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On the extent of mantle hydration caused by plate bending

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ABSTRACT

When bent at subduction zones, oceanic plates are damaged by normal faulting, and this bending-related faulting is widely believed to cause deep mantle hydration, down to ∼20–30 km deep. The buoyancy of water (or equivalently, confining pressure), however, makes it difficult to bring water down even if faulting is deep. Extension associated with plate bending generates negative dynamic pressure, but the magnitude of such dynamic pressure is shown to be insufficient to overcome confining pressure. Seismic velocity anomalies that have been used to infer the extent of mantle hydration are reviewed, and it is suggested that small crack-like porosities, which can be produced by thermal cracking and further enhanced by bending-related faulting, is sufficient to explain such velocity anomalies. The presence of such porosities, however, does not necessarily lead to the substantial hydration of oceanic plates because of confining pressure. Whereas the depth extent of bending-generated porosities is uncertain, the theory of thermal cracking can be used to place a lower bound on the amount of water contained in the slab mantle (0.03–0.07 wt% H_2O), and this lower bound is suggested to be more than sufficient to explain the lower-plane earthquakes of the double seismic zone by dehydration embrittlement.

1. Introduction

Oceanic plates are often thought to be deeply faulted when they are bent at subduction zones (Christensen and Ruff, 1988; Seno and Yamanaka, 1996; Ranero et al., 2003). The alteration of an oceanic plate by seawater takes place as soon as the plate forms at a mid-ocean ridge, but bending at a subduction zone could potentially hydrate the plate down to the depths of several tens of kilometers if faulting is correspondingly deep (Peacock, 2001). The extent of hydration in subducting plates is important for the deep water cycle (e.g., Rüpke et al., 2004), and quite a few observational and theoretical studies on this issue have been published in recent years (e.g., Grevemeyer et al., 2007; Faccenda et al., 2008, 2009; Van Avendonk et al., 2011; Garth and Rietbrock, 2014; Naif et al., 2015).

Water is, however, buoyant with respect to silicate rocks, and given the magnitude of confining pressure (i.e., the difference between lithostatic and hydrostatic pressures), which increases at the rate of ∼23 $\mathrm{MPa\,km^{-1}}$, it is not obvious how deeply water can infiltrate even when faulting is deep. Peacock (2001) suggested that downward water transport might be possible by seismic pumping (Sibson et al., 1975), which is based on the dilatancy-diffusion hypothesis for shallow earthquakes (Scholz et al., 1973). As the validity of the dilatancy-diffusion hypothesis is questionable in a number of aspects (e.g., Main et al., 2012), it has become difficult to defend the original seismic pumping mechanism. Instead of dilatancy, tectonic deformation may be able to generate a sufficient hydraulic gradient to allow downward water transport (e.g., McCaig, 1988), and based on numerical modeling, Faccenda et al. (2009) suggested that plate bending could yield strongly negative 'tectonic' pressure that promotes deep mantle hydration. As shown in this paper, however, the generation of such negative pressure may be in direct conflict with the dynamics of brittle deformation.

Observational efforts to constrain the extent of hydration have been notable particularly in the field of active-source seismology (e.g., Ranero and Sallarès, 2004; Grevemeyer et al., 2007; Van Avendonk et al., 2011), with conclusions invariably in favor of the deep hydration of incoming plates. The interpretation of estimated seismic velocity structure in these studies, however, is not unique and seems to have overlooked an important complication caused by the topology of porosity (e.g., Korenaga et al., 2002). A 10% reduction in the P-wave velocity of mantle rocks, for example, can be caused by ∼20% serpentinization (equivalent to ∼2 wt% H_2O) or by ∼0.1% of crack-like residual porosity (equivalent to ∼0.03 wt% H_2O).

The purpose of this paper is three-fold: (1) derive a theoretical bound on dynamic pressure caused by brittle deformation, (2) examine the degree of nonuniqueness associated with the interpretation of crustal and mantle seismic velocities, and (3) estimate the likely extent of mantle hydration by assembling relevant geo-

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physical observations. In doing so, I will also discuss the potential role of thermal cracking (Korenaga, 2007b) in mantle hydration as well as the origin of intermediate-depth earthquakes. I will start with theoretical considerations on dynamic pressure.

2. Dynamic pressure in the brittle regime

In two-phase flow, fluid pressure and solid pressure can be different owing to surface tension and matrix compaction (McKenzie, 1984; Bercovici and Ricard, 2003), but under the simplifying assumptions of zero surface energy and constant porosity, these pressures are equal (e.g., Spiegelman and McKenzie, 1987; Faccenda et al., 2009). Fluid flow is driven by buoyancy as well as dynamic pressure gradients, and to enable downward water transport, the effect of dynamic pressure, which is caused by the deformation of the solid phase, should be greater than that of buoyancy. In the limit of static or steady-state faulting, the magnitude of dynamic pressure associated with bending-related (normal) faulting can be estimated by simple force balance, as shown below.

Assuming that the stresses in the x, y, and z directions are the principal stresses and that no strain in the y direction, the stress state appropriate for normal faulting owing to horizontal extension in the x direction may be expressed as (e.g., Turcotte and Schubert, 1982)

$$\sigma_{xx} = -\rho g z + \Delta\sigma_{xx}, \tag{1}$$

$$\sigma_{yy} = -\rho g z + \nu \Delta\sigma_{xx}, \tag{2}$$

$$\sigma_{zz} = -\rho g z, \tag{3}$$

where ρ is density, g is gravitational acceleration, and z is depth, $\Delta\sigma_{xx}$ is tensional deviatoric stress, and ν is Poisson's ratio. The deviatoric stress that can be supported by a fault with dip β is given by

$$\Delta\sigma_{xx} = \frac{2\mu(1-\lambda)\rho g z}{\sin 2\beta + \mu(1-\cos 2\beta)}, \tag{4}$$

where μ is the friction coefficient, and λ is pore fluid pressure normalized by lithostatic pressure ($\rho g z$). Dynamic pressure corresponding to normal faulting may be thus written as

$$\begin{aligned} p &= -\frac{1}{3}(\sigma_{xx} + \sigma_{yy} + \sigma_{zz}) - \rho g z \\ &= -\frac{1+\nu}{3}\,\frac{2\mu(1-\lambda)\rho g z}{\sin 2\beta + \mu(1-\cos 2\beta)}. \end{aligned} \tag{5}$$

As shown in Fig. 1a, the dynamic pressure can fully compensate for confining pressure only when the fault dip is very low (~10–20°), which is much lower than the optimal dip for normal faulting (~60°). Outer rise earthquakes with normal faulting exhibit dip angles close to the optimal value (Christensen and Ruff, 1988), indicating that the associated dynamic pressure can reduce confining pressure by ~50% at most. The dynamic pressure is less negative for lower friction coefficient or higher pore fluid pressure, and the case of $\mu = 0.8$, $\lambda = 0$, and $\nu = 0.25$ shown in Fig. 1a is likely to serve as the lower bound. The λ value is ~0.3 when pore fluid pressure is hydrostatic, and higher pore fluid pressure acts to further decrease the magnitude of dynamic pressure. Equation (5) holds for all depths as long as brittle deformation takes place, so normal faulting does not reduce confining pressure sufficiently to allow downward water transport.

The above calculation is based on a static or steady-state force balance, but the consideration of a more dynamic situation, e.g., rupture propagation, would not affect the conclusion. The magnitude of any dynamic effect on stress caused by an earthquake can be estimated by dividing the radiated seismic energy by the

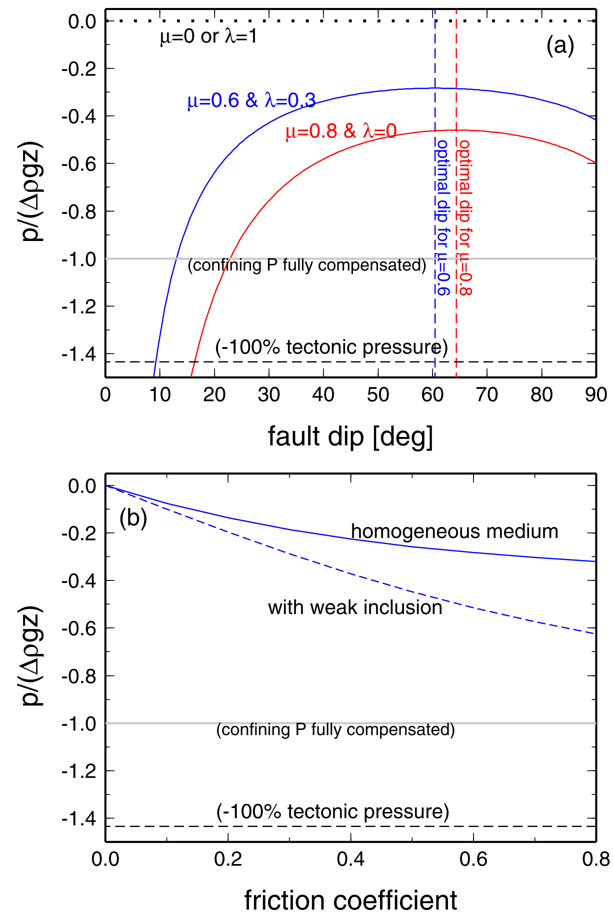

Fig. 1. (a) Dynamic pressure generated by normal faulting as a function of fault dip β, according to equation (5). Pressure is normalized by confining pressure $\Delta\rho g z$ ($= (\rho - \rho_w) g z$), where ρ_w is water density. Red curve denotes the case of $\mu = 0.8$ and $\lambda = 0$, and blue curve the case of $\mu = 0.6$ and $\lambda = 0.3$. In both cases, the Poisson's ratio ν is set to 0.25. Dotted line denotes the case of $\mu = 0$ or $\lambda = 0$. The location of optimal fault dip, corresponding to the minimum deviatoric stress (i.e., $\tan 2\beta = -1/\mu$), is shown by vertical line. The level of dynamic pressure required to fully compensate confining pressure is shown by horizontal gray line, and −100% tectonic pressure, which was somehow achieved in the numerical model of Faccenda et al. (2009), is by horizontal dashed line. (b) Dynamic pressure by normal faulting as a function of friction coefficient, with $\lambda = 0.3$ (hydrostatic), $\nu = 0.25$, and the optimal fault dip. Solid line denotes the case of a homogeneous medium (equation (5)), while dashed line denotes the maximum effect caused by viscosity heterogeneities along the optimally dipped fault (equation (6)). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

volume involved. This is the so-called the stress drop, which is on the order of only 1–10 MPa (Shearer, 2009).

The numerical modeling of Faccenda et al. (2009) is based on the visco-elasto-plastic code of Gerya and Yuen (2007), in which brittle deformation is taken into account. It is therefore puzzling that their models exhibit strongly negative dynamic pressure, enough to compensate for lithostatic pressure down to the depth of a few tens of km, along with the formation of normal faults with dip of ~60°. The 'tectonic' pressure in Faccenda et al. (2009) is defined as deviation from lithostatic pressure, and −100% tectonic pressure seen in their models is equivalent to dynamic pressure entirely canceling lithostatic pressure, which is greater than confining pressure by ~40%. The friction coefficient used is in the range of 0.4–0.6, which conforms to the fault dip seen in the models, but the amplitude of dynamic pressure seems too large.

One way to explain the strongly negative pressure of Faccenda et al. (2009) is to assume that a fault zone is inherently weaker than the surrounding rocks; viscosity heterogeneities could dis-

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