

FAST TRACK PAPER

On the possibility of a steady state backflow in the mantle wedge

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SUMMARY

Viscous coupling between the subducting slab and the overriding mantle wedge may drive a corner flow in the forearcs and, consequently, may affect the temperature in the mantle wedge. We developed a numerical model of a generic pseudo-plastic subduction plate, in which the steady-state thermal structure of the mantle wedge is calculated without any *a priori* assumptions on the geometry of the subducting slab. The results showed that both the temperature field and the geometry of the slab are strongly controlled by weakness zones in the topmost part of the subducting plate and in the cold tip of the mantle wedge. We showed that a backward flow may evolve along the plate–wedge interface if the cold portion of the mantle wedge is weak, which, likely, is true as this region is hypothesized to be strongly serpentinized. We concluded that the self-consistent generation of the plate geometry should be addressed in the kinematic subduction models because such a strongly non-linear system may lead to unforeseen effects.

Key words: Subduction zone processes; Rheology: crust and lithosphere.

1 INTRODUCTION

During subduction, the oceanic crust and the upper part of the lithosphere are buried deeply into the mantle. Subducted materials undergo high-pressure metamorphic changes, and consequently, a portion of these metamorphosed rocks again appear at the Earth's surface. The exhumed rocks found in the forearc regions are mainly composed of eclogites (Tsujimori *et al.* 2006a), which might be surprising since eclogite is substantially denser (3.5 g cm^{-3}) than the surrounding peridotite (3.3 g cm^{-3}). Therefore, various mechanisms of high-pressure and ultrahigh-pressure metamorphic rock uplift were suggested, but nowadays, buoyancy is considered to be the only force sufficient for exhumation of rocks from depths of about 100 km to a lower crustal level (Platt 1993).

The exhumation rates are comparable to plate motion rates, that is, are of the order of centimeters per year: the youngest eclogite found at the Earth's surface exhumed from a depth of about 75 km is about 4 Myr old (Baldwin *et al.* 2004). The exhumation rates, however, do not correlate with the convergence rates. Whereas the subduction rate of the Adriatic plate was about 2 cm yr^{-1} and the exhumation rate observed in the Western Alps is 1.3 cm yr^{-1} , in the Himalaya, the convergence rate was estimated to $10\text{--}18 \text{ cm yr}^{-1}$ and the exhumation rate is just about 0.5 cm yr^{-1} (Guillot *et al.* 2001). The common feature of eclogites observed at the Earth's surface is simultaneous presence of serpentinites reported from many places, for example, Alps (Hermann *et al.* 2000), Himalaya (Guillot *et al.* 2000), and classical examples of high-pressure low-temperature eclogites are from California (Tsujimori *et al.* 2006b). Serpentinites

are supposed to play a crucial role in the exhumation processes due to lowering the bulk density of the exhumed serpentinite–eclogite mixture (Hermann *et al.* 2000); the density of antigorites is about 2.8 g cm^{-3} , and therefore, they are extremely buoyant with respect to the surrounding peridotites and eclogites. Thus, widely reported serpentinization of forearc wedges (Hyndman & Peacock 2003) is a common attribute of almost all subduction zones, which allows exhumation processes.

The corner flow of a buoyant material in the forearc mantle wedge has been modelled several times in its simplest form, as a 1-D problem (Schwartz *et al.* 2001; Gerya & Stöckert 2002; Raimbourg *et al.* 2007). The published simplified models assume a channel-localized backward flow along the interface between the subducting slab and the mantle wedge and demonstrate that not only the low density of serpentinites but also their low viscosity are necessary for the rocks uplift. These 1-D models, however, are oversimplified, as they do not deal with changes of the temperature distribution due to the upward flow, which is important, particularly because serpentinites are stable only at relatively low temperatures (Ulmer & Trommsdorf 1995).

On the other hand, the thermal structure of the forearc regions has been modelled using kinematic steady state models, for more than 15 yr (Peacock 1996). The number of these models is constantly growing, for example, Abers *et al.* (2006), Kneller *et al.* (2007) and Peacock *et al.* (2005), and the popularity of the kinematic steady-state models in the geophysical community increases (van Keken *et al.* 2008), but these models still do not take the exhumation processes into account. The purpose of steady-state forearc models

is primarily to calculate the temperature distribution in the mantle wedge and consequently derive the P – T conditions at the top of the subducting plate (Peacock *et al.* 1994, 2005). Their long-term goal is to answer the question, at which depths the subducted sediments and the oceanic crust start to melt and eventually map the magma transportation paths (Manea *et al.* 2005). Since exhumed rocks carry a significant record of temperature at greater depths, they should be utilized in these models as well. Basically, the record preserved in the exhumed rocks shows two different types of retrograde P – T paths: Alpine-type exhumation is characterized by nearly isothermal decompression whereas Franciscan-type is characterized by decompression and simultaneous cooling (Ernst 1988). Recently, the Alpine-type exhumation has been found, for instance, in eclogite samples from southern Mexico (decompression from 1.6 GPa to 0.6 GPa at temperatures about 600°C; Meza-Figueroa *et al.* 2003) or in exhumed eclogites in Usagaran belt in Tanzania (decompression from 1.7 GPa to 0.7 GPa at lower crustal levels at the temperature 750°C, before cooling and further uplift; Herms 2004). The other constraint on the temperature field model is given by the measurements of the surface heat flow. In forearc regions, measured surface heat flow is generally low, about 40 mW m^{−2} (Ziagos *et al.* 1985; Hyndman *et al.* 1995), but values even as low as 15 mW m^{−2} have been reported (Grevemeyer *et al.* 2005).

Although it is believed that a viscous coupling between the subducting slab and the overriding mantle drives the flow in the mantle wedge, all the current kinematic models use the approach in which the temperature field is calculated both in the wedge and in the subducting plate, but the flow field is calculated only in the mantle wedge (van Keken *et al.* 2002). It has been shown that the treatment of the shallow portion of the lithosphere in these numerical models considerably affects the calculated temperature at the plate–wedge boundary (Conder 2005); however, the published steady state models prescribe a fixed position of the boundary between the subducting slab and the mantle wedge. In such an approach, a part of the flow field (solution) is imposed on the results *a priori*.

Besides 1-D models of the exhumation in a relatively narrow subduction channel, more complex 2-D models are in use (Gerya *et al.* 2002; Stöckert & Gerya 2005). These are based on kinematic prescription of subducting plates and are focused on exhumation of high-pressure metamorphosed rocks in a wedge-shaped subduction channel. Together with field observations (Jolivet *et al.* 2005), they lead to conclusions that for the exhumation, which takes place in the subduction channel and/or in the tip of the wedge (Agard *et al.* 2009), the processes that are able to soften the rocks are more important than possible density variations and their buoyancy effects.

Our study follows this line of research; however, the subducting plate is incorporated into our models in a semi-dynamic way. We consider a pseudo-plastic viscosity and calculate the steady state of the subduction zone, using a technique that does not require the location of the plate boundaries to be prescribed in advance. We particularly investigate the effect of rheological weakening in the wedge to self-organization of the flow and temperature field and the possibility to provide physical explanation of exhumation processes in greater depths. Our models are focused to a subcrustal level only since further exhumation to the surface is controlled by shear failure along faults (Platt 1993).

2 MODELLING APPROACH

We calculate the steady-state thermal convection, solving the equations of conservations of mass, momentum and energy

$$\nabla \cdot \mathbf{v} = 0, \quad (1)$$

$$-\nabla p + \nabla \cdot [\eta(\nabla \mathbf{v} + (\nabla \mathbf{v})^T)] = 0, \quad (2)$$

$$\rho c_p \mathbf{v} \cdot \nabla T = \nabla \cdot (k \nabla T) + \mathbf{D} : \nabla \mathbf{v} + \rho \alpha T \mathbf{v} \cdot \mathbf{g}, \quad (3)$$

where p denotes the non-hydrostatic pressure, η is the dynamic viscosity, $\mathbf{v}$ is the velocity, ρ is the density, c_p is the isobaric heat capacity, T is the temperature, k is the thermal conductivity, α is the thermal expansion coefficient and the symbol $\mathbf{D} = \eta(\nabla \mathbf{v} + (\nabla \mathbf{v})^T)$ denotes the deviatoric part of the stress tensor. In the Stokes eq. (2), the body forces are not taken into account, as we are focused on the flow in the wedge driven solely by a traction from the subducting plate.

A viscoplastic rheology is used for the lithosphere. Both the diffusion and dislocation creeps are taken into account. The general viscosity law for constant water content is given by

$$\eta = \left(\frac{d^m}{A C_{OH}^r} \right)^{\frac{1}{n}} \dot{\epsilon}^{\frac{1-n}{n}} \exp \left(\frac{E + pV}{nRT} \right), \quad (4)$$

where A is a constant determined experimentally, d is the grain size, C_{OH} is the concentration of hydroxyl ions, E is the activation energy, V is the activation volume, p is the lithostatic pressure and R denotes the gas constant (Hirth & Kohlstedt 2003). This formula includes both Newtonian diffusion creep with $n = 1$ and non-Newtonian dislocation creep with $n \geq 3$. At the cold regions ($T \leq T_{PLAST} = 800^\circ\text{C}$) plastic behaviour is applied by introducing a pseudo-plastic viscosity $\eta = \eta_{yield} = \sigma_{yield}/2\dot{\epsilon}$, where σ_{yield} is the yield stress, which is a material parameter assumed to be different in different regions. The effective viscosity is then given by the equation

$$\frac{1}{\eta_{eff}} = \begin{cases} \frac{1}{\eta_{yield}} + \frac{1}{\eta_{diff}} + \frac{1}{\eta_{disl}}, & T < T_{PLAST} \\ \frac{1}{\eta_{diff}} + \frac{1}{\eta_{disl}}, & T \geq T_{PLAST} \end{cases} \quad (5)$$

where η_{diff} and η_{disl} are viscosities for diffusion and dislocation creeps defined in (4).

The computational domain is a rectangular box 300 × 100 km. The inflow through the left-hand boundary into the domain is prescribed (see the details of the model description in Fig. 1), but on the part of the bottom boundary, where the outflow is considered, only the normal velocity component is given, and the tangential velocity component is a part of the solution since the free-slip condition (no tangential force) is applied here. Such settings allow the model to freely develop the angle of subducting flow.

We use the finite element code developed before (Kukačka & Matyska 2004), which is based on the stream function formulation. The size of the rectangular mesh cells used for flow field calculation is 5 km; however, in areas of interest (i.e. in the upper part of the subducting plate and in the mantle wedge), the mesh is refined to a size of 1 km.

The stream function ψ is defined by the relations

$$v_x = \frac{\partial \psi}{\partial y}, \quad v_y = -\frac{\partial \psi}{\partial x}, \quad (6)$$

and therefore, any constant could be added to it. Utilizing this non-uniqueness of the stream function, we define $\psi=0$ in the left-hand lower corner of the domain. Consequently, the value ψ_B of the streamline, which separates the subducting plate from the overriding plate and the mantle wedge, is the product of the domain height H and the given velocity of the incoming lithosphere v_0 . The key point of the presented methods consists in the fact that in a steady state regime, the streamlines separate different parts of the fluid, since no mass flows through them. Thus, the streamlines can be employed to distinguish different structural parts of the lithosphere.

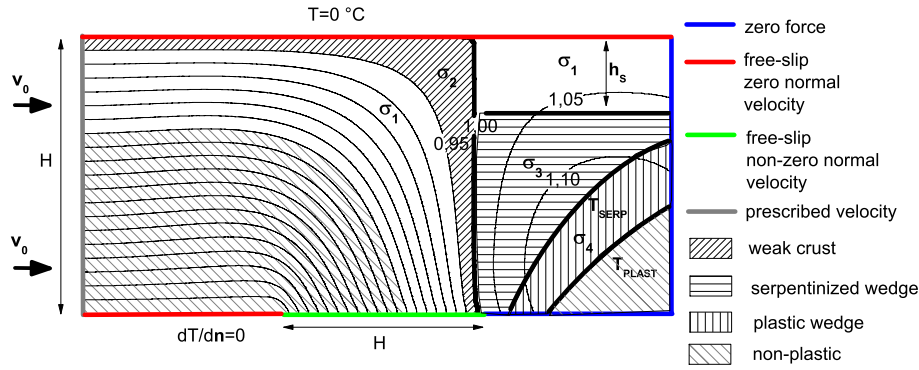


Figure 1. The model domain, 300×100 km, with boundary conditions applied to the velocity and the temperature. The streamlines in the case of constant viscosity are shown; the stream function is scaled by the value $\psi_B = Hv_0$, and thus, the contour of the value 1 denotes the interface between the incoming lithosphere and the mantle wedge. Regions with different yield stress value are hatched; in the mantle wedge two important isotherms are schematically shown: T_{SERP} specifies the serpentinites stability field and T_{PLAST} specifies the region of pseudo-plastic behaviour in the wedge. The region with a temperature higher than T_{PLAST} is marked as non-plastic.

In our model, we define the following parts of the lithosphere: oceanic crust and the upper part of the subducting plate; the rest of the subducting plate; continental crust in the upper part of the overriding plate; serpentinized mantle wedge and deserpentinized mantle wedge. Whereas the border between the topmost part of the subducting plate and the rest of the plate is roughly given by the thickness of the oceanic crust, the border between serpentinized and deserpentinized wedge is defined by the temperature T_{SERP} , below which serpentinites are stable (Ulmer & Trommsdorf 1995). These parts are assumed to have different values of the yield stress σ_{yield} , which is therefore defined as a function of the temperature T , the depth h and, particularly, the stream function ψ (see also Fig. 1):

$$\sigma_{yield}(\psi, T, h) = \begin{cases} \sigma_1, & \psi < (1 - \delta)\psi_B \text{ or } \psi > \psi_B, h < h_s, \\ \sigma_2, & \psi \in ((1 - \delta)\psi_B, \psi_B), \\ \sigma_3, & \psi > \psi_B, T < T_{SERP}, h > h_s, \\ \sigma_4, & \psi > \psi_B, T > T_{SERP}. \end{cases} \quad (7)$$

The yield stress in the uppermost part of the subducting oceanic plate (its thickness fraction of the overall plate thickness is δ) is σ_2 , whereas in the rest of the plate, the yield stress is σ_1 . In the overriding lithosphere, up to the depth denoted by h_s , there is the yield stress value σ_1 . The value h_s corresponds to the thickness of the continental crust in the forearc regions. The mantle wedge is then divided by the temperature value T_{SERP} into two parts, with the yield stress values σ_3 and σ_4 . In this way, the zones of weakening defined by different yield stress values are incorporated into the model without actual prescription of the plate–wedge interface position. Therefore, the spatial settlement of both the weak oceanic crust and the mantle wedge, which is supposed to be weak because it is highly serpentinized (Hyndman & Peacock 2003), is the outcome of the modelling.

We use the viscosity parameters for wet diffusion creep, $n = 1$, $E = 335$ kJ, $V = 4 \times 10^{-6}$ m³ mol⁻¹, and wet dislocation creep, $n = 3.5$, $E = 480$ kJ, $V = 11 \times 10^{-6}$ m³ mol⁻¹ (Hirth & Kohlstedt 2003). The absolute value of viscosity depends on the water content and the grain size, which are not known precisely. We set these values so that the viscosity at a depth of about 100 km is 10^{20} Pa s for both diffusion creep and dislocation creep, at strain rate 10^{-15} s⁻¹ for most of the simulations. The thickness of the continental crust h_s is fixed to 30 km.

3 RESULTS

We analysed the influence of the different magnitudes of the yield stress on the flow pattern and the temperature field in the mantle wedge and the plate–wedge boundary. In the parametric study, we first defined a viscosity weakening just in the oceanic crust. Fig. 2 shows the temperature and velocity fields for different magnitudes of the yield stress values σ_2 , whereas in all other parts, the yield stress is $\sigma_1 = \sigma_3 = \sigma_4 = 1000$ MPa. The results for all studied plate velocities v_0 ranging from 1 to 10 cm yr⁻¹ show that the plate-like behaviour develops only if there is a substantial drop in the strength of the oceanic crust compared with the strength of the overlying mantle. A lower ratio between the yield stress in the oceanic crust and in the overlying plate results in a less steeper dip angle, and the subducting plate sinks with smaller dip angle at shallower depths and with bigger dip angle in higher depths. Results, moreover, show that the wedge is very cold, as the surface heat flow above the wedge is as low as 15 mW m⁻². This is caused by the reduced viscous coupling between the subducting plate and the overlying mantle, which leads to a stagnant mantle wedge cooled by the incoming lithosphere.

The tip of the wedge is, however, supposed to be serpentinized and therefore weak, and we set the value $T_{SERP} = 700$ °C. This is because below this temperature, a number of hydrated minerals is stable (for instance antigorite starts to dehydrate at temperatures about 720 °C at pressures of about 2 GPa; Ulmer & Trommsdorf 1995). In this case, we observed a significant change in the flow field: Fig. 3 shows the temperature and the flow fields for different weakening magnitudes in the wedge for the convergence rate 5 cm yr⁻¹. Depending on the ratio between yield stress magnitudes in the oceanic crust and the cold portion of the mantle wedge, the backward flow may evolve along the plate–wedge interface. The velocity of the backward flow depends on the viscous coupling between the subducting crust and the wedge; if the effective viscosity of the crust is higher than that in the wedge, the circulation in the wedge is stronger. Such a drastic change in the flow field results in a remarkable change in the temperature field: the backward flow drags the warmer geotherms upward. As a result, the surface heat flow along the forearc is almost constant or may eventually increase if the viscous coupling is too strong, that is, if the effective viscosity of the subducting crust is higher than that of the cold nose of mantle wedge. The forced backward flow of weak rocks along the plate–wedge interface, thus, provides a feasible physical mechanism of exhumation processes in the greater depths.

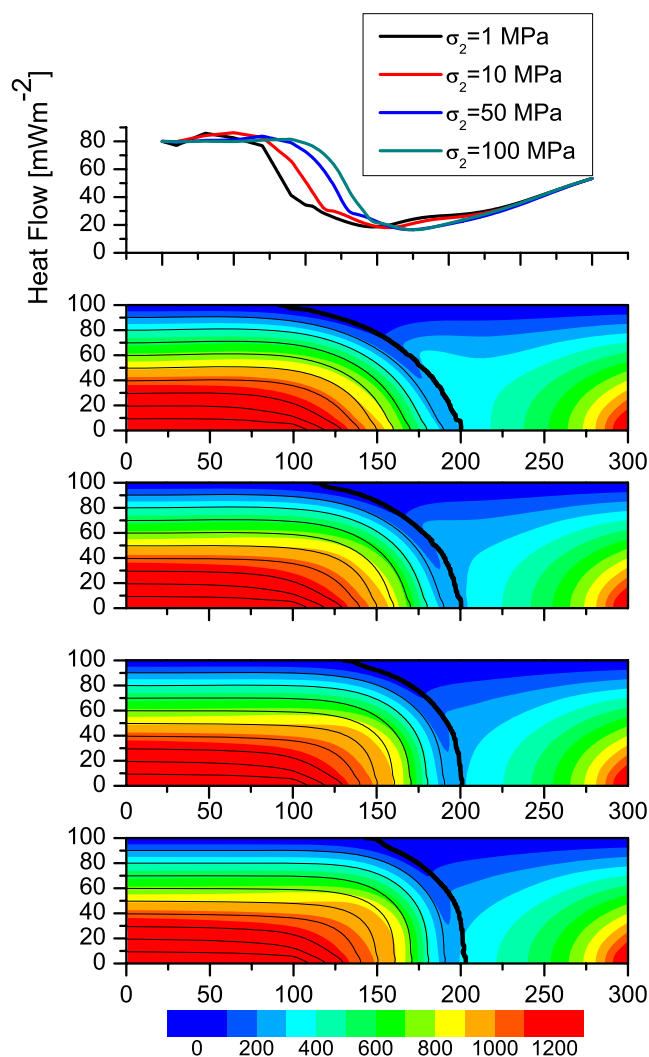


Figure 2. The temperature, the stream function (contour interval $0.1 \psi_B$) and the surface heat flow for a subducting plate of rate 5 cm yr^{-1} . Results calculated for different magnitudes of the yield stress in the oceanic crust without any weakening in the mantle wedge.

Fig. 4 shows the case, where $\sigma_1 = \sigma_4 = 1000 \text{ MPa}$ and $\sigma_2 = \sigma_3 = 10 \text{ MPa}$, in detail. A close inspection in the wedge close to the subducting plate reveals that some streamlines of the backward flow coincide with temperature isolines. It clearly shows that the decompression along the retrograde path in the return flow is isothermal or nearly isothermal. The example of P – T path along a streamline marked by the arrow in Fig. 4 is shown in Fig. 5, where the decompression from the depth of about 2.5 – 1.6 GPa at 600°C is depicted. The sample going along this path, approximately from a depth of 80 to 40 km , is cooled down only by 50°C , from which about 10°C could be attributed to adiabatic cooling.

The pattern of backward flow in the forearc wedge is observed for different convergence rates (see Fig. 6), and the vigour of the flow naturally depends on the subducting plate velocity. The results clearly demonstrate that there is a significant trade-off among the magnitude of the weakening in the yield stress, convergence rate, subduction zone geometry and the thermal state of the forearc. Weakening in the oceanic crust is necessary to produce small dip angles at shallow depths. If the weakening in the tip of the wedge is of the same order as in the oceanic crust (which likely is true in its cold

portion where one may assume similar serpentinization/hydration), the viscous traction is sufficient to drag the backward corner flow in the serpentinized portion of the wedge. Such an uplift brings upwards a warmer material from greater depths, which consequently influences the temperature in the wedge and effectively reduces the region where serpentinites are stable, that is, region with a low value of the yield stress.

Faster subducting plates cool down a larger portion of the mantle wedge. When the larger portion of the wedge is serpentinized (cold enough so as serpentinites are stable in it), the driven backward flow brings more heat from deeper parts upward, which restricts the cold portion of the wedge. Thus, the size of the serpentinized wedge is given by a balance between these two effects. As shown in Fig. 4, in the cold nose of the wedge, the stress is about two orders of magnitude larger than in the rest of the mantle wedge. Together with the assumed high water content, it is an indicator that in the region of the backward flow B-fabric olivine may exist (Lassak *et al.* 2006; Kneller *et al.* 2007). Lattice preferred orientation of B-type olivine is thought to be the reason why these forearc regions show trench-parallel fast seismic direction (unlike this, in the regions farther from the trench the fast direction axis is roughly subduction-parallel), and we may hypothesize that in the regions, where fast seismic direction is trench-parallel, the backward flow occurs.

Eq. (3), describing the heat transfer, contains the dissipative term $\mathbf{D} : \nabla \mathbf{v}$. Since there is no substantial increase in the temperature in the region of backward flow, where the strain rate is large (see Fig. 4), with the increasing convergence rate, it is obvious that this term is negligible.

4 CONCLUSIONS

The presented model shows that the dip angle of the subducting plate is controlled mainly by the viscosity weakening, here modelled as a drop of the yield stress in the oceanic crust. The viscosity weakening is likely related to the water content in the subducted rocks, and therefore, we may conclude that the dip angle is related to the water content. This is consistent with our earlier findings (Kukačka & Matyska 2004) and could also explain why traditionally searched relationships among the dip angle and other parameters of subduction plates (Jarrad 1986) are statistically unimportant (Cruciani *et al.* 2005; Billen & Hirth 2007). Unlike the classical steady state models (Peacock 1996; van Keken *et al.* 2002), the presented method allows simultaneously to calculate both the flow and temperature fields and to predict the geometrical layout in the subduction zone (i.e. the dip angle). The obtained shape of the subducting plate is consistent with what has been observed in many subducting plates, that is, the dip angle is small in shallow depths, and with the increasing depth, the dip angle increases (Jarrad 1986).

The most important outcome from our modelling is that if there is a weakened portion in the mantle wedge, an upward flow may naturally evolve at the tip of the wedge along the plate–wedge interface. The extent of such a region of backward circulation is comparable to dimensions of wedge-shaped exhumation channels, modelled with kinematically prescribed slabs (Stöckert & Gerya 2005). The presented numerical model is too simple to cover the exhumation processes in their whole complexity; however, the calculated backflow in the wedge may be a part, if not a source, of the exhumation processes. We emphasize that in our model, the body forces are not taken into account, which means that they are likely not necessary for the exhumation of rocks, which is also consistent with earlier numerical models (Gerya *et al.* 2002). The fact that the

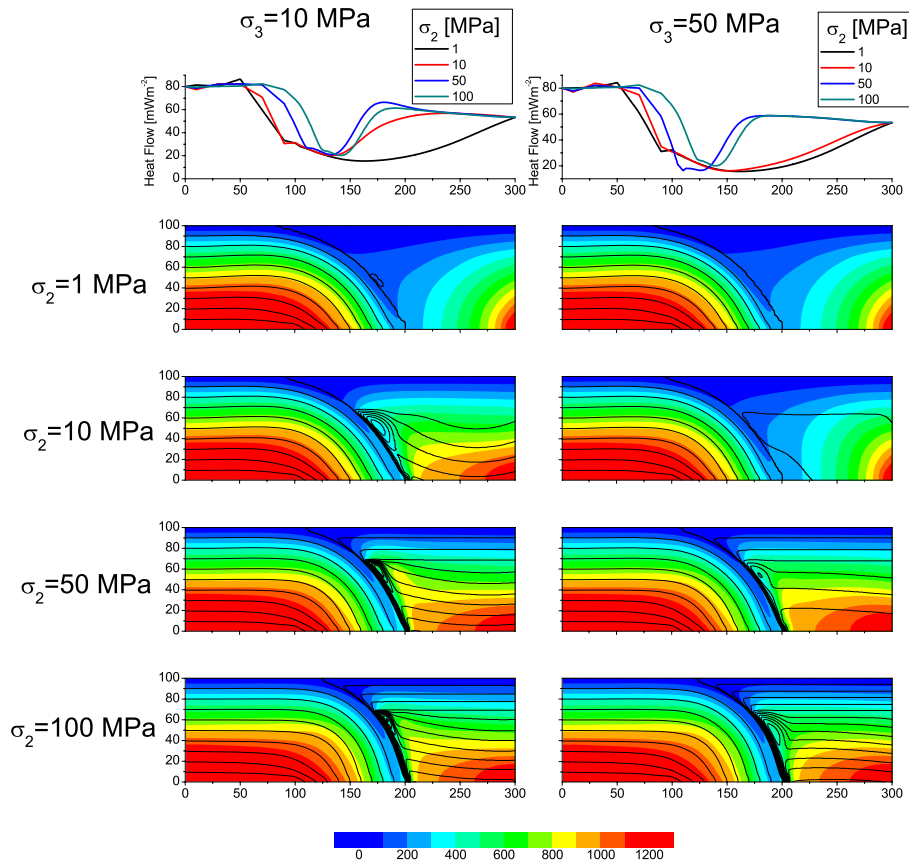


Figure 3. The temperature, the stream function (contour interval $0.1\psi_B$ for $\psi < \psi_B$ and $0.005\psi_B$ otherwise) and the surface heat flow for a subducting plate of rate 5 cm yr^{-1} . Results for two different magnitudes of the yield stress in the cold tip of the wedge (columns) and different magnitudes of the yield stress in the oceanic crust (rows) shown.

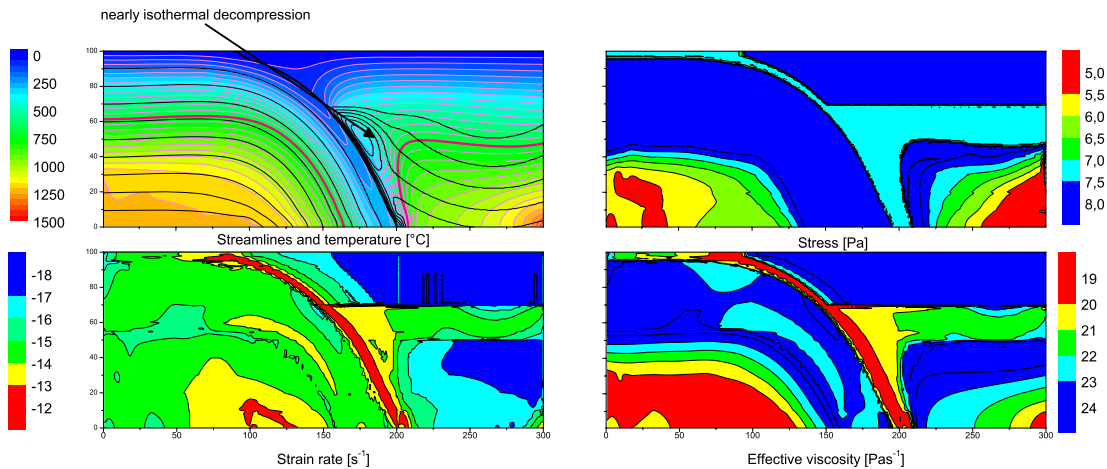


Figure 4. The temperature and the streamlines (top left-hand panel), stress (top right-hand panel), strain rate (bottom left-hand panel) and effective viscosity in the case of $\sigma_2 = \sigma_3 = 10 \text{ MPa}$. Convergence rate is 5 cm yr^{-1} . The isotherm T_{SERP} is marked red. An arrow points to the area within the serpentine stability field, where streamlines of the upward flow and isotherms coincide (i.e. isothermal decompression occurs there).

exhumation from depths of about 80 km to lower crustal levels might be an outcome of the interplay between viscous forces is surprising, but it helps to explain why the exhumation of eclogite–serpentine melange can occur even when serpentinization is only partial (the density of the melange might be very close to the density of surrounding peridotites). In our model, partial serpentinization is not a problem, since mere partial serpentinization results in sufficient strength reduction (Escartin *et al.* 2001).

The backward flow along the plate–wedge interface drastically changes the temperature distribution in the wedge. At the surface, the upward flow results in an increase in surface heat flow. In the case without the backward flow (no weakening in the wedge), the surface heat flow is as low as $15\text{--}20 \text{ mW m}^{-2}$, which is possible but very rarely observed (Grevemeyer *et al.* 2005). If the weakening in the wedge is assumed, the surface heat flow is much higher and can be close to 60 mW m^{-2} , which is above the generally

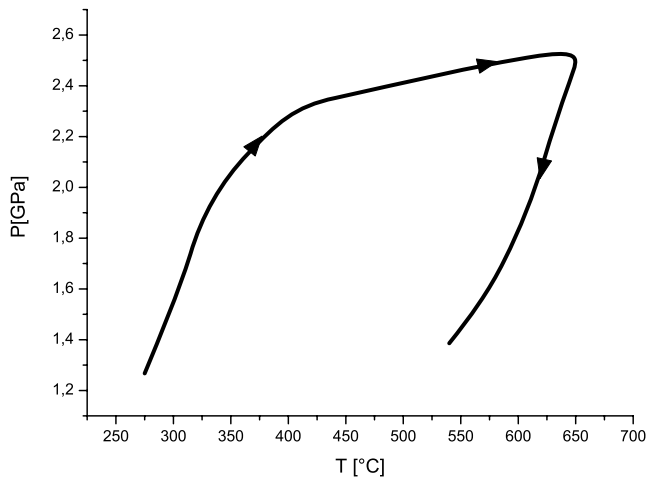


Figure 5. An example of P - T path in the case of $\sigma_2 = \sigma_3 = 10$ MPa (convergence rate 5 cm yr^{-1}) along a streamline marked by the arrow in Fig. 4. The retrograde part of the path shows nearly isothermal decompression.

observed values (Hyndman *et al.* 1995). It means that the vigour of the backward flow is likely smaller than in the presented cases. It is interesting that we did not obtain higher surface heat flow values for any set of free model parameters, which is in agreement with surface heat flow observations (Ziagos *et al.* 1985). It is related to the fact that a stronger backward flow bringing more heat from the interior reduces the size of the weak portion of wedge tip,

which inhibits the upward flow. The presented model demonstrated, however, that when a balance between these two effects is achieved, a nearly isothermal decompression during the exhumation is possible.

Anyway, the results show that the steady state models prescribing a fixed geometry of the subduction zones overlook a significant feature of the forearc system. The oceanic crust must be weak enough to enable a plate-like behaviour of subduction (i.e. plunging the oceanic plate beneath the overriding plate); however, in such a case, the viscous coupling between the oceanic crust and the hydrated and weak tip of the wedge may be sufficient to produce a backward flow. Even if the exhumation is likely an episodic process (Agard *et al.* 2009), the return flow may be a steady state feature in forearcs, and therefore, it should be included into the kinematic subduction models to shed more light onto the forearc thermal state.

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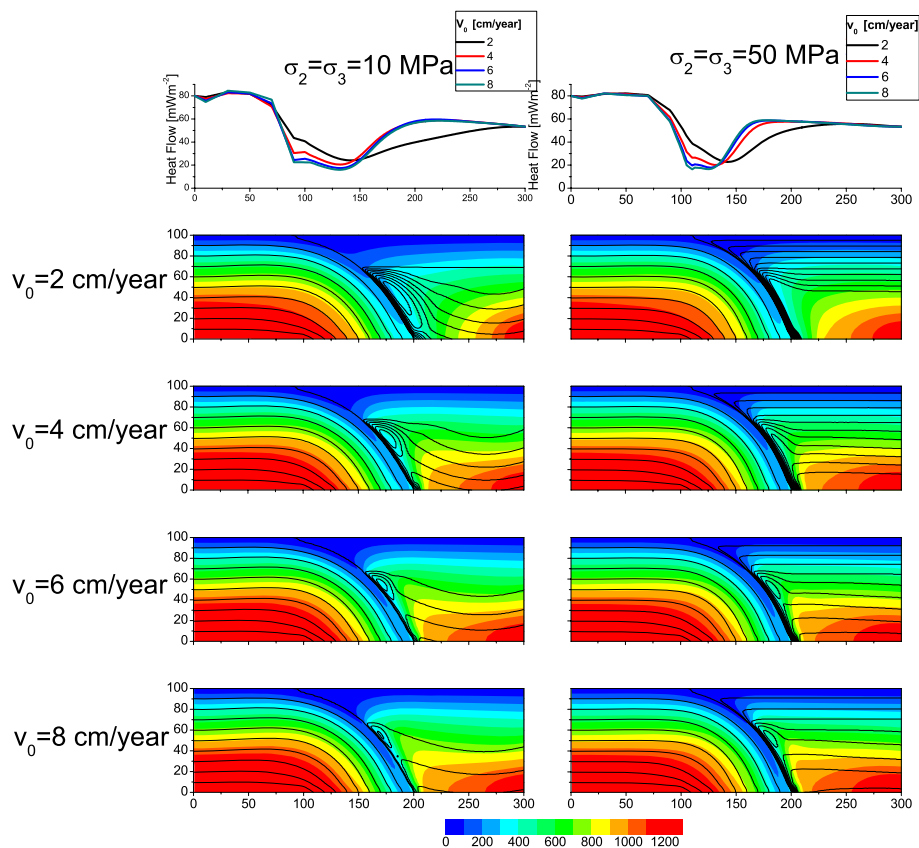


Figure 6. The temperature, the stream function (contour interval $0.1\psi_B$ for $\psi < \psi_B$ and $0.005\psi_B$ otherwise) and the surface heat flow for different convergence rates, ranging from 2 – 8 cm yr^{-1} . Results for two different sets of the yield stress $\sigma_2 = \sigma_3 = 10$ MPa and $\sigma_2 = \sigma_3 = 50$ MPa are shown.

REFERENCES

- Abers, G.A., van Keken, P.E., Kneller, E.A., Ferris, A. & Stachnik, J.C., 2006. The thermal structure of subduction zones constrained by seismic imaging: implications for slab dehydration and wedge flow, *Earth planet. Sci. Lett.*, **241**, 387–397.
- Agard, P., Yamato, P., Jolivet, L. & Burov, E., 2009. Exhumation of oceanic blueschists and eclogites in subduction zones: timing and mechanisms, *Earth-Sci. Rev.*, **92**, 53–79.
- Baldwin, S.L., Monteleone, B.D., Webb, L.E., Fitzgerald, P.G., Grove, M. & Hill, E.J., 2004. Pliocene eclogite exhumation at plate tectonic rates in eastern Papua New Guinea, *Nature*, **431**, 263–267.
- Billen, M.I. & Hirth, G., 2007. Rheologic controls on slab dynamics, *Geophys. Geochem. Geosyst.*, **8**, doi:10.1029/2007GC001597.
- Conder, J.A., 2005. A case for hot slab surface temperatures in numerical viscous flow models of subduction zones with an improved fault zone parametrization, *Phys. Earth planet. Inter.*, **149**, 155–164.
- Cruciani, C., Carmitani, E. & Doglioni, C., 2005. Slab dip vs. lithosphere age: no direct function, *Earth planet. Sci. Lett.*, **238**, 298–310.
- Ernst, W.G., 1988. Tectonic history of subduction zones inferred from retrograde blueschist P-T paths, *Geology*, **16**, 1081–1084.
- Escartin, J., Hirth, G. & Evans, B., 2001. Strength of slightly serpentinized peridotites: implications for the tectonics of oceanic lithosphere, *Geology*, **29**, 1023–1026.
- Gerya, T.V. & Stöckert, B., 2002. Exhumation rates of high pressure metamorphic rocks in subduction channels: the effect of rheology, *Geophys. Res. Lett.*, **29**, doi:10.1029/2001GL014307.
- Gerya, T.V., Stöckert, B. & Perchuk, A.L., 2002. Exhumation of high-pressure metamorphic rocks in a subduction channel: a numerical simulation, *Tectonics*, **21**, 1–15.
- Grevemeyer, I., Kaul, N., Diaz-Naveas, J.L., Villinger, H.W., Ranero, C.R. & Reichert, C., 2005. Heat flow and bending-related faulting at subduction trenches: case studies offshore of Nicaragua and Central Chile, *Earth planet. Sci. Lett.*, **236**, 238–248.
- Guillot, S., Hattori, K.H. & de Sigoyer, J., 2000. Mantle wedge serpentinization and exhumation of eclogites: insight from eastern Ladakh, northern Himalaya, *Geology*, **28**, 199–202.
- Guillot, S., Hattori, K.H., de Sigoyer, J., Nägler, T. & Auzende, A., 2001. Evidence of hydration of the mantle wedge and its role in the exhumation of eclogites, *Earth planet. Sci. Lett.*, **193**, 115–127.
- Herman, J., Müntener, O. & Scambelluri, M., 2000. The importance of serpentine mylonites for subduction and exhumation of oceanic crust, *Tectonophysics*, **327**, 225–238.
- Herms, P., 2004. Fluids in a 2 Ga old subduction zone—deduced from eclogite-facies rocks of the Usagaran belt, Tanzania, *Eur. J. Mineral.*, **14**, 361–373.
- Hirth, G. & Kohlstedt, D., 2003. Rheology of the upper mantle and the mantle wedge: a view from the experimentalist, in *Inside the Subduction Factory*, ed. Eiler, J., Am. Geophys. Union, Washington, DC.
- Hyndman, R.D. & Peacock, S.M., 2003. Serpentinization of the forearc mantle, *Earth planet. Sci. Lett.*, **212**, 417–432.
- Hyndman, R.D., Wang, K. & Yamano, M., 1995. Thermal constraints on the seismogenic portion of the southwestern Japan subduction thrust, *J. geophys. Res.*, **100**, 15 373–15 392.
- Jarrad, R.D., 1986. Relations among subduction parameters, *Rev. Geophys.*, **24**, 217–283.
- Jolivet, L., Raimbourg, H., Labrousse, L., Avigad, D., Leroy, Y., Austrheim, H. & Andersen, T.B., 2005. Softening triggered by eclogitization, the first step toward exhumation during continental subduction, *Earth planet. Sci. Lett.*, **237**, 532–547.
- Kneller, E.A., van Keken, P.E., Katayama, I. & Karato, S., 2007. Stress, strain, and B-type olivine fabric in the fore-arc mantle: sensitivity tests using high-resolution steady-state subduction zone models, *J. geophys. Res.*, **112**, B04406, doi:10.1029/2006JB004544.
- Kukačka, M. & Matyska, C., 2004. Influence of the zone of weakness on dip angle and shear heating of subducted slab, *Phys. Earth planet. Inter.*, **141**, 243–252.
- Manea V.C., Manea, M., Kostoglodov, V. & Sewell, G., 2005. Thermo-mechanical model of the mantle wedge in Central Mexican subduction zone and a blob tracing approach for the magma transport, *Phys. Earth planet. Inter.*, **149**, 165–186.
- Meza-Figueroa, D., Ruiz, J., Talavera-Mendoza, O. & Ortega-Gutierrez, F., 2003. Tectonometamorphic evolution of the Acatlan Complex eclogites (southern Mexico), *Can. J. Earth Sci.*, **40**, 27–44.
- Lassak, T.M., Fouch, M.J., Hall, C.E. & Kaminski, E., 2006. Seismic characterization of mantle flow in subduction systems: can we resolve a hydrated mantle wedge?, *Earth planet. Sci. Lett.*, **243**, 632–649.
- Peacock, S.M., 1996. Thermal and petrologic structure of subduction zones, in *Subduction: Top to Bottom*, *Geophysical Monograph* 96, pp. 119–133, eds Bebout, G., Scholl, D.W., Kirby, H.S. & Platt, J.P., Am. Geophys. Union, Washington, DC.
- Peacock, S.M., Rushmer, T. & Thompson, A.B., 1994. Partial melting of subducting oceanic crust, *Earth planet. Sci. Lett.*, **121**, 227–244.
- Peacock, S.M., van Keken, P.E., Holloway, S.D., Hacker, B.R., Abers, G.A. & Ferguson, R.L., 2005. Thermal structure of the Costa Rica–Nicaragua subduction zone, *Phys. Earth planet. Inter.*, **149**, 187–200.
- Platt, J.P., 1993. Exhumation of high-pressure rocks: a review of concepts and processes, *Terra Nova*, **5**, 119–133.
- Raimbourg, H., Jolivet, L. & Leroy, Y., 2007. Consequences of progressive eclogitization on crustal exhumation, a mechanical study, *Geophys. J. Int.*, **168**, 379–401.
- Schwartz, S., Allemand, P. & Guillot, S., 2001. Numerical model of the effect of serpentinites on the exhumation of eclogitic rocks: insight from the Monviso ophiolitic massif (Western Alps), *Tectonophysics*, **342**, 193–206.
- Stöckert, B. & Gerya, T.V., 2005. Pre-collisional high pressure metamorphism and nappe tectonics at active continental margins: a numerical simulation, *Terra Nova*, **17**(2), 102–110.
- Tsujimori, T., Sisson, V.B., Liou, J.G., Harlow, G.E. & Sorensen, S.S., 2006a. Very-low-temperature record of the subduction process: a review of worldwide lawsonite eclogites, *Lithos*, **92**, 609–624.
- Tsujimori, T., Matsumoto, K., Wakabayashi, J. & Liou, J.G., 2006b. Franciscan eclogite revised: reevaluation of the P–T evolution of tectonic blocks from Tiburon Peninsula, California, U.S.A., *Mineral. Petrol.*, **88**, 243–267.
- van Keken P.E., Kiefer, B. & Peacock S.M., 2002. High-resolution models of subduction zones: implications for mineral dehydration reactions and the transport of water into the deep mantle, *Geochem. Geophys. Geosyst.*, **3**, Q05017, doi:10.1029/2001GC000256.
- van Keken, P.E. et al., 2008. A community benchmark for subduction zone modeling, *Phys. Earth planet. Inter.*, **171**, 187–197.
- Ulmer, P. & Trommsdorf, V., 1995. Serpentine stability to mantle depths and subduction-related magmatism, *Science*, **268**, doi:10.1126/science.268.5212.858.
- Ziagos, J.P., Blackwell, D.D. & Moser, F., 1985. Heat flow in Southern Mexico and the thermal effects of subduction, *J. geophys. Res.*, **90**, 5410–5420.