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Subduction zones: observations and geodynamic models

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Abstract

This review of subduction and geodynamic models is organized around three central questions: (1) Why is subduction asymmetric? (2) Are subducted slabs strong or weak? (3) How do subducted slabs interact with phase transformations, changes in mantle rheology, and possibly chemical boundaries in the mantle? Based on laboratory measurements of the temperature dependence of olivine, one would conclude that the core of a subducting slab is at least 10,000 times more viscous than ambient mantle; however, there are a number of complementary but independent observations that suggest that slabs are much weaker than this. Slabs undergo significant deformation in the upper mantle and may thicken to twice their original width by the time they reach the base of the transition zone. The lack of a clear correlation between the observed dip angle of deep slabs and plate velocity, rate of trench migration, and slab age in modern subduction zones is consistent with hypothesis that subduction is a time-dependent phenomenon. Both tank and numerical convection experiments with plates conclude that subduction is not a steady phenomenon, but that slabs bend, thicken, stretch, and change dip through time. This is at odds with the assumptions used in steady-state slab thermal models, where slab deformation is not considered.
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1. Introduction

The formation, evolution, and eventual subduction of oceanic lithosphere dominates the large-scale dynamics of the mantle–lithosphere system and the heat budget of the mantle (e.g. Davies and Richards, 1992). Plates organize flow within the mantle, with descending flow at subduction zones and ascending flow in the intervening regions. Subducted slabs are the major source of buoyancy (in this case negative buoyancy) that drives mantle flow (Hager, 1984). Thus, it becomes clear that understanding the process of subduction is important for furthering our understanding of the thermal and chemical evolution of the Earth. Plate motions also provide the most direct observations of the scale of mantle motion as evidenced by the agreement between a kinematic flow model using plate reconstructions to advect subducting slabs for the last 200 Ma and long-wavelength global seismic tomography (Lithgow-Bertelloni and Richards, 1995).

It is important to properly incorporate subduction in geodynamic models because: (1) plate motions provide the most direct observation of the scale of mantle convection (e.g. Hager and O'Connell, 1978, 1981; Richards and Davies, 1992); (2) the negative buoyancy of subducting slabs provides the dominant driving force for plate motions (e.g. Forsyth and Uyeda, 1975; Hager, 1984); (3) seismic tomographic images of subduction zones provide the most direct observation of mass flux between the upper and lower mantle

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(e.g. Grand et al., 1997; van der Hilst et al., 1997); (4) plates and slabs impose a long-wavelength platform to the pattern of convection in the mantle that would not otherwise develop in uniform-viscosity or stagnant-lid mode of convection at the Rayleigh numbers appropriate for the mantle (e.g. Gurnis, 1989; Lowman and Jarvis, 1996); (5) the long-wavelength platform of mantle convection has a significant impact on the heat flow and thermal evolution of the mantle (Gurnis, 1989; Lowman and Jarvis, 1996). The key question becomes what is the correct way to incorporate subduction in convection models? The answer is that this depends on the nature of the problem being addressed. For some global investigations, it may be sufficient to impose plate velocities as surface boundary conditions because the most important affect of plates and slabs at the longest wavelength is organization, or platform that plates and slabs impose on mantle flow. In order to understand problems at the scale of crustal recycling and/or volatile evolution, it is probably necessary to model the deformation associated with the bending and unbending of the slab in the shallow subduction zone.

2. Why is subduction asymmetric?

This is one of the most basic features of plate tectonics and has presented a challenge for geodynamic models because the unique, one-sided motion of the two plates that come together at a subduction zone is not a feature of classical Rayleigh–Benard convection. Several solutions to the asymmetric subduction problem have been proposed and these will be examined in detail below. It is important first to clarify the relationship between features in convection models and observable features on Earth.

The surface of the Earth and the core-mantle boundary are both isothermal, zero-stress boundaries for the mantle. The Rayleigh number is a dimensionless parameter that can be thought of as a ratio of the magnitude of the buoyancy force, resulting from temperature variations within the fluid, to viscous stresses. For classical Rayleigh–Benard convection (convection in an incompressible fluid with isothermal top and bottom boundaries, uniform viscosity, and constant thermodynamic properties where the density assumed constant in every term of the equation except for the buoyancy force term) the Rayleigh number and the geometry (i.e. a Cartesian box of specified length or spherical shell with a specified ratio of inner and outer radii, for example) uniquely describe the problem. While Rayleigh–Benard convection is a useful starting point for understanding the physics of fluid flow, there are a number of assumptions in Rayleigh–Benard convection that can be removed in order to better approximate the Earth (e.g. the mantle is slightly compressible, the mantle is not isoviscous and the thermodynamic properties of the mantle vary with temperature and pressure). The Rayleigh number remains a useful measure of the vigor of convection even though it is no longer sufficient to uniquely describe the conditions of more complex fluid problems (see Table 1). For example, in a vigorously convecting fluid the temperature variations in the interior of the fluid are small and most of the temperature difference between the surface and bottom of the fluid is confined to thin layers near the top and bottom of the fluid (and thin vertical columns that separate the nearly isothermal interiors). These layers are called thermal boundary

Table 1
Values for mantle Rayleigh number

Symbol	Quantity	Value	Reference
ρ_0	Density	3.7×10^3 kg/m^3	Dziewonski and Anderson (1981)
g_0	Magnitude of gravity	9.8 m/s^2	
h	Depth of the core–mantle boundary	2.891×10^6 m	
ΔT	Superadiabatic temperature increase	2×10^3 K	
α_0	Thermal expansivity	(5–20) $\times 10^{-6}$ K^{-1}	Chopelas and Boehler (1989)
κ_0	Thermal diffusivity	(5.4–24.0) $\times 10^{-7}$ m^2/s	Brown (1986)[a]
η_0	Viscosity	(3–10) $\times 10^{21}$ Pa s	Mitrovica (1996)
$Ra = \rho_0 g_0 \alpha_0 \Delta T h^3/\kappa_0 \eta_0$	Rayleigh number	(0.365–21.6) $\times 10^6$	

[a] For thermal conductivity, $c_p = 1.25$ kJ/kg K, density.

layers and their thickness scales with the $-1/4$ to $-1/3$ power of the Rayleigh number (McKenzie et al., 1974). The relationship between thermal boundary layer thickness and Ralyleigh number has the same form (although different coefficients) for Rayleigh–Benard and compressible convection (Jarvis and Peltier, 1989).

The lithosphere of the Earth is the cold, top thermal boundary layer of the mantle. The conceptual picture of a plate as a passive raft that sits on top of the mantle neglects the fact that oceanic plates are created from and return to the mantle. It is generally accepted that the cold, negatively-buoyant slabs provide the dominant force that drives plate motions, historically referred to as 'slab-pull' after Forsyth and Uyeda (1975). The terms plate, lithosphere, and thermal boundary layer are often used synonymously when describing convection calculations with dynamic plates and/or subduction. The lithosphere is comprised of chemically and rheologically distinct crustal and residual mantle components, or layers. The average thickness of the oceanic crust, 6 km, is four to five times smaller than the distance between adjacent grid points in most numerical calculations. Compositional and rheological components of the lithosphere have been included in some convection calculations (Gaherty and Hager, 1994; Lenardic and Kaula, 1996; Van Keken et al., 1996; Christensen, 1997). The general consensus is that buoyant crust is not able to separate from the denser lithosphere (e.g. Gaherty and Hager, 1994) unless there is a low viscosity region separating the crust and lithosphere (Van Keken et al., 1996). Thus, most mantle convection calculations ignore the compositional and rheological variations that would be introduced by including a chemically distinct crust.

Mantle convection studies began with the examination of the Rayleigh–Benard problem described above. Symmetric downwellings (e.g. the thermal boundary layers on both sides of the downwelling participate in the formation and development of the downwelling limb) are always observed Rayleigh–Benard problems (e.g. Gaherty and Hager, 1994). Downwellings associated with subduction do not behave in this manner; instead one plate descends into the mantle beneath the other. The importance of this difference cannot be over-emphasized. When both downwelling limbs participate in the formation of the downwelling, usually the coldest, near-surface part of the boundary layer remains at the surface and the lower part of the boundary layers participate in the formation of the downwelling. In our current view of subduction, most, if not all, of the subducting plate becomes the subducting slab. In this case, although there is only one boundary layer forming the downwelling, the temperatures within that downwelling may be colder than the case of symmetric downwellings which have more material participating in the downwelling in total but less of the coldest parts of the downwelling. There has been very little systematic study of the details of slab thermal structure, so it is not possible to state with certainty which case will produce the greatest density anomaly. Different formulations of asymmetric subduction may produce different results.

Corner flow models of subduction have shown that the angle of subduction could be controlled by imposing asymmetric velocity boundary conditions (Stevenson and Turner, 1977; Tovish et al., 1978). These studies show that for a given slab buoyancy, there is a critical angle at which a dipping slab will be in steady equilibrium. Angles shallower than the critical angle are unstable, while angles steeper than the critical angle will evolve toward the critical angle. Tank experiments with migrating trenches (Griffiths et al., 1995; Guillou-Frottier et al., 1995) and numerical experiments (Scott and King, in preparation) confirm that downwellings with angles shallower than a critical value undergo significant deformation and often break up.

Davies (1977) suggested that plate motions are an important component in the development of slab geometries. Using observed plate velocities to drive a kinematic flow in a 3D spherical model Hager and O'Connell (1978) predicted the geometry of subduction zones surprisingly well. The flow model in that study did not include buoyancy forces which would have increased the vertical component of slab motion relative to the kinematic flow because the cold, dense slabs will sink in a viscous fluid. This would have reduced the fit of the predicted slab dips to the observed slab geometries. This leaves a rather odd paradox because, it is clear that slabs are more dense that their surrounding (e.g. Hager, 1984) and provide a major driving force for plate motions (Forsyth and Uyeda, 1975) yet some observations (like slab geometry) appear to be better fit by models that ignore slab

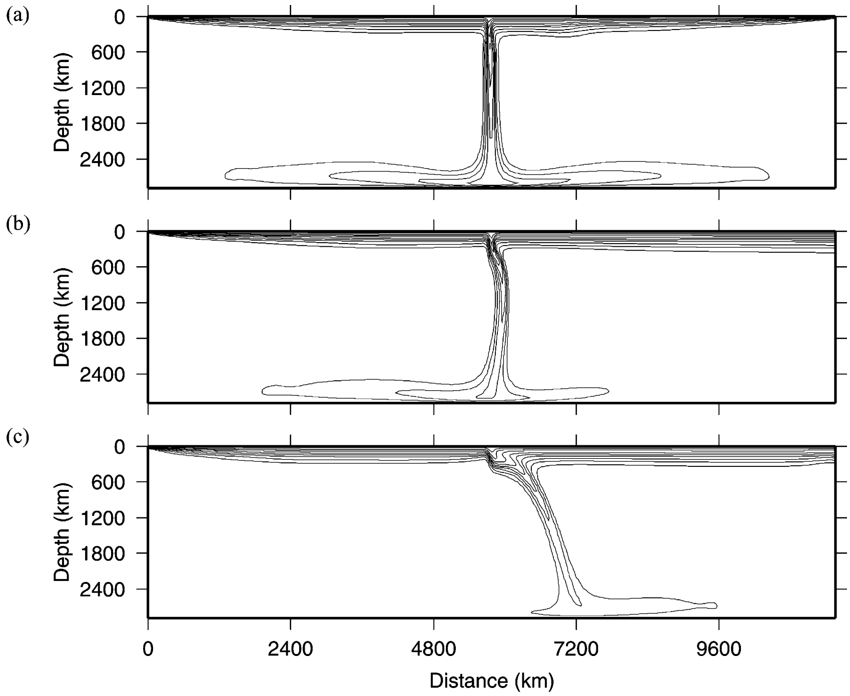

Fig. 1. Thermal fields from two convection calculations with plates. In both calculations the Rayleigh number is 10^6, the top and bottom are isothermal free-slip boundaries, the fluid is internally heated, and the viscosity is temperature dependent. The initial temperature field has a square root of age plate thermal field on the left side of the box and a uniform boundary layer on the right side of the box. Weak material zones allow the plate to deform at the trench and the ridge. The 'ridge' is in the left corner of the box and the 'trench' is in the center of the box. Further details on this type of calculation can be found in Chen and King (1998). (a) The side velocity boundary conditions are free-slip (material is free to move vertically along the edge). (b) Identical to (a) except that the velocity of a single node on the right-hand side is set to zero, breaking the symmetry of the top boundary layer velocities. (c) Identical to (b) except that the side velocity boundary conditions are an imposed horizontal velocity of 20 mm per year. This is identical to moving the trench and plate relative to the center of mass of the mantle.

density. As we will see below, this remains a problem for convection models.

Without imposing some form of asymmetry in the velocity boundary conditions, it is difficult, if not impossible, to generate an asymmetric downwelling. In Fig. 1a, the temperature field from a convection calculation that attempts to reproduce subduction is shown. The top surface of the model has a free-slip boundary condition, and there is a weak zone at the center of the box that keeps the downwelling form migrating. Even though the right-hand side of the box has a uniform thermal boundary layer and the left-hand side of the box has a square-root of age temperature profile characteristic of uniform plates, both sides of the upper boundary layer participate in the downwelling. The development of a square-root of age temperature profile on the right-hand side of the box between the non-dimensional distances of 3.5 and 4.0 illustrates that this side of the thermal boundary layer is moving like a uniform plate and is moving toward the central downwelling. When asymmetric velocity boundary conditions are applied, the resulting downwelling starts out with a dip that is approximately the critical-angle dip predicted by half-space theory (Fig. 1b); however, the downwelling quickly steepens to a near-90° dip within a distance of about 1–2 times the thickness of the boundary layer. In this case, the right-hand side of the top boundary layer is stationary, as shown by the parallel isotherms because, a zero velocity boundary condition is imposed on one node, breaking the symmetry between the left- and right-hand sides. A limitation of 2D geometries is illustrated in Fig. 1b. Because the flow is confined to the 2D plane, the upwelling flow driven by the separation of the plates at the ridge draws fluid from the right side of the box. This flow pulls the slab back under the moving plate. In a 3D geometry, the upwelling flow will draw from the 3D region and the stress on the slab induced by the upwelling flow will not be as strong as it is in the 2D geometry. Models with the most realistic slab geometries generally have both an asymmetric surface condition and trench migration, as shown in the model in Fig. 1c.

Attempts to circumvent the problem of return flow in a 2D geometry have included the use of periodic boundary conditions (e.g. Gurnis and Hager, 1988; Lowman and Jarvis, 1996; Chen and King, 1998; Han and Gurnis, 1999) or a cylindrical geometry (Puster et al., 1995; Zhong and Gurnis, 1995; Ita and King, 1998). Yet, even 3D geometries are not sufficient to explain the deep structure of subducted slabs (e.g. Bercovici et al., 1989; Tackley et al., 1993, 1994; Ratcliff et al., 1995, 1996; Bunge et al., 1996, 1997). In 3D constant viscosity calculations (e.g. Tackley et al., 1993, 1994) the global platform of the flow could be described as linear, sheet-like downwellings and cylindrical, plume-like upwellings. Although these features do not resemble plumes or slabs in any detail, the results are encouraging. When temperature-dependent rheology is included in 3D spherical calculations, the platform changes, i.e. cylindrical downwellings at the pole and sheet-like upwellings (Ratcliff et al., 1995, 1996). Even though calculations have moved from 2D Cartesian geometries to more a appropriate geometry for the mantle (3D spherical models), the results suggest that there is still the need for an additional symmetry-breaking condition to produce asymmetric downwellings.

A series of investigations of convection with temperature-dependent and/or non-Newtonian rheologies documented a transition from the free-slip platform found in constant viscosity calculations through a sluggish-lid platform to a stagnant-lid platform of convection as the stiffness of the boundary layer increases (Van den Berg et al., 1991; Solomatov, 1993; Christensen, 1984; Solomatov, 1995; Moresi and Solomatov, 1995; Solomatov and Moresi, 1996; Tackley, 1996, 1998). The sluggish- and stagnant-lid convective planforms do not include asymmetric downwellings or piecewise-uniform surface velocities, both of which are characteristics of subduction zones that we would like to be able to model (Weinstein and Christensen, 1991). While the structure of the top thermal boundary layer might seem somewhat removed from the issue of subduction, it is really quite important. In stagnant- and sluggish-lid convection, most of the top boundary layer remains at the surface and only the weakest part of the boundary layer participates in the active flow. Thus, most of the cold boundary layer is never recycled into the interior of the fluid. This is quite different from our picture of subduction, where most, if not all, of the subducting plate descends into the interior of the fluid. This has important implications for the plate driving force due to slabs and for gravity anomalies over subduction zones because the temperature structure is related to the density structure through the coefficient of thermal expansion.

In the stagnant lid platform, a significant amount of the negative buoyancy in the top thermal boundary layer does not participate in the downwelling (Moresi and Solomatov, 1995; Conrad and Hager, 1999). Furthermore, as a result of the one-sided nature of subduction, the surface heat flow from stagnant-lid convection calculations is significantly smaller than the heat flow from models with temperature-dependent rheology when a plate formulation is included (Gurnis, 1989).

The addition of a strain-weakening (non-Newtonian) component to the rheology weakens parts of the cold thermal boundary layer, especially at regions of high stress (such as in the corners of the computational domain). When coupled with temperature-dependent rheology, this can produce nearly uniform surface velocities (Weinstein and Olson, 1992; King et al., 1992; Larsen et al., 1993); however, power-law rheology calculations are typically unstable and the plate-like behavior quickly breaks down as these calculations evolve away from the carefully chosen initial conditions. Some researchers have argued that subduction forms at pre-existing zones of weakness in the lithosphere (e.g. Toth and Gurnis, 1998). Rheologies based solely on the properties of mantle minerals have not successfully produced piecewise-uniform surface velocities and slab-like downwellings. As a result, in order to generate slab-like asymmetric downwellings, convection calculations have used imposed velocity boundary conditions (e.g. Stevenson and Turner, 1977; Davies, 1986, 1988, 1989; Christensen, 1996), a priori specified mechanically weak zones (e.g. Gurnis and Hager, 1988; King and Hager, 1994), and/or dipping viscous faults (e.g. Zhong and Gurnis, 1992, 1994a, 1995, 1996, 1997). It is beyond the scope of this review to address the strengths and limitations of each of these approaches in detail. The reader is referred to King et al. (1992), Zhong et al. (1998), Han and Gurnis (1999), Bercovici et al. (1999) and references therein for further discussion of plate generation methods.

A criticism of the calculations discussed above is that they are limited, for the most part, to two-dimensional geometries, or three-dimensional geometries with reflecting side-wall boundary conditions. Symmetric, near-90°-dipping downwellings are also the observed platform in three-dimensional spherical convection models (e.g. Tackley et al., 1994; Bunge et al., 1997). Even when temperature-dependent rheology is included in spherical calculations, dipping slab-like features are not observed (Ratcliff et al., 1995, 1996). The use of periodic boundary conditions in two-dimensions eliminates the effect of the side walls and two-dimensional calculations can be formulated in such a way that they are formally equivalent to calculations that explicitly allow the trench to move relative to the grid (cf. Han and Gurnis, 1999).

Trench migration has been one of the most studied mechanisms for producing dipping slabs (Davies, 1986; Gurnis and Hager, 1988; Zhong and Gurnis, 1995; Griffiths et al., 1995; Christensen, 1996; Ita and King, 1998). It is interesting to note that while numerical models models and tank experiments have demonstrated that trench migration is an effective mechanism for generating shallow dipping slabs, a compilation of data from various subduction zones (Jarrard, 1986) shows almost no correlation between the dip of a slab in the 100–400 km depth range and the rate of trench migration (Fig. 2). There are several cases where three-dimensional models have been employed to study specific regions with considerable success matching seismic structure (van der Hilst and Seno, 1993; Moresi and Gurnis, 1996). In these cases, the focus of attention was in the western Pacific, where slabs are old and steeply dipping. In addition, these cases have focused on subduction zones where a significant amount of trench migration has been documented. We can reconcile the apparent lack of correlation of the global trench migration

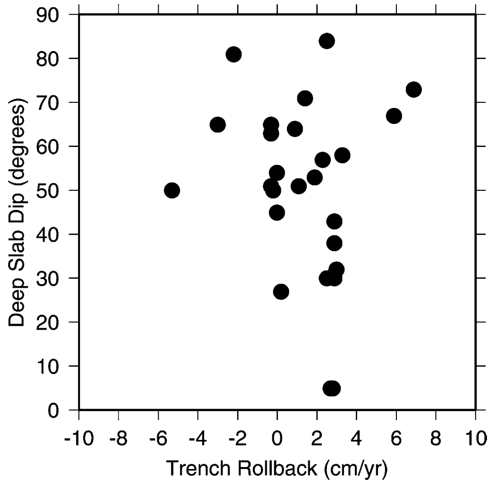

Fig. 2. Deep slab dip, as measured by the shape of the Wadati-Benioff zone in the 100-400 km depth range, vs. trench rollback using the Minster and Jordan plate motion model. Data taken from (Jarrard, 1986).

observations with slab geometry and the success of the regional subduction studies by recognizing that subduction is a time-dependent phenomenon and the shape of subducting slabs evolves with time. Both numerical and tank experiments show that buoyancy driven subduction is a time-dependent phenomenon (Gurnis and Hager, 1988; Griffiths et al., 1995; Becker et al., 1999). The geometries of present day subduction zones provide a single snap shot of a time-dependent phenomenon and each subduction zone is at a different stage in this process.

There are several observations that have not been fully exploited in convection modeling. The first of these is the difference in dip between eastward dipping and westward dipping slabs (e.g. Le Pichon, 1968; Ricard et al., 1991; Doglioni, 1993; Marotta and Mongelli, 1998; Doglioni et al., 1999) which is illustrated in Fig. 3. The difference between Chilean style subduction zones and Mariana style subduction zones has been recognized for some time (e.g. Uyeda and Kanamori, 1979). However, it is not easy to separate the effects of dip direction from other factors because most of the eastward dipping slabs are being over-ridden by the North and South American plates (i.e. continents) while many of the slabs in the Pacific are being over-ridden by island arcs or oceanic plates. Yet, a series of flexure calculations that include the effect of motion of the slab-trench system relative to the underlying mantle fit slab geometries remarkably well (Marotta and Mongelli, 1998). One of the reasons that this observation has not been explored is that many of the techniques used to generate subduction features in two dimensions are not practical to implement in a complete sphere in three dimensions calculations. Furthermore, the grid resolution that is required to study subduction problems exceeds what can be practically achieved in a three-dimensional spherical shell at present.

There is also a difference in the average dip angle between the populations of slabs where the overriding plate is continental or oceanic plate (Furlong et al., 1982; Jarrard, 1986). While the average of the two populations is distinctly different (40° versus 66° from Jarrard's data), there are notable exceptions including the shallow west-dipping Japan slab and the steep eastward-dipping Solomon and New Hebrides slabs (see Fig. 3). Studies of supercontinent breakup observe shallow dipping slabs under continents (e.g. Lowman and Jarvis, 1996). Because we are currently in a phase where continents are generally moving away from the site of the former supercontinent and over-riding oceanic plates, it is not clear whether this correlation once again reflects the effect of trench migration, or mechanical differences in the overriding plate.

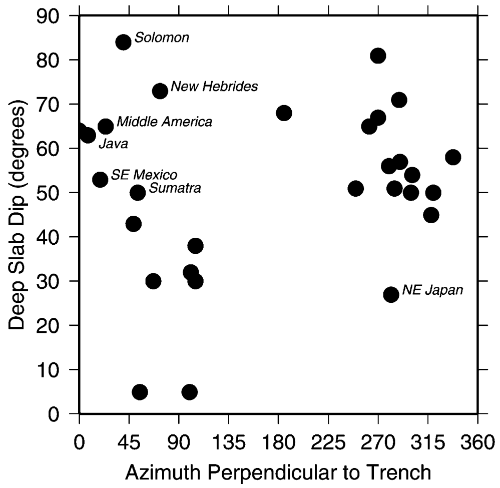

Fig. 3. Deep slab dip, as measured by the shape of the Wadati–Benioff zone in the 100–400 km depth range, vs. azimuth perpendicular to the trench. Data taken from Jarrard (1986).

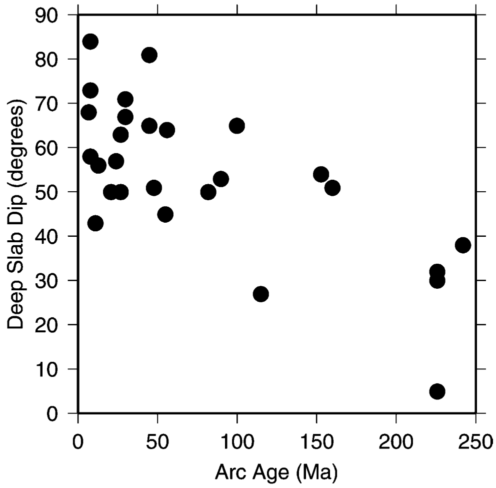

Fig. 4. Deep slab dip, as measured by the shape of the Wadati-Benioff zone in the 100–400 km depth range, vs. duration of subduction. Data taken from Jarrard (1986).

It is interesting to note that the strongest correlation in the data set compiled by Jarrard (1986) is the correlation of the duration of subduction with dip of the deep part of the slab, defined as 100–400 km (Fig. 4). This was the only significant correlation that Jarrard found in the global compilation of subduction zone data. While it is not possible to isolate this from the other factors that seem to influence slab geometry, this correlation is consistent with the assumption that subducted slabs are evolving, time-dependent features that are not at steady-state equilibrium. Furthermore, it is encouraging that the observed subduction zone geometries shown in Fig. 4 agree with the general findings of the models, that slab dip angles become more shallow with time. In both numerical and laboratory experiments, young buoyancy driven slabs steepen with age from the time of the initiation of subduction until the slab reaches the transition zone (Gurnis and Hager, 1988; Griffiths et al., 1995; Becker et al., 1999). Once the slabs reach the top of the lower mantle slab dip angles become progressively shallower if there is ocean-ward trench migration.

The area where the asymmetry of subduction is most apparent is the subduction of young lithosphere. The most striking example is the subduction of the Pacific plate underneath Alaska. Perhaps, not coincidentally, this is an area where convection calculations have had little success. It is highly probable that the subduction of young lithosphere is not driven by the negative buoyancy of the subducting slab, which is the underlying assumption in most convection subduction studies. Convection models have focused on the case of older slabs where the negative buoyancy of the slab itself provides the driving motion for the plate system (e.g. slab-pull). There are a few examples of studies of the subduction of young lithosphere (e.g. England and Wortel, 1980; Vlaar, 1983). Young slabs have less negative buoyancy than older slabs, and thus, younger slabs should have shallower dip angles than older slabs, all other things being equal.

It is important to recognize that while buoyancy within a slab is dominated by the effect of temperature, phase transformations can have significant effects (e.g. Daessler and Yuen, 1996; Bina, 1996, 1997; Christensen, 1997; Ita and King, 1998; Schmeling et al., 1999; Marton et al., 1999). The impact of phase transformations on buoyancy is most pronounced if the olivine to spinel phase transformation is kinetically hindered in cold slabs (see Kirby et al., 1996 and references therein). The presence of metastable olivine can reduce slab decent velocities by as much as 30% (Kirby et al., 1996; Schmeling et al., 1999; Marton et al., 1999) and will have the greatest effect on the oldest and coldest slabs. The stresses induced by the differential buoyancy are consistent with the pattern of deep earthquakes (Bina, 1997).

Many researchers have taken an approach which bypasses some of the problems of the development of the asymmetric downwelling by solving for the thermal structure of a subducting slab using a kinematic model where the velocity field is specified (e.g. Minear and Toksoz, 1970, Peacock, 1991, 1996, Peacock et al. 1994; Molnar et al., 1979; Helffrich et al., 1989; Kirby et al., 1996). The attraction of this approach is that the required input to the kinematic model includes plate age, plate velocity, and a specified slab dip. These quantities are easily observed (or measured), making it practical to set up a thermal structure calculation for specific subduction zones. In a convection calculation, the plate velocity, slab velocity and slab dip are not input controls, making it more difficult to set up a calculation with a geometry that resembles specific subduction zones. There are limitations to the kinematic approach. First, the velocity of the slab is not driven by the buoyancy structure of the slab. A result of the independence of slab buoyancy and plate/slab velocity that is built into the kinematic model is that old slabs and young slabs can descend with the same velocity, in apparent violation of the generally held theory that slab density is a major driving force for plate motion (e.g. Forsyth and Uyeda, 1975). In most, if not all kinematic models, the velocity of the slab is the same as the velocity of the plate (or some fraction thereof). Second, the deformation that occurs as a plate bends, breaks, and generally deforms as it passes through a subduction zone (Conrad and Hager, 1999) is not accounted for in kinematic models. It is reasonable to ask how our limited understanding of the deformation at subduction zones affects buoyancy driven slab models approximation the deformation at a trench. A third, and perhaps the most important, limitation of kinematic slab thermal models is that the slab behaves rigidly, with no internal deformation. There are a number of observations which show that slabs deform as they sink, especially in the region of the transition zone (see Lay, 1994 and references therein). Because there has been no attempt to directly compare the thermal structures from buoyancy driven slabs and kinematic slabs, it is not possible to quantify the degree to which the different assumptions impact the thermal models.

3. Are slabs strong or weak?

The strength of subducted slabs is an important mechanical property of the mantle–lithosphere system, yet flow models can reproduce some observations at subduction zone with strong or weak slabs. One example is the global distribution of seismicity with depth. Seismicity rates decay exponentially with depth in the upper mantle, reaching a minimum between the depths of 300 and 500 km. Seismicity rate then increases until 700 km at which point all seismicity abruptly ends. The pattern of seismicity with depth has been modeled using both strong (Vassiliou et al., 1984) and weak (Tao and O'Connell, 1993) slabs meeting an increase in viscosity at about 670 km depth.

Laboratory investigations of the deformation mechanism of mantle minerals demonstrate that the creep strength of mantle minerals is a strong function of temperature, pressure, strain-rate, and grain-size (e.g. Karato and Li, 1992; Karato and Wu, 1993; Karato and Rubie, 1997; Riedel and Karato, 1997). Any reasonable estimate of the temperature difference between the core of a subducting slab and average mantle leads to the conclusion that the majority of a slab must be much stronger than the mantle. However, if great earthquakes periodically cut through the entire thickness of the elastic oceanic lithosphere (Kanamori, 1971), then the mechanical properties of individual minerals may not be representative of the properties of the subducted slab as a whole. The role of water adds an additional complicating factor as laboratory work has shown that the viscosity of olivine can be reduced by as much as a factor of 100 in the presence of water (Hirth and Kohlstedt, 1996). The consideration of water raises additional problems because there are a series of dehydration reactions that take place in the slab (e.g. Peacock, 1996) and the partitioning of water between phases is not well constrained.

There are numerous seismic studies which show that slabs can bend, kink and thicken. These include the location of Wadati–Benioff zones (WBZ) as mapped through earthquake hypocenters (Isacks and Barazangi, 1977; Giardini and Woodhouse, 1984, 1986; Fischer et al., 1988, 1991), the study of earthquake focal mechanisms (Isacks and Molnar, 1968, 1971; Vassiliou, 1984; Giardini and Woodhouse, 1986; Holt, 1995), the study of moment tensors (Bevis, 1988; Fischer and Jordan, 1991; Holt, 1995) and tomography (van der Hilst et al., 1991, 1993, 1995; Fukao et al., 1992). The seismic observations have been extensively reviewed by (Lay, 1994) and here I will only highlight a few examples.

The rate of accumulation of seismic moment in WBZ can be used to estimate the average down dip strain rate in subducting slabs (Bevis, 1986; Fischer and Jordan, 1991; Holt, 1995). This assumes that the amount of aseismic deformation is small and that the viscous deformation is parallel with the brittle layer with the same deformation-rate tensor. Thus, the seismic deformation rate probably represents a minimum estimate of slab deformation. Because slabs are generally in down-dip compression or down-dip extension, the average change in the length of a slab due to the cumulative seismicity can be related to the sum of the seismic moments of the individual seismic events (Bevis, 1988). Even after accounting for the scatter in focal mechanisms, in the depth range of 75–175 km the average strain rate in subducting slabs is estimated to be $10^{-15}\,s^{-1}$ (Bevis, 1988). This compares with a characteristic asthenospheric strain rate of $3 \times 10^{-14}\,s^{-1}$ (Turcotte and Schubert, 1982; Hager and O'Connell, 1981). This leads to the surprising suggestion that the difference in strain rates between the asthenosphere and subducting slabs is no more than a factor of 3. Taking the average decent rate of slabs this strain rate corresponds to a total accumulated strain in the depth range of 75–175 km of 5%. Fischer and Jordan (1991) find that the seismic data require a thickening factor of 1.5 or more in the seismogenic core of the slab in central Tonga and complex deformation in northern and southern Tonga. Holt (1995) argues that the average seismic strain rate in Tonga represents as much as 60% of the total relative vertical motion between the surface and 670 km.

The geometry of WBZ also indicates that slabs are significantly deformed. Subducting slabs bend to accommodate over-ridding plates and the descend into the mantle; they also unbend, otherwise WBZs would curve back on themselves (Lliboutry, 1969). The dip of a subducting slab changes over a narrow depth interval, from 10–20° in the inter-plate thrust zone to 30–65° at depths near 75 km (Isacks and Barazangi, 1977). This change in shallow dip is a feature that is not well modeled by any slab model. In addition to deformation as the slab descends into the mantle, there is also deformation in the horizontal plane at subduction zones. Because oceanic plates themselves are spherical caps, the degree of misfit between subducting slab and the best fit spherical cap provides an estimate of the amount of slab deformation (Bevis, 1986). Based on this analysis, the Alaska-Aleution, Sumatra-Java-Flores, Caribbean, Scotia, Ryukyu, and Mariana arcs undergo a minimum of 10% strain (Bevis, 1986; Giardini and Woodhouse, 1986).

While convection models can reproduce many subduction zone observations with both strong and weak slabs, there are several indications from convection studies that slabs are weak. Houseman and Gubbins (1997) use a dynamic model of the lithosphere that assumes that the properties of the slab are uniform through out the entirety of the slab; the slab is both more viscous and more dense than the surrounding mantle. They find that the shape of the deformed slab is a strong function of the viscosity of the slab. They also observe a buckling mode that produces slab geometries similar to the Tonga slab. The effective viscosity of the slab needed to produce this slab geometry is 2.5×10^{22} Pa s, no more than 200 times the upper mantle viscosity and maybe only a factor of two greater than the lower mantle viscosity (cf. King and Ita, 1995).

The regional gravitational potential, or geoid, high over subduction zones has provided at important constraint on mantle rheology (Hager, 1984). In the most simple terms, the geoid at subduction zones is a balance between two large terms that are opposite in sign and almost cancel. Those terms come from the positive gravitational potential due to the dense, slab and the negative gravitational potential due to the deformation of the surface (i.e. dynamic topography). If the mantle were static, and the slab did not sink, then this term would be zero. Hager (1984) showed that in an isoviscous mantle, the dynamic topography generated by a reasonable slab density anomaly creates a dynamic topography contribution to the geoid that exceeds the geoid anomaly due to the slab, resulting in geoid lows over the subduction zone, opposite of the observation. The dynamic topography contribution can be reduced, resulting a in geoid high, with an increase in viscosity at approximately 670 km. There have been a number of efforts to improve the uniform viscosity flow model used by Hager by including: both depth dependent and temperature dependent viscosity; the effect of phase transformations; and plate formulations with temperature-dependent rheology (Zhong and Gurnis, 1992; King and Hager, 1994; Moresi and Gurnis, 1996; Zhong and Davies, 1999). The surprising result from this work is that weak slabs provide a much better fit to the geoid than strong (high viscosity) slabs.

It would be easy to dismiss any one of these observations taken by itself; however, the number and diversity of the observations that slabs are not significantly stronger than ambient mantle suggests that this should receive serious consideration. A weak slab is not necessarily inconsistent with the laboratory observations. While the viscosity of olivine is a strong function of temperature, it is also a strong function of grain size (Riedel and Karato, 1997). Grain size reduction resulting from recrystalization of minerals at phase boundaries may counter balance the effects of temperature on viscosity. In addition, the brittle component of the slab may be mechanically weakened by the faulting and thus the single crystal measurements of viscosity may not correctly describe the deformation of the slab.

It is important to recognize that assumptions regarding slab viscosity are built into some slab thermal models. For example, in the Minear and Toksoz (1970) and Peacock et al. (1994) models, the kinematic velocity field used to advect the temperatures is uniform except in the corner where the plate becomes the slab. Because the slab has a uniform velocity, there is zero strain, and hence zero deformation. The evidence reviewed above suggests that slabs thicken with depth. Because the interior of a slab warms primarily by diffusion, this suggests that the interior of thickening slabs will be colder than the kinematic models predict, all other things being equal. Because there have been no direct comparison between kinematic model, which do not include slab deformation and dynamic models, which can include slab deformation, it is impossible to make a quantitative statement.

4. How do slabs interact with phase, rheology, and (possibly) chemical boundaries in the mantle?

The eventual fate of deep slabs has been a topic of debate for several decades (see Lay, 1994 and Christensen, 1995 for reviews). One of the original arguments for a barrier to convection at 670 km depth was the cessation of earthquakes at this depth. While this is approximately the depth that corresponds to the pressure of the olivine-spinel to perovskite plus ferro-periclase phase transformation, both a chemical boundary and rheological boundary have been proposed to explain the cessation of earthquakes.

Early attempts to model subduction zones used a non-Newtonian fluid with an endothermic phase transformation. This approach produced stiff downwelling limbs that descended at a 90° dip angle as a direct consequence of the free-slip (reflecting) boundary conditions on the sides of the domain (Christensen and Yuen, 1984). The symmetry of the downwelling (i.e. material from both sides of the thermal boundary layer at the downwelling) is the major shortcoming of this approach. Christensen and Yuen identified three planforms of deep slabs at a phase boundary: slab penetration, slab stagnation, and partial slab penetration. Fluid experiments with corn syrup which used a setup designed to produce an asymmetric downwelling confirm these basic planforms (Kincaid and Olson, 1987). The tank experiments induced an asymmetric

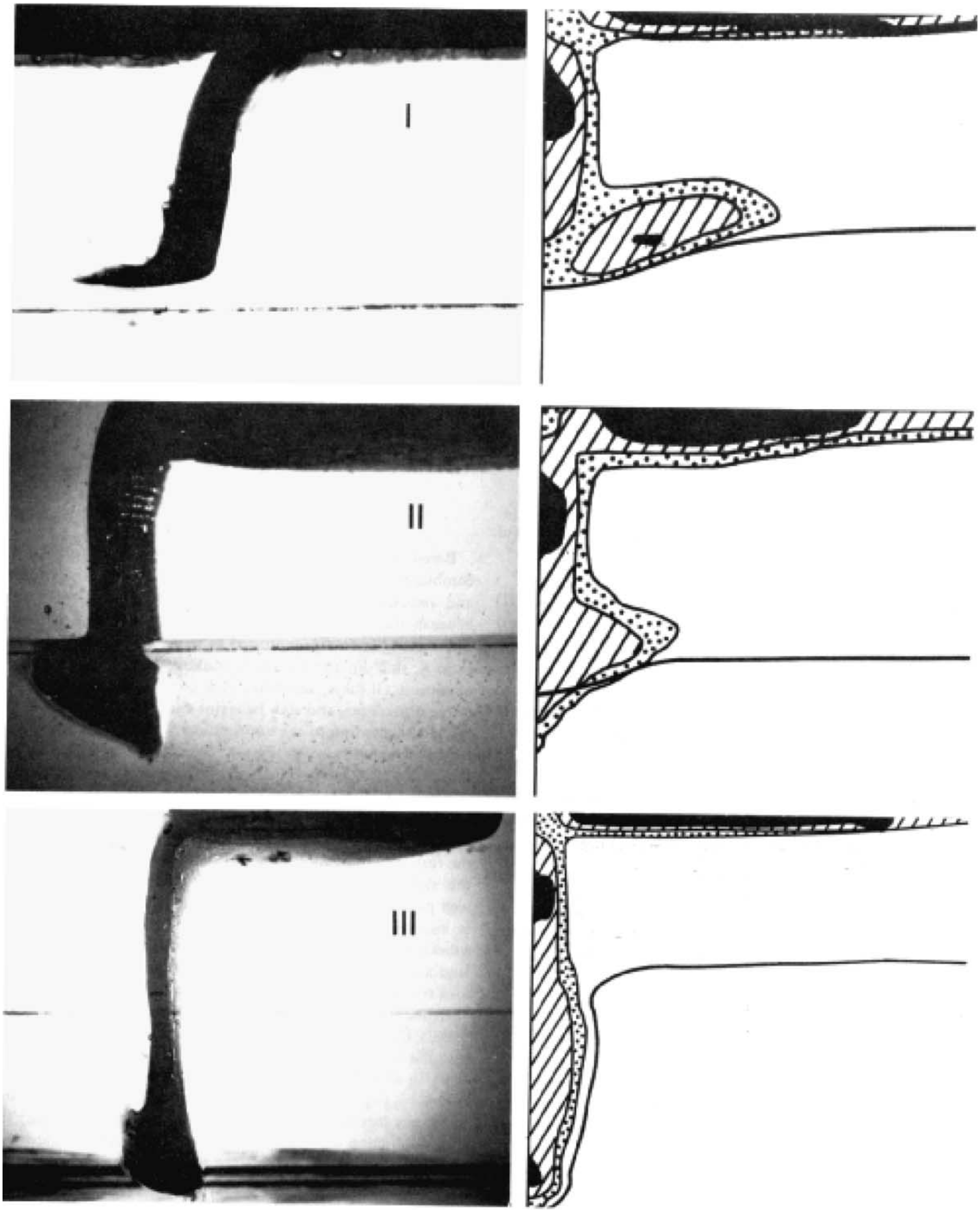

Fig. 5. Experimentally observed styles of slab penetration through a density discontinuity (left) compared with two-dimensional calculation by Christensen and Yuen (1984) (right). Part I: Slab deflection with $R \approx -0.2$. Part II: Partial slab penetration with $R \approx 0.0$. Part III: Complete slab penetration with $R \approx 0.5$. From Kincaid and Olson (1987) with permission.

downwelling by placing two cold sheets of concentrated sucrose solution on the top surface of the fluid in the tank. The larger of the two sheets was introduced with a shallow dipping bend at its leading edge; this provided the instability that allowed this plate to subduct under the other plate. The agreement between the tank and numerical experiments (Fig. 5) suggests that the three planforms discussed above are robust features of cold downwellings interacting with phase transformations and/or chemical boundaries. Subsequent numerical (Zhong and Gurnis, 1994b; King and Ita, 1995; Ita and King, 1998) and tank (Griffiths et al., 1995; Guillou-Frottier et al., 1995) experiments have confirmed the basic results and added additional insight into the interaction between subducting slabs and phase transformations.

Many investigators have studied subduction with one or more phase changes. There is a general consensus that has emerged from both 2D and 3D calculations in both Cartesian and spherical geometries. The evolution of the flow can be described by three distinct stages: (1) the initial impingment of the leading edge of the subducting slab on the spinel to perovskite plus ferro-periclase phase boundary; (2) a period of increased trench roll-back and draping of the slab on the phase boundary and; (3) virtual cessation of trench roll-back as the slab penetrates into the lower mantle. The interaction of subduction, trench-migration, and phase changes is now reasonably well understood. Several numerical studies have clearly demonstrated that the effect of a phase transformation on the pattern of flow in a convecting fluid is sensitive to the Rayleigh number, initial conditions, boundary conditions, and the equation of state approximation (Ita and King, 1994; King et al., 1997).

5. Conclusion

It is well established from both laboratory and numerical experiments, that trench migration can produce shallow dipping downwelling features. Detailed plate reconstructions have been successfully used to explain slab geometries in the western Pacific, lending support to the hypothesis that trench migration is a controlling factor in slab geometry. There may be other factors, including the properties of the overriding plate and the direction of subduction (i.e. eastward versus westward dipping slabs). The fact that slab dip angles and trench migration are not strongly correlated in the global compilation of subduction zones observations at first appears to contradict the findings of subduction zone modeling; however it is important to remember that both the numerical and laboratory experiments and the global compilation of subduction zone parameters indicate that subduction is a time-dependent phenomenon.

A more surprising result, consistent with a variety of observations, is that slabs appear to be much weaker than the predictions of slab viscosity based on temperature dependent viscosity alone. At the base of the transition zone, the viscosity contrast between the slab and lower mantle may be as small as a factor of three. This can be reconciled with laboratory creep data from single crystals if the effective viscosity of a subducting slab is weakened by faulting or as a result of the reduction of mineral grain sizes due to dynamic recrystalization as a result of phase transformations or faulting.

Both of these results require reconsideration of various slab thermal models because, the most popular and widely used slab thermal models do not account for slab deformation or time-dependent behavior. If a slab thickens by as much as a factor of two, then thermal models that assume that slabs cool by conduction are underestimating the cooling time for subducting slabs to reach thermal equilibrium with the mantle.

The subduction of young lithosphere has received little attention (e.g. England and Wortel, 1980; Vlaar, 1983). Convection models assume that subduction is driven by the negative buoyancy of the slab and it is not clear whether young, warm slabs subduct due to their own negative buoyancy or are being forced under other plates by a global force/mass balance. The Cocos plate subducting under Central America and the Juan de Fuca plate subducting under North America are both examples where young lithosphere is being subducted. In these cases, subducting may be driven by the over-riding North American continental landmass.

The Earth appears to be the only terrestrial body on which the process of subduction is currently active. This is an important but often neglected observation. A complete understanding of the process of subduction should not only explain the observations from subduction zones on Earth, it must also explain why subduction is presently not occurring on Venus or Mars. The most popular explanation, that plate tectonics requires water to weaken the crust, has only been tested at the most basic level (Lenardic and Kaula, 1994).

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