used in the Stokes problem, Eqs.(2) and (3), to obtain a new velocity field. The iterative process

$$T^0 \rightarrow \psi^1 \rightarrow T^1 \rightarrow \psi^2 \rightarrow \dots \rightarrow T^n \rightarrow \psi^{n+1} \rightarrow T^{n+1} \rightarrow \psi^{n+2} \rightarrow \dots \quad (11)$$

continues until the convergence is reached. To overcome numerical issues underrelaxation is required.

We set the parameters so as to model a generic subduction zone: in the first set of simulations, the slab thickness  $H$  is 100 km and the bottom dip angle  $\delta$  is set to  $60^\circ$  which is the average deep angle (taken between 100 and 400 km) over all subduction zones (Jarrad, 1986). We used the slab velocity  $v_0$  equal to 4.5 cm/yr. Other parameters were set as described in Table 1.

On the left boundary the oceanic geotherm giving the surface heat flow  $80 \text{ mW m}^{-2}$  was prescribed, whereas on the right boundary the cold continental geotherm producing the surface heat flow  $40 \text{ mW m}^{-2}$  was applied up to the depth of 300 km. Below this depth zero heat flux through the boundary was assumed.

First, we investigated the case where individual domains have a constant viscosity (i.e.  $K_T = K_p = 1$ ) in order to gain an insight into the physics of the problem. It turned out that the value of viscosity  $\eta_1$  does not affect the flow in the wedge. What is important is the viscosity  $\eta_3$  in the uppermost part of the subducting slab. If it is sufficiently lower than the viscosity of the continental crust, about three orders of magnitude, a relatively narrow zone of weak viscosity is formed at the top of the subducting slab. The weak oceanic crust acts as a lubricant

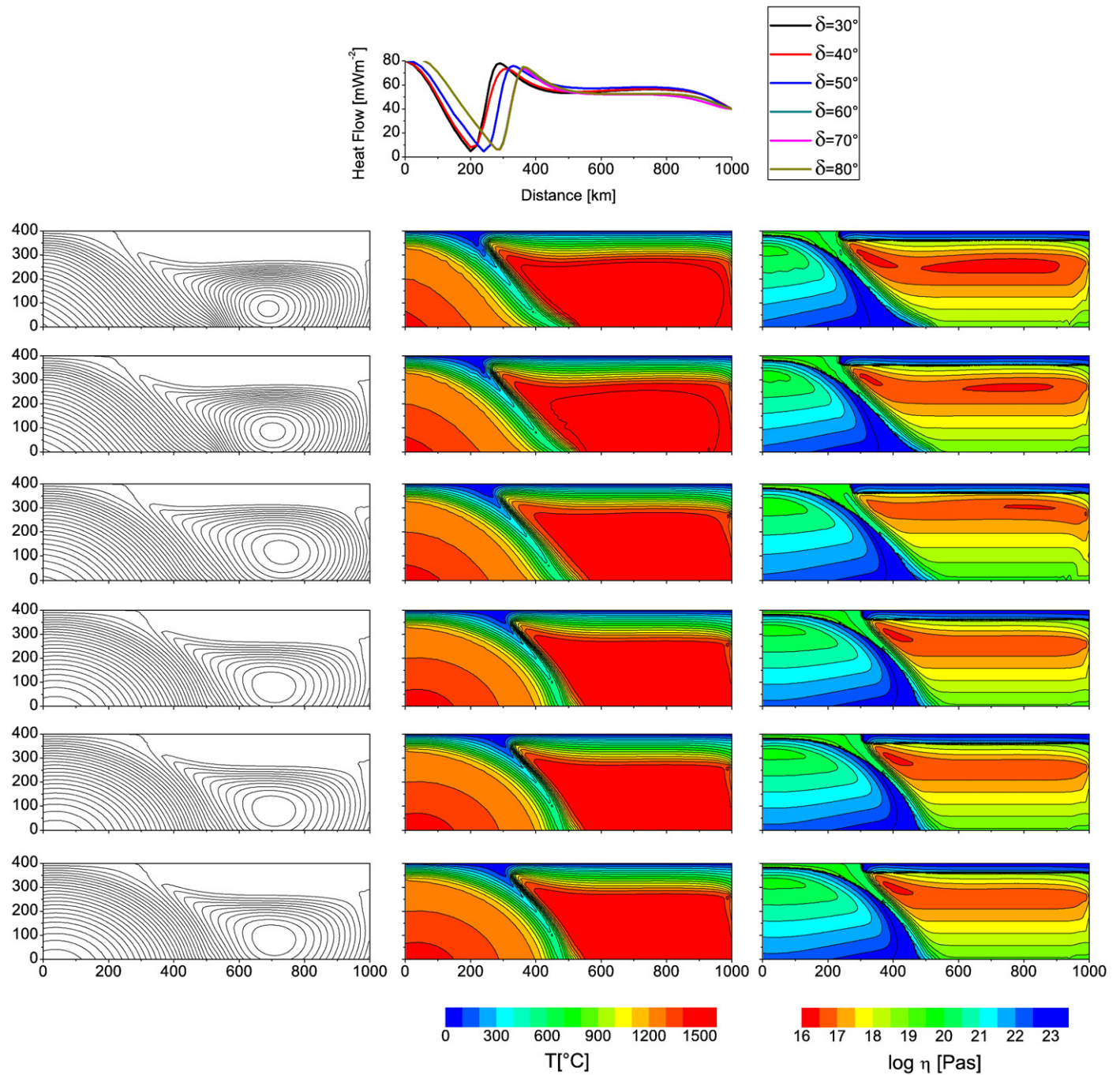


Fig. 7. Temperature- and pressure-dependent viscosity ( $K_T = 10^5$ ,  $K_p = 10^4$ ) with the adiabatic temperature gradient applied at the bottom for different dip angles  $\delta$  changing from  $30^\circ$  to  $80^\circ$ .

and effectively reduces viscous forces transferred from the subducting slab to the overriding lithosphere. If, however, the viscosity in the wedge below the continental crust is reduced, lubrication becomes inefficient and a substantial flow occurs in the wedge. It is demonstrated in Fig. 2, where the adiabatic temperature gradient is applied at the whole bottom boundary. The viscosity of the wedge changes in a range  $10^{16}$ – $10^{22}$  Pa s: for high viscosity values the wedge is nearly stagnant and heat is transferred dominantly by conduction, which results in relatively high temperatures at the bottom but low surface heat flow. For low viscosity values a substantial corner flow occurs and convection dominates — heat from the bottom is transferred to the surface and surface heat flow increases, but a cold material neighboring the subducting slab cools down an inconsiderable part of the bottom. A balance between these two heat transfer mechanisms appears when the viscosity values of the uppermost part of the subducting plate and the wedge are roughly equal. Then a significant part of the wedge is hot (temperatures above  $1200^\circ\text{C}$ ) and simultaneously a pattern of elevated surface heat flow may be observed.

In the case of pure temperature-dependence (i.e.  $K_T > 1$  and  $K_p = 1$ ), the corner flow becomes more localized to the region near the subducting plate. Stronger temperature-dependence results in stronger localization and it substantially affects the temperature field. It is documented in Fig. 3, where the parameter scaling temperature-dependence  $K_T$  varies between 1 and  $10^3$ . Since the adiabatic temperature gradient is again applied at the bottom, the narrower corner flow does not carry enough heat from the bottom and the whole wedge is cooled down with increasing flow localization. The temperature field in the wedge is very sensitive to the thermal

boundary condition applied at the bottom; moving the point  $x_T$  from where a temperature value  $1450^\circ\text{C}$  is fixed, towards the subducting plate results in an increase of wedge temperatures and surface heat flow as well, which is shown in Fig. 4. Due to the flow localization effect and the strong sensitivity to the bottom boundary condition, the purely temperature-dependent viscosity does not successfully reproduce the observed pattern of the surface heat flow.

If also the pressure-dependence of viscosity is employed, the temperature field in the wedge may change drastically. Fig. 5 shows the case with  $K_T = 10^3$  and  $K_p = 10^2$  under variable thermal boundary conditions applied at the bottom. Even if a pattern of elevation in surface heat flow is developed only for a transition point close to the sinking slab ( $x_T = 700, 800$  km), a clear trend is obvious: the corner flow can carry more hot material from greater depths and distances.

Fig. 6 shows simulations when even stronger temperature- and pressure-dependence  $K_T = 10^5$  and  $K_p = 10^4$  is used. In this case, an increase in surface heat flow is obtained even in the case of the adiabatic temperature gradient at the whole bottom boundary. The strong pressure-dependence of the viscosity leads to a low viscosity layer beneath the continental crust, which allows rapid corner flow (velocities of the same order as the convergence rate). Such a rapid flow can carry upward the materials, which are far from the subducting slab and therefore not cooled down, and, consequently, almost uniformly hot region below the back-arc is developed. A surface manifestation of such hot regions, elevated surface heat flow, occurs.

The question arises, whether the elevated back-arc surface heat flow obtained for strongly temperature- and pressure-dependent viscosity is robust enough, i.e. whether it is not sensitive to reasonable

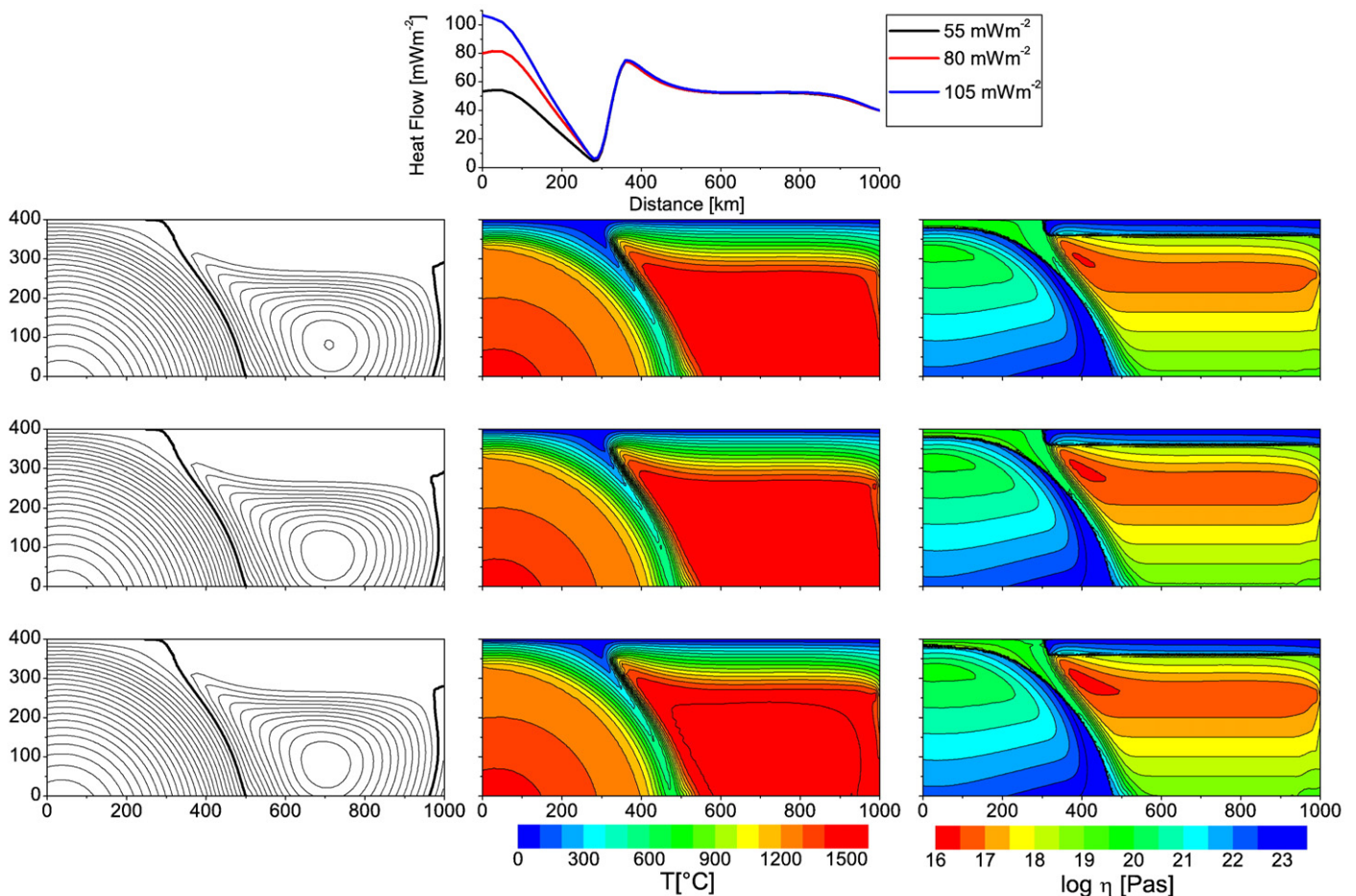


**Fig. 8.** Temperature- and pressure-dependent viscosity ( $K_T = 10^5$ ,  $K_p = 10^4$ ) with the adiabatic temperature gradient applied for different oceanic geotherms giving the surface heat flow value in the range of  $50$ – $100$   $\text{mW m}^{-2}$ .

changes of other parameters of the model. We have found that such a pattern is stable over a broad range of tested bottom dip angles ( $30^\circ$ – $80^\circ$ , see Fig. 7), different convergence rates (4–10 cm/yr) and different oceanic geotherms (yielding 50–100 mW m<sup>-2</sup> at the surface, Fig. 8). On the other hand, if a pressure-dependence is not considered, the corner flow is focused to the region close to the subducting slab, which is extensively cooled down with the cold incoming lithosphere, and thus such a viscosity does not produce hot back-arcs. Note that our models are still semi-kinematic as they are driven by kinematic boundary conditions. As the horizontal temperature gradient in Figs. 5–8 is negligible below back-arcs, the non-hydrostatic gravity forces need not be taken into account in this region. In other words, the resultant wedge flow pattern of our models is generated by the interaction of the wedge with the subducting plate, which is created as the response to the inflow (outflow) into (from) the computational domain.

#### 4. Conclusions

We have used the method developed earlier (Kukačka and Matyska, 2004) to calculate surface heat flow in subduction back-arc regions. The presented non-linear models allow to calculate simultaneously both the temperature and flow field without incorporating assumptions about the position of the whole slab–wedge boundary, which is needed information in recent models (van Keken et al., 2002; Currie et al., 2004); only the location and dip angle at the outflow are prescribed. We investigated how the different kinds of viscosity affect the mantle wedge flow and, particularly, the surface heat flow pattern, which is directly observable quantity. Observed surface heat flow in almost all back-arcs is remarkably higher than in the continental lithosphere (Currie and Hyndman, 2006). Therefore, we searched for a viscosity law, which would be capable to reproduce this effect under various thermal conditions at the base of the back-arc. Numerical models revealed that the back-arc heat flow increase may be reproduced even with a constant viscosity of the wedge if its value is sufficiently low. More realistic, temperature-dependent viscosity, however, does not reproduce such a pattern and the results are extremely sensitive to the thermal condition applied at the bottom. Strong temperature-dependence leads to localization of the corner flow to a region close to the subducting slab and such a localized flow is unable to advect enough heat from the depth. On the other hand, considered pressure-dependence may compensate this effect since it results in a low viscosity layer below the back-arc continental crust. If the viscosity of this layer is sufficiently low (about  $10^{17}$  Pa s), the pattern of surface heat flow increase is stable under various thermal boundary conditions and various settings of the subducting slab including convergence rate or dip angle. Note that we omitted the radiogenic heating; if it were added into the wedge crust, it would result in an additional increase of surface heat flow in the wedge. There is still a space for such an increase as the modeled heat flow in back-arcs is still slightly lower than the observed. Our heat flow outputs may thus be considered as the heat flow part of deep origin.

In general, our preferred viscosity distribution is in agreement with the idea of a low viscosity wedge (Billen and Gurnis, 2001) although the magnitude of viscosity decrease seems to be rather high. It is possible that such a viscosity weakening could be also caused by compositional effects and/or water content, anyway a very low viscosity zone in shallow portion of the back-arcs could be the physical reason why the back-arcs are relatively hot.

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