Tectonophysics 494 (2010) 149–154

Contents lists available at ScienceDirect

Tectonophysics

journal homepage: www.elsevier.com/locate/tecto

Short Communication

Fluid flow during unbending: Implications for slab hydration, intermediate-depth earthquakes and deep fluid subduction

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ARTICLE INFO

Article history:
Received 10 June 2010
Received in revised form 12 August 2010
Accepted 13 August 2010
Available online 20 August 2010

Keywords:
Tectonic stress profile
Slab unbending
Deep water cycle
Rheological weakening
Deep earthquakes

ABSTRACT

We calculate the tectonic stress profile and associated direction of fluid flow during unbending and dehydration of oceanic plates, for a range of critical parameters that affect their combined elasto-plastic and viscous behaviour, such as bending curvature, age, pore fluid pressure and viscous flow laws. In all models, negative pressure gradients are established at Moho depths, down to the base of the slab elastic core. Fluids released at these depths flow downward across the plate, "wetting" or further hydrating the underlying dry levels, and ultimately accumulate at the base of the elastic core, increasing the pore fluid pressure and triggering deep seismicity. The thickness of the "wet" layer increases for low bending curvatures, low pore fluid pressure, old slabs and dry viscous rheologies. "Wetting" of the upper 10–30 km of the slab has important implications for its rheological and anisotropic structure. Unbending favours the redistribution and trapping of significant amounts of fluids in subducting oceanic plates that will subsequently be released at the base of the upper mantle.

1. Introduction

Deformation of the oceanic plate at the mid ocean ridge and at the trench outer rise induces percolation of seawater into the slab lithosphere down to mantle depths (Ranero et al., 2003, 2004; Grevemeyer et al., 2007; Contreras-Reyes et al., 2008; Faccenda et al., 2009), leading to the hydration of dry mafic and ultramafic rocks of the oceanic crust and mantle, respectively. Slab hydration implies the formation of hydrous minerals such as amphibole, chlorite, serpentine and talc that are able to store significant amounts of chemically bound water in their crystalline lattice (roughly around 2–12 wt.%, Schmidt and Poli, 1998). During subduction and warming of the slab, such hydrous phases become progressively unstable at temperatures above 600–800 °C and dehydration reactions take place at depths between 50 and 200 km, triggering the observed intermediate-depth seismicity (e.g., Peacock, 2001; Hacker et al., 2003; Brudzinski et al., 2007). If the slab is sufficiently cold, however, hydrous mantle rocks can transform into Dense Hydrous Magnesium Silicate (DHMS) phases (such as phase A; Schmidt and Poli, 1998), with the consequence that chemically bound water can be transported down to the base of the upper mantle (Rüpke et al., 2004).

Buoyant fluids produced by the breakdown of hydrous minerals are thought to flow upward along different paths, such as high permeability reaction zones, cracks, faults and shear zones oriented according to the slab thermal structure and state of stress (Hacker et al., 2003), leaving behind a nearly dry slab. However, at depths where the pore fluid pressure approaches the solid pressure, fluid flow is driven not only by the fluid buoyancy but also by the spatial variation in tectonic stress causing deformation (e.g., Connolly and Podladchikov, 2004; Phipps Morgan and Holtzman, 2005; Katz et al., 2006). In particular, it has been recognized that the gradient in tectonic pressure (rather than its absolute magnitude) is fundamental for driving many processes of fluid flow in porous rocks (Mancktelow, 2008). For instance, Faccenda et al. (2009) showed that at the trench outer rise a strong variation of the tectonic under-pressure may induce the formation of sub-hydrostatic or even inverted solid pressure gradients that favour downward fluid flow in the subducting slab.

Slab unbending also induces strong gradients in tectonic pressure (e.g., Babeyko and Sobolev, 2008) that may significantly affect the fluid flow during slab dehydration at intermediate depths. In this paper we investigate how fluid flow is affected by stresses arising from slab unbending and show that the transition from tectonic overpressure in the oceanic crust to tectonic under-pressure focused in the slab mantle core favours downward fluid flow and fluid trapping inside the slab.

2. Plate flexure model

The lithospheric state of stress of a plate undergoing bending and subsequent unbending (Fig. 1) is modelled using the method and parameters described in Mueller et al. (1996a). The in-plane differential stresses (σ'_{xx}) resulting from flexure of the oceanic lithosphere are bounded by a yield surface representing perfectly plastic (i.e., brittle) deformation in the upper part and (generally non-linear) viscous deformation in the lower part. Stresses below yield in the brittle part are

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doi:10.1016/j.tecto.2010.08.002

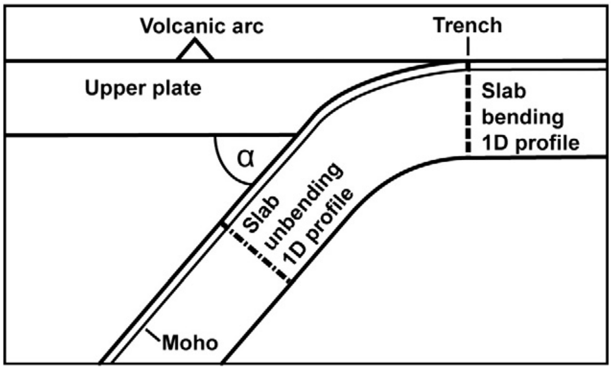

Fig. 1. Schematic representation of a subduction zone showing the location of the 1D profiles where the lithospheric states of stress of the oceanic plate during bending and unbending were computed.

maintained elastically. Since the initial in-plane force is assumed to be zero (no significant regional stresses), the differential stresses are iteratively adjusted to obtain a null stress integral. Despite the simplicity of such one-dimensional models, which are computed statically and assume a constant strain rate through the entire plate, previous studies have successfully reproduced the lithospheric state of stress (reflected in patterns of seismicity) of the oceanic lithosphere at the outer rise (Chapple and Forsyth, 1979; Mueller et al, 1996b) and in the unbending area (Wang, 2002). In our models, the curvature of the bent slab is in the range of 1–5$\times 10^{-6}$ m and is assumed to have been fully unbent at a depth of 100 km (i.e. the slab below this depth is straight).

2.1. Slab rheology

The 1-D temperature profile of a 90 km thick oceanic lithosphere with a 7 km thick crust is calculated using the cooling plate model of Carslaw and Jaeger (1984), which allows for a better description of plates older than 70–80 Ma. The thermal structure used during unbending is the same as during bending. Brittle behaviour at shallow depths is governed by Byerlee's (1978) law of rock friction, with fluid pressure lowering the effective normal stress by a factor $(1-\lambda)$, where λ is the pore fluid factor (ratio of fluid to solid pressure). At greater depths and temperatures, viscous creep is activated. Several recent studies have concluded that at low-T and high stresses an exponential (Peierls) creep may be active, producing most of the observed deep seismicity (e.g., Karato et al., 2001). Hence, viscous deformation is modelled considering both power-law dislocation creep (Mueller et al., 1996a) and exponential creep (Katayama and Karato, 2008) mechanisms. Also, because of the presence of fluids, pore fluid pressures and flow laws are varied to simulate either dry or wet conditions. Since fluids are thought to penetrate into the upper part of the slab during bending (Ranero et al., 2003, 2004; Grevemeyer et al., 2007; Contreras-Reyes et al., 2008; Faccenda et al., 2009), the pore fluid pressure gradient is assumed to be hydrostatic ($\lambda = 0.3$, Wang, 2002) in the brittle layer, while dry flow laws are used for the viscous deformation of the lower section. On the other hand, in the unbending area the pore fluid pressure gradient is supposed to approach the lithostatic pressure gradient in the dehydrating portions of the slab. Following Wang (2002), we use $\lambda = 0.95$ for the crust and $\lambda = 0.8$ for the mantle to account for a higher degree of hydration of the crust. We do not account for extensional stresses arising from oceanic crust eclogitization because, in our models, unbending strain rates are markedly higher than the ones caused by metamorphism. Both dry and wet viscous flow laws are used for comparison during unbending.

2.2. Fluid flow

The fluid flow normal (z coordinate positive downward) to the slab surface is calculated using a simple Darcy's law where, considering the range of depths (50–200 km) at which slab dehydration takes place, we reasonably assume that the pore fluid pressure (P_f) is equal to the solid pressure (P_s). The Darcy's law is then:

$$v_z = -\frac{k}{\eta_f \phi}\left(\frac{\partial P_s}{\partial z} - \rho_f g_Z \cos\alpha\right) \quad (1)$$

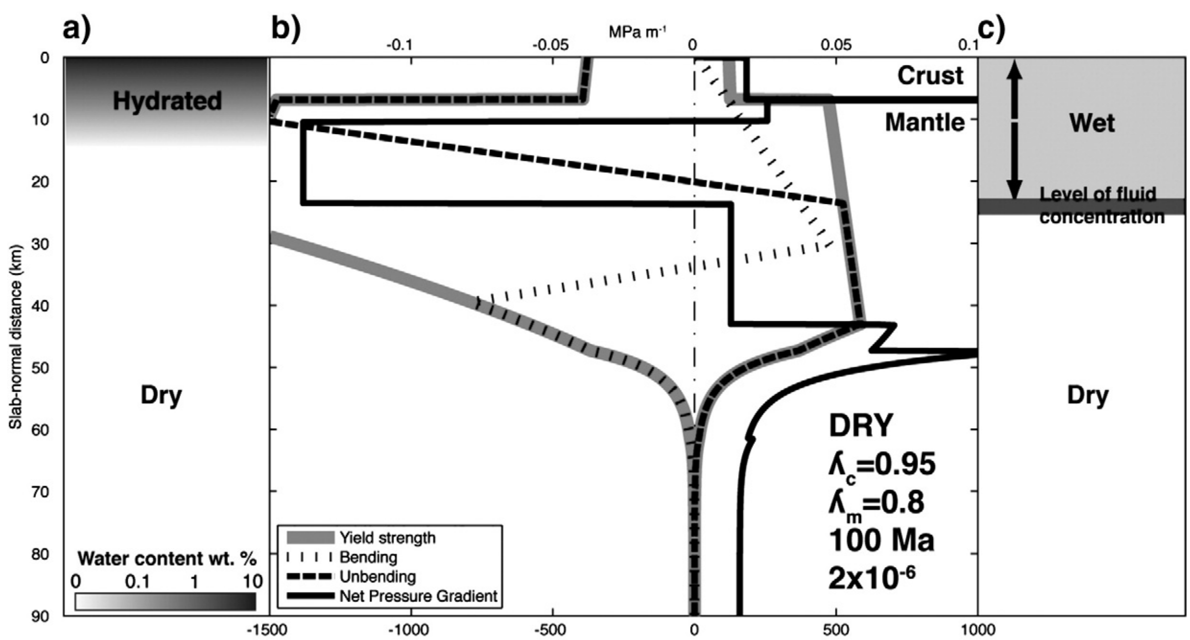

Fig. 2. a) Schematic, slab-normal section showing the water concentration in a subducting oceanic plate before dehydration. Hydration is high in the crust and decreases with depth. The colour scale roughly indicates the water concentrations. b) Stress profiles due to bending and unbending (lower horizontal axes) and net pressure gradient (upper horizontal axes) during unbending for a 100 Ma old slab using dry viscous flow laws and assuming a maximum bending curvature of 2×10^{-6} m^{-1}, $\lambda_c = 0.95$ for the crust and $\lambda_m = 0.8$ for the mantle. c) Schematic, slab-normal section showing the water concentration in a subducting oceanic plate after dehydration. The black arrows indicate the direction of fluid flow. The upper layer remains "wet" (1000–2000 wt.ppm of water) while fluids concentrate at the base of the elastic core.

where v_z is the Darcy's velocity component normal to the slab surface, k and ϕ are the permeability and porosity of the medium, η_f and ρ_f are the fluid viscosity and density, g_Z is the vertical component of the gravity acceleration and α is the slab dip (45° for all models presented here). For the purpose of this study, we are more interested in which direction the fluid flows once it has been released in the unbending area than on the rate at which this flow occurs. The downward or upward direction ($d_{z\pm}$) of fluid flow normal to the slab is opposite to the sign of the net pressure gradient (∇P_z):

$$d_{z\pm} = -sgn(\nabla P_z) \quad (2)$$

$$\nabla P_z = \frac{\partial P_s}{\partial z} - \rho_f g_Z \cos\alpha \quad (3)$$

Hence, downward fluid flow (positive $d_{z\pm}$) occurs when the net pressure gradient is negative. From the force balance equation, the solid pressure gradient along the slab-normal is:

$$\frac{\partial P_s}{\partial z} = \rho_s g_Z \cos\alpha - \frac{\partial \sigma'_{xx}}{\partial z} \quad (4)$$

where σ'_{xx} is the in-plane tectonic stress (positive under extension and negative under compression) resulting from the bending and unbending of the slab, and ρ_s is the solid density. The first term on the right of Eq. (4) is the isotropic component due to body forces (the lithostatic gradient) and the second is a tectonic component due to the deviatoric stress (the tectonic stress gradient). Substituting Eq. (4) into Eq. (3) yields:

$$\nabla P_z = \Delta\rho g_Z \cos\alpha - \frac{\partial \sigma'_{xx}}{\partial z} \quad (5)$$

where $\Delta\rho$ is the density contrast between the solid and fluid phases.

3. Results

Fig. 2 shows (a) a schematic section of the water concentration in a subducting oceanic plate before dehydration, (b) the differential stress and net pressure gradient profiles of our reference model due to bending and unbending, and (c) a schematic section of the water concentration in a subducting oceanic plate after dehydration. Slab bending induces extension and tectonic under-pressure in the upper part of the oceanic plate and compression and tectonic overpressure in the core. Conversely, slab unbending induces tectonic overpressure in the upper part of the slab and tectonic under-pressure in the core (e.g., Babeyko and Sobolev, 2008). The net pressure gradient (Eq. (5)) during unbending is positive at the top and bottom of the profile where the plate yields, but negative in the elastic slab core close to the neutral surface marking the transition from compressional differential stresses above to extensional differential stresses below. It follows that if the crust and shallow lithospheric mantle are dehydrating

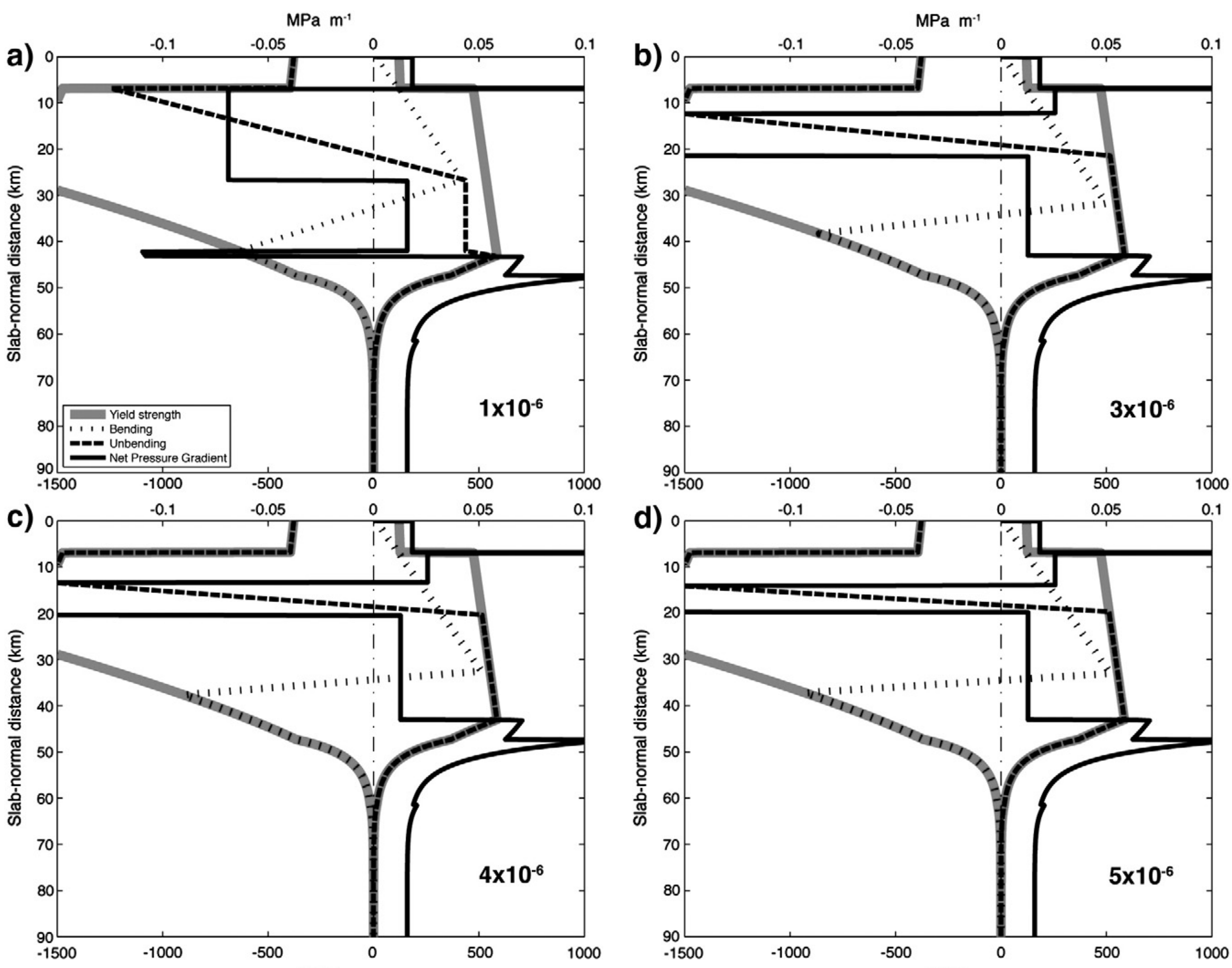

Fig. 3. Stress profiles due to bending and unbending (lower horizontal axes) and net pressure gradient (upper horizontal axes) during unbending for slabs with different maximum bending curvatures respect to the reference model (Fig. 2b).

during unbending, fluids released in the uppermost part of the slab will be expelled upward toward the mantle wedge, whereas those produced at the base of the crust or in the underlying mantle will migrate downward and concentrate in the region undergoing extensional deformation (Fig. 2c).

We carried out the same calculations as for the reference model changing parameters such as bending curvature, slab age, pore fluid pressure and viscous creep flow laws (Figs. 3–5). In all models, negative net pressure gradients form at the bottom of the crust and extend down into the mantle for 5–20 km, depending on the slab rheological structure and bending curvature. Deep downward fluid flow is established for low bending curvatures (Fig. 3), old plates (Fig. 4), low pore fluid pressure (Fig. 5a–c) and dry viscous creep flow laws (Fig. 5d). The thickness of the elastic core (i.e., where downward fluid flow is established) is reduced when yielding is more widespread through the plate because of high bending curvatures (Fig. 3), for "hot" thermal structures of the slab (Fig. 4), high pore fluid pressures (Fig. 5a–c) and wet viscous rheologies (Fig. 5d). The slab dip does not have a marked effect on the extent to which fluid may flow downward inside the slab because the negative net pressure gradients are O (10^{-1}–10^{0}) MPa m^{-1} while the buoyancy term in Eq. (5) (that induces upward fluid flow) is O ($\leq 10^{-2}$) MPa m^{-1}.

4. Discussion and implications

The results shown in Figs. 2–5 should be seen as snapshots of an evolving system whose time-integrated behaviour is the sum of the single instantaneous solutions. In fact, while subducting, the bending curvature and age of oceanic plates will in general change through time and, in particular, the rheological structure may also change as fluids migrate through, and react with, the solid matrix. Despite this, the different 1D profiles show that during slab unbending downward fluid flow should occur in all subduction settings and promote hydration of the oceanic plate to at least Moho depths, because the variation in the tectonic stresses across the elastic core of the plate has a stronger and opposite effect compared to that of fluid buoyancy. Indeed, tomographic images show that subducting oceanic plates are hydrated down to mantle depths in many subduction settings (Ranero and Sallares, 2004; Grevemeyer et al., 2007; Contreras-Reyes et al., 2008), which would support our models where fluids released from the lower crust and uppermost mantle migrate and progressively concentrate at the bottom of the elastic core, leaving behind a "wet" slab (water content: 1000–2000 wt. ppm, Mainprice, 2007).

Hydration or "wetting" of dry mafic and ultramafic rocks induces considerable rheological weakening (1–3 orders of magnitude reduction of the viscosity; Karato and Wu, 1993; Hilairet et al., 2007) that may favour slab deformation, especially at the upper to lower mantle discontinuity. Also, low temperature, high-stress deformation experiments on wet olivine have shown that shear instabilities form at much lower differential stresses when compared to dry counterparts (Katayama and Karato, 2008), implying that deep earthquakes could nucleate in the cold and "wet" slab core more easily than previously thought. Furthermore, such "wet" rocks would

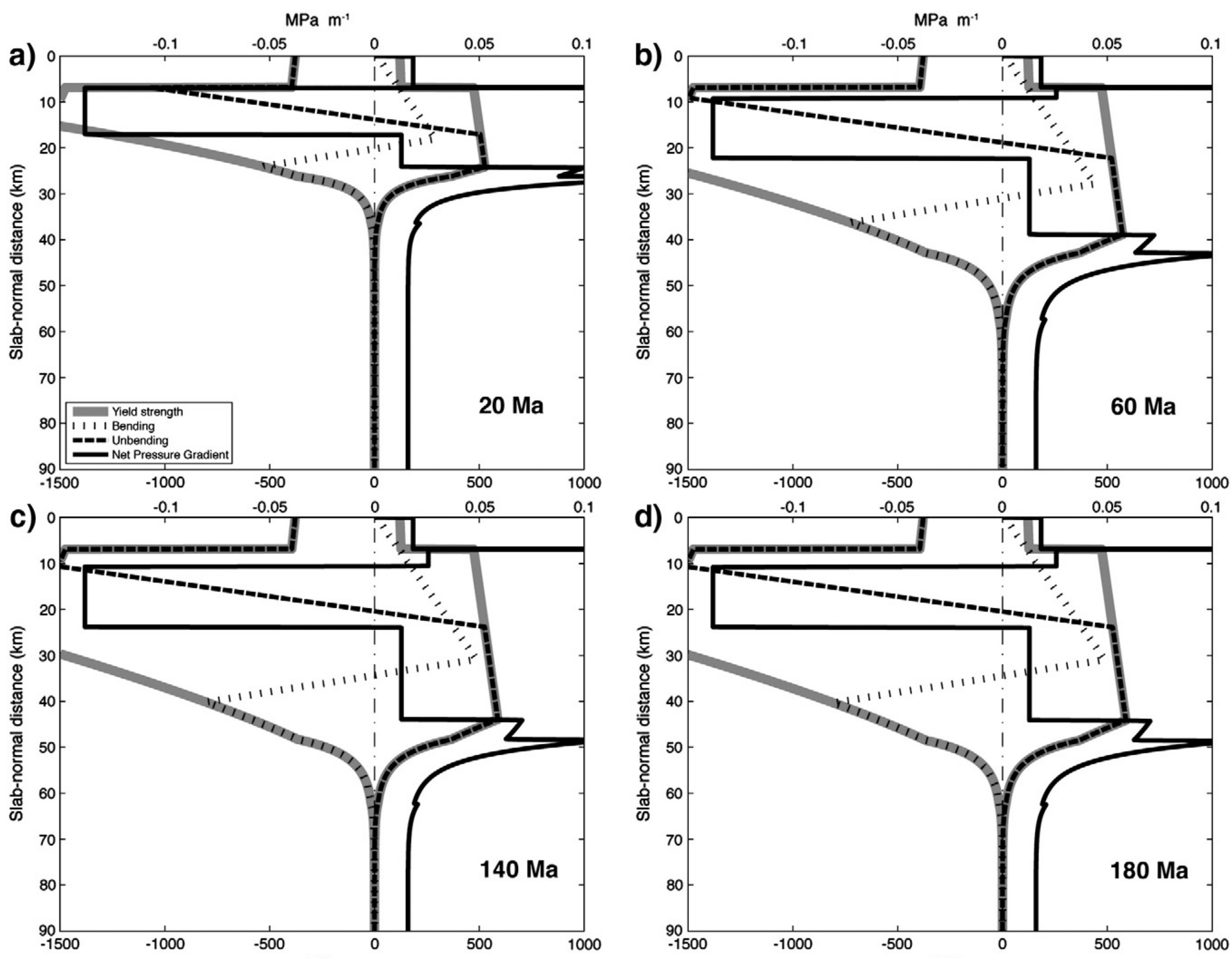

Fig. 4. Stress profiles due to bending and unbending (lower horizontal axes) and net pressure gradient (upper horizontal axes) during unbending for slabs with different ages respect to the reference model (Fig. 2b).

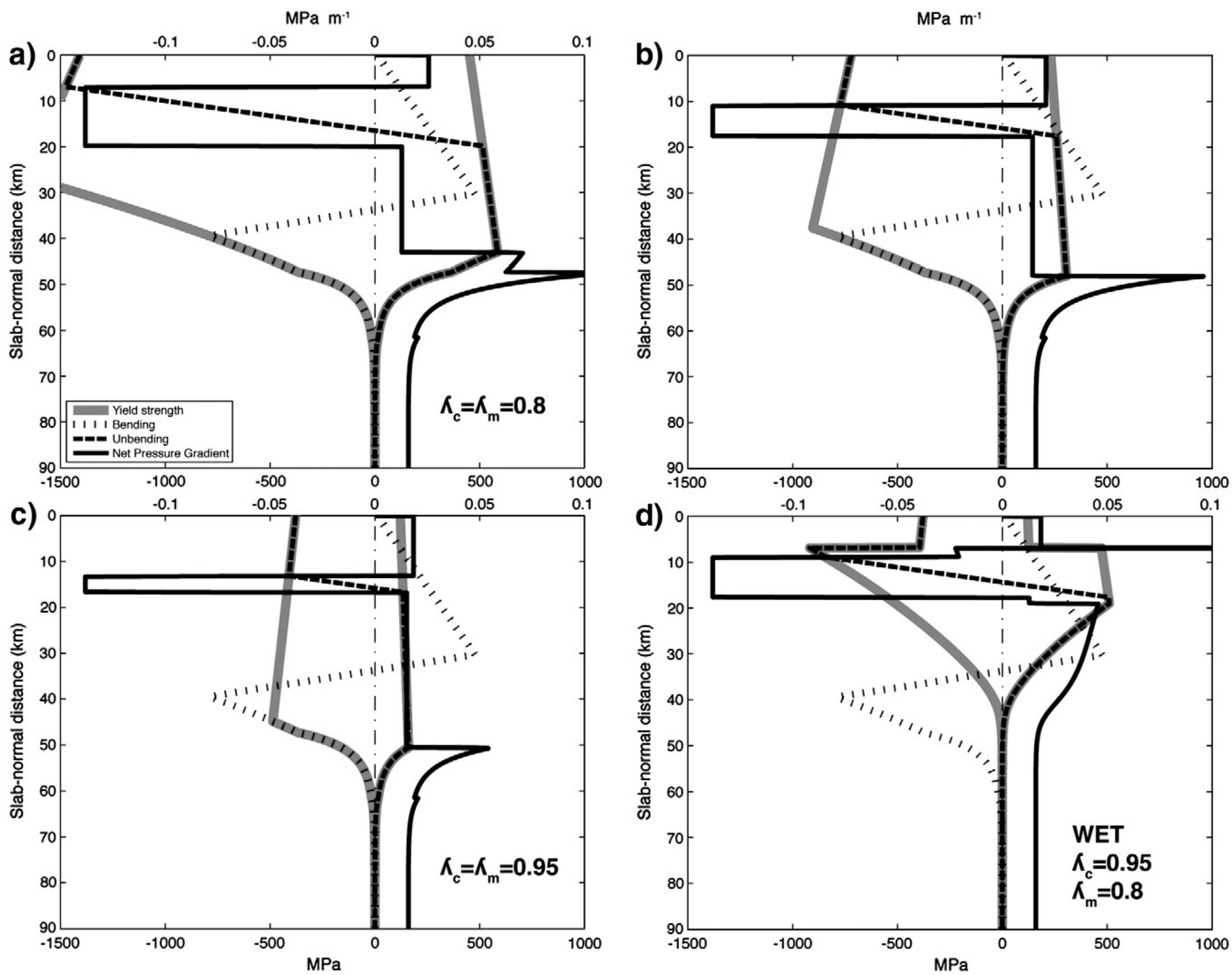

Fig. 5. Stress profiles due to bending and unbending (lower horizontal axes) and net pressure gradient (upper horizontal axes) during unbending for slabs with different pore fluid pressure distribution (a–c) or viscous flow law used during unbending (d) respect to the reference model (Fig. 2b).

contain many times the concentration of water necessary for a C-type crystallographic preferred orientation (CPO) to develop (Jung and Karato, 2001; Mainprice, 2007), producing a different anisotropic fabric compared to the original A-type CPO in the upper 5–20 km of the slab mantle.

On the other hand, fluids trapped at the bottom of the elastic core could subsequently flow up-dip along the slab, producing a low seismic velocity layer (Abers, 2005) and triggering earthquakes due to the elevated pore fluid pressure (e.g., Jung et al., 2004). Indeed, downward fluid flow during unbending is an efficient mechanism for explaining intermediate-depth earthquakes nucleating in the lower plane of Double Benioff Zones (DBZs) up to 30 km thick (e.g., in NE Japan; Igarashi et al., 2001). For instance, recent high-resolution tomographic studies of the mantle beneath Hokkaido, Japan, revealed two layers of low seismic velocity in the subducting Pacific slab corresponding to two planar zones of strong seismicity, with the lower one having low Vp/Vs values typical of a peridotite containing free fluids (Zhang et al., 2004; Nakajima et al., 2009). In the light of our results, the lower plane of the DBZ could more likely reflect the layer where fluids accumulate due to downward fluid flow during unbending rather than the location where dehydration of rocks hydrated at the trench outer rise occurs (e.g. Peacock, 2001). Recently, it has been found that an analogue stagnant zone where fluids concentrate could also form at mid-crustal levels during compressional tectonics in the continental crust (Connolly and Podladchikov, 2004).

Another important implication of our results is that only part of the chemically bound water released at intermediate depths will flow upward toward the mantle wedge and, ultimately, be recycled back to the Earth's surface. Instead, a considerable portion of water will be incorporated in mafic and ultramafic rocks and transported down to the base of the upper mantle, either as a free water component in the crust and underlying mantle (10–30 km of "wet" slab with 1000–2000 wt. ppm of water, Mainprice, 2007) or as bound water at the base of the elastic core, where the concentration of fluids maybe high enough to enable the formation of DHMS phases (water content: 3–18 wt.%, Ohtani, 2005) (Fig. 2c). Despite the fact that with the present methodology we cannot give an exact estimate, the simple 1D models do imply that significant amounts of water may be redistributed in the subducting oceanic plate due to downward fluid percolation induced by slab unbending and this water can be subsequently transported into the transition zone.

Together with geophysical observations (e.g., Ranero et al., 2003, 2004; Grevemeyer et al., 2007) and numerical models (Faccenda et al., 2009) considering bending-related slab hydration, these results suggest a simple model where tectonic pressure markedly affects the water circulation in subducting oceanic plates. During extensional brittle deformation at the trench outer rise, inverted pressure gradients form along active normal faults forcing seawater to flow from the upper and fluid-filled oceanic crust downward to mantle depths, where it becomes chemically bounded in hydrous phases (Faccenda et al., 2009). During unbending, fluids produced by the breakdown of hydrous minerals will migrate away from the region of maximum tectonic overpressure (close to the Moho), which acts as a "watershed". Fluids produced in the oceanic crust will migrate upward into the mantle wedge, whereas those produced in the lithospheric mantle will migrate downward toward an extending region with strong tectonic under-pressure. Subsequently, the water trapped in the slab mantle will migrate up-dip along a level of neutral (slab-normal) pressure gradient and be progressively consumed by reactions with the relatively dry rocks through which it flows.

5. Conclusions

Using simple 1D models of visco-elasto-plastic plates, we calculated differential stress profiles and the associated fluid flow direction of a dehydrating slab undergoing bending and unbending and showed that part of the fluids released via dehydration reactions at intermediate depths can be retained in the slab. Unbending induces strong and inverted tectonic pressure gradients at Moho depths that force the fluids to flow downward across the slab. According to our interpretation of the models, fluids released by the dehydrating lithospheric mantle will tend to accumulate at the bottom of the elastic core, gradually increasing the pore fluid pressure and triggering deep seismicity. In this way, the lower plane of the DBZ could form without requiring extremely deep (30 km for old plates such as the Pacific slab under NE Japan) slab hydration at the trench outer rise. Depending on the slab age, pore fluid pressure, curvature and composition (i.e., dry or wet rheology), the upper 10–30 km of the subducting slab should be considered as a "wet" layer. "Wetting" of such a layer implies (1) strong weakening of the cold slab core where shear instabilities (i.e., deep earthquakes) could nucleate, (2) a change of the anisotropic fabric (CPO) initially acquired at the mid ocean ridge, and (3) the incorporation of a significant amount of water that could be carried down to the transition zone.

Acknowledgements

This work was supported by Swiss National Science Foundation Fellowship PBEZP2-126187 "Slab deformation dynamics: seismic cycle, deep earthquakes and spontaneous slab hydration" to Manuele Faccenda. We thank Taras Gerya for discussion.

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