Patterns of deep seismicity reflect buoyancy stresses due to phase transitions

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Abstract. Thermal perturbation of mantle phase relations in subduction zones gives rise to significant buoyancy anomalies. Finite element modeling of stresses arising from these anomalies reveals transition from principal tension to compression near $\sim$400 km depth, down-dip compression over $\sim$400-690 km (peaking at $\sim$550 km), and transition to rapidly fading tension below $\sim$690 km. Such features, even when complicated by olivine metastability, are consistent with observed patterns of deep seismicity. That such a simple model, neglecting all effects other than buoyancy anomalies due to temperature and to thermal perturbation of olivine phase relations, successfully generates so many observed features of deep seismicity suggests that these buoyancy anomalies are significant contributors to the stress field in subducting slabs. It also suggests that the depth distribution of deep seismicity may largely reflect the state of stress in the slab rather than simply a particular mechanism of stress release.

Introduction

Deep earthquakes [*Green and Houston*, 1995] are characterized by several distinctive features: down-dip compressional focal mechanisms [*Vassiliou and Hager*, 1988], normally distributed depth of occurrence about $\sim$550–600 km [*Frolich*, 1989], and abrupt cessation of seismicity at $\sim$690 km depth [*Rees and Okal*, 1987]. These patterns must reflect some combination of the stresses present in slabs of subducting lithosphere and the failure mechanisms which release these stresses. Buoyancy forces constitute one important contributor to slab stresses: both those due to the negative thermal buoyancy of the cold slab (responsible for its sinking) and those due to buoyancy anomalies arising from thermal perturbation of mantle phase relations within the cold slab [*Turcotte and Schubert*, 1971, 1972; *Bina and Liu*, 1995]. Extending earlier work [*Ito and Sato*, 1992; *Bina*, 1996], I use a finite element method to solve the full elastic plane strain problem for these buoyancy forces. I demonstrate that the mechanisms, depth distribution, and cessation of deep seismicity all correspond to features of the stress field contribution arising solely from the buoyancy forces in the slab.

I construct a very simple 2-D model of a subduction zone. Temperatures are calculated (on a 120×90 grid) from the finite difference algorithm of N. H. Sleep [*Toksöz et al.*, 1973] for 140 Ma lithosphere subducting at 8 cm/yr at a dip of $60°$ using an initial GDH1 [*Stein and Stein*, 1992] lithospheric

temperature structure. Initial pressures are obtained by radial integration of a reference density model [*Dziewonski et al.*, 1975]. Composition is assumed to be that of mantle olivine [*Ringwood*, 1982], a uniform $(Mg_{0.9}Fe_{0.1})_2SiO_4$. For this model, mineral phase assemblages are determined by a free energy minimization algorithm based upon simulated annealing [*Bina*, 1998], using thermodynamic parameters [*Fei et al.*, 1991] for olivine polymorphs (α, β, γ), magnesiowüstite (mw), and silicate perovskite (pv). The resulting phase assemblages and their associated buoyancy anomalies are shown in Figure 1 for two cases: that of stable equilibrium and that in which α olivine persists metastably below 1000 K [*Sung and Burns*, 1976; *Rubie and Ross*, 1994; *Wang et al.*, 1997]. Note that the equilibrium $\alpha \rightarrow \alpha + \beta \rightarrow \beta \rightarrow \beta + \gamma \rightarrow \gamma$ transition series in the mantle is deflected upwards in the cold slab (and replaced by the $\alpha \rightarrow \alpha + \gamma \rightarrow \beta + \gamma \rightarrow \gamma$ series in the coldest core of the slab), yielding negative buoyancy anomalies, and the $\gamma \rightarrow \gamma + pv + mw \rightarrow pv + mw$ series is deflected downwards, yielding a positive buoyancy anomaly. This pattern of deflection is consistent with the topography of seismic discontinuities observed in some subduction zones [*Collier and Helffrich*, 1997].

Recently, I obtained a crude estimate of stress from these buoyancy structures by integrating the down-dip component of the buoyancy forces while neglecting elasticity and assuming the slab to act as a perfect stress guide, and I argued that the maximum in depth distribution of seismicity at $\sim$550–600 km corresponds to the maximum in down-dip compressive stress within the slab [*Bina*, 1996]. Here I employ the finite element method (FElt software [*Gobat and Atkinson*, 1995], with top boundary fixed vertically, using 4-node isoparametric elements) to solve the full elastic plane strain problem which results from these buoyancy structures, obtaining the displacements and stresses which minimize the strain energy at static equilibrium [*Segerlind*, 1984]. The resulting principal stresses, σ_1 and σ_2, are shown in Figure 2. For the equilibrium case, the down-dip slab stresses change from tensional to compressional just below $\sim$400 km. The slab exhibits down-dip compressional stresses over the depth range ($\sim$400–690 km) of deep seismicity, in accord with observed focal mechanisms [*Vassiliou and Hager*, 1988]. Furthermore, the depth variation of compressive stress magnitude (peaking near $\sim$550 km) matches well the observed seismicity distribution [*Frolich*, 1989]. Finally, the large down-dip compressional stresses quickly yield to fading tensional stresses just above $\sim$690 km, in agreement with the observed abrupt cessation of seismicity near this depth [*Rees and Okal*, 1987]. The latter two points can be seen clearly in profiles down the slab temperature minimum, shown with the depth distribution of global seismicity [*NEIC*, 1994] in Figure 3. Note that the maximum compressive stresses fall at the high end of the broad range of stress drops determined
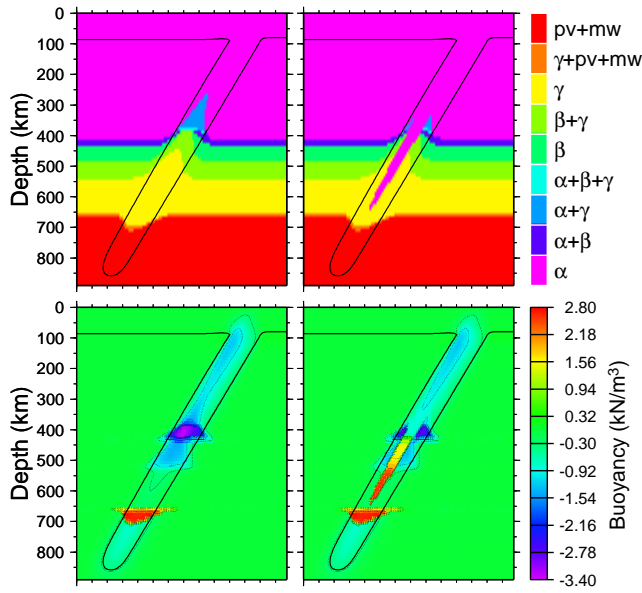
Figure 1. TOP: Slab and mantle phase assemblages for equilibrium case (left) and for metastable persistence of α olivine (right). Bold line is 1500 K isotherm. BOTTOM: Corresponding buoyancies ($g\Delta\rho$) relative to ambient mantle. Positive buoyancies float; negatives sink.

Figure 3. Magnitude of the (absolute) maximum principal stress, σ_{max}, profiled down-dip along slab temperature minimum, for equilibrium case (top) and for metastable persistence of α (bottom). Also shown are profiles 25 km normal to slab above (dashed) and below (dotted) temperature minimum. Center panel shows depth distribution of global seismicity for 1964-1994.

for deep earthquakes [*Houston and Williams*, 1991].

The case of metastably persisting olivine [*Kirby et al.*, 1996] is similar except for the intrusion of a positively buoy-
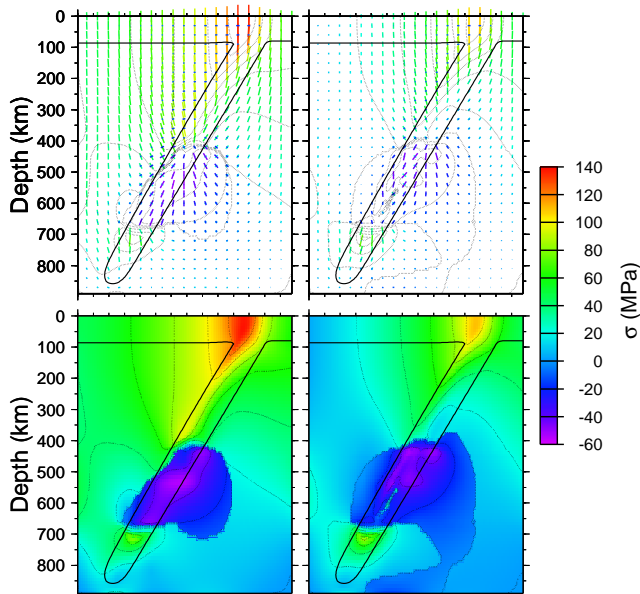
ant wedge of α into the negatively buoyant interior of the slab (Figure 1). This results in a narrow interior region, beginning just above $\sim$550 km, in which the (absolute) maximum principal stress is tensional and is not oriented down-dip. Along with rotated stresses of decreased compressional magnitude in the outer portions of the slab, this skews the compressive stress maximum toward shallower depths (Fig-

Figure 2. TOP: Calculated principal stresses for equilibrium case (left) and for metastable persistence of α (right). Segment lengths are proportional to stress magnitudes. Bold line is 1500 K isotherm. Dotted lines contour magnitude of the (absolute) maximum principal stress. Negative stresses are compressional; positives are extensional. BOTTOM: Magnitude of the (absolute) maximum principal stress, σ_{max}.
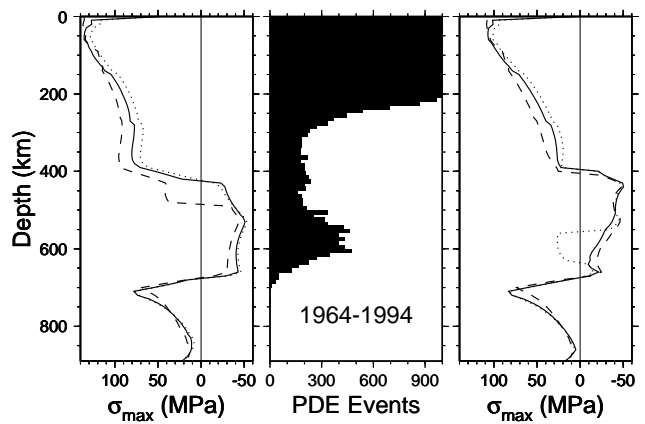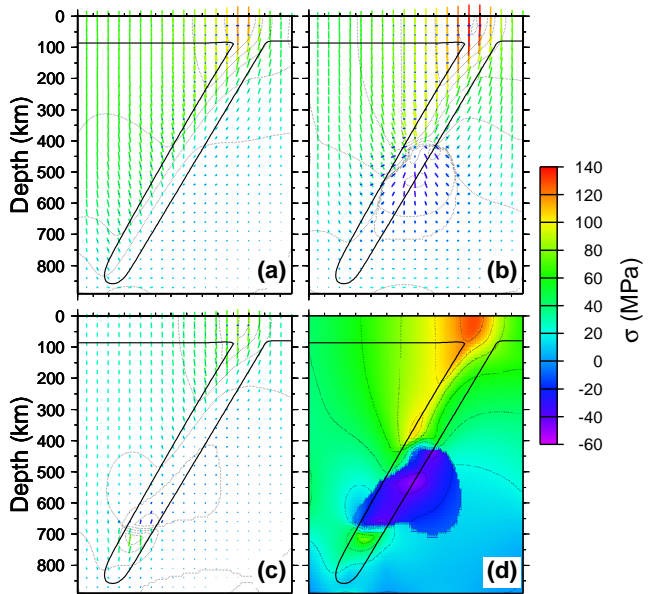

Figure 4. (A): Same as top left of Fig. 2, but for α phase only (i.e., no phase transitions). (B): Same as top left of Fig. 2, but for α, β, and γ phases only (i.e., only upper mantle phase transitions). (C): Same as top left of Fig. 2, but for γ, pv, and mw phases only (i.e., only deeper mantle phase transitions). (D): Same as bottom left of Fig. 2, but using uniform, constant elastic properties of α-$(Mg_{0.9}Fe_{0.1})_2SiO_4$ at room temperature and pressure throughout model.

ures 2–3).

It is interesting to note the relative contributions of the various buoyancy anomalies to the stress field. Thermal buoyancy alone leaves the slab largely in tension (Figure 4a). The upper mantle phase transitions involving α, β, and γ yield a compressive stress maximum at the appropriate depth, but the stresses drop off rapidly below (Figure 4b). The deeper reactions involving γ, pv, and mw yield large compressive stresses only in their immediate vicinity (Figure 4c). The combined effects of all of these anomalies are necessary to produce a stress field matching global seismic observations (Figure 3). It is also important to note the role of variable elastic properties; the stiffness of the cold slab acts to guide the stress contributions of the various buoyancy anomalies. Calculations employing uniformly constant elastic properties of olivine at room temperature and pressure throughout the model yield smaller, more localized compressive stress maxima (Figure 4d), indicating that temperature dependence of the elastic moduli [*Bina and Helffrich*, 1992] tends to concentrate these stresses throughout the cold slab interior (Figure 2).

In constructing this simple model, I have deliberately neglected many additional factors in order to isolate the stress contributions arising from buoyancy forces attendant upon thermal perturbation of phase relations. For example, a viscoelastic rheology may be more appropriate in the high-temperature mantle wedge region [*Karato*, 1997], further enhancing the slab's stress-guide efficiency beyond that seen in the contrast between Figures 2 and 4d. Dynamical effects have been ignored here, yet the torque exerted on the slab by buoyancy (and possibly other) forces may bend the slab toward the horizontal near 660 km, as observed in some seismic studies [*Lundgren and Giardini*, 1992; *Lay*, 1994; *van der Hilst*, 1995]. While olivine probably comprises at least 60% of the mantle by volume [*Jeanloz*, 1995], phase transformations involving pyroxene, garnet, and silicate ilmenite will also contribute to buoyancy anomalies, but such contributions should be distributed over broader depth ranges than those due to olivine polymorphism [*Bina and Wood*, 1984; *Akaogi et al.*, 1987; *Wood and Rubie*, 1996]. Feedback of latent heats of transformation into slab thermal structure and reaction kinetics will affect the fine structure of phase boundaries [*Daessler and Yuen*, 1993] and their associated stresses [*Yoshioka et al.*, 1997], and chemical layering within the slab will introduce further fine structure [*Helffrich et al.*, 1989; *Karato*, 1997]. Additional stresses may arise from viscous drag between slab and mantle [*Davies*, 1980], further enhancing the down-dip tension at shallow depths evident in Figure 2. At greater ($\geq$ 660 km) depths [*Vassiliou et al.*, 1984], interaction of the slab with putative contrasts in chemical composition [*Zhao and Anderson*, 1994] or viscosity [*King*, 1995] may further amplify the local compressive stresses in Figure 4c.

In addition, internal stresses may be generated in response to accumulated heterogeneous volume changes [*Goto et al.*, 1985]. Neglecting buoyancy forces, *Goto et al.* [1987] computed internal stresses arising from the volume changes accompanying transformation of an 80-km (down-dip) length of slab from α to γ (over $\sim$7 GPa with no β field) at depth in the mantle, employing slab and mantle models of a particular viscoelastic rheology underlain by a layer of partial melt, for a variety of slab thermal parameters. While they, too, found that a downward-deflected

transition at great ($\gtrsim$600 km) depth is required to generate compressive stresses consistent with observed seismicity, they were compelled to invoke kinetic hindrance of the $\alpha \to \gamma$ transition to these depths, since they were unaware of other sources of resistant forces below 600 km such as $\gamma \to pv + mw$ breakdown. However, their 80-km step size and omission of both β-phase stability and γ-phase breakdown call for caution in quantitative interpretation of their results. Moreover, such internal strains should be fully relaxed by grain-size reduction during nucleation and growth of new phases [*Karato*, 1997] and thus are highly unlikely to be retained across phase transitions. In contrast to their modeling of accumulated of internal strains over time, my analysis focuses upon body forces applied at each instant in time.

Despite my omission of such potential complicating effects, the very fact that such a simple model, incorporating only buoyancy anomalies due to temperature and to thermal perturbation of olivine phase relations, successfully generates so many observed features of deep seismicity indicates that these buoyancy anomalies contribute significantly to the state of stress in subducting slabs. Furthermore, it also suggests that observed patterns of deep seismicity largely reflect the state of stress in the slab, rather than simply reflecting a particular failure mechanism of seismogenesis. If this is the case, then certain consequences of the stress field, such as the depth of seismicity cutoff, should vary with the dip angle, age, and rate of subduction [*Helffrich and Brodholt*, 1991]. The extraordinarily rapid subduction in the Tonga-Kermadec area, for example, should produce a colder slab and hence greater depression of the $\gamma \to \gamma + pv + mw \to pv + mw$ transition and its associated buoyancy anomaly. Indeed, consistent with this hypothesis, seismicity beneath Tonga-Kermadec appears to extend 25–30 km deeper than beneath other subduction zones [*Okal and Bina*, 1997].

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