7.08 Mantle Downwellings and the Fate of Subducting Slabs: Constraints from Seismology, Geoid Topography, Geochemistry, and Petrology

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7.08.1 Introduction

Subduction zones are a major component of the dynamic-plate, convecting-mantle system, forming the downwelling limb of the mantle thermal boundary layer at Earth's surface. Subducting slabs are the dominant source of gravitational potential energy (GPE), that is, the driving force for mantle convection and plate tectonics. It is through subducted slabs that oceanic crust and sediments, including volatile compounds such as water, are recycled back into the mantle. Earth is the only terrestrial planet with active subduction zones and it is possible that it is the only planet where the process of subduction as we know it ever occurred. It is conceivable that without subduction, Earth would not have been a suitable place for life to develop. Without the recycling of volatile compounds back into the mantle at subduction zones, it is possible that Earth would have developed a runaway greenhouse similar to Venus.

There have been a number of recent reviews of subduction zones (e.g., Bebout et al., 1996; King, 2001; Stern, 2002; Eiler, 2003; van Keken and King, 2005) with some specifically focusing on seismology (Lay, 1994; van der Hilst et al., 1997), the mantle...
wedge (van Keken, 2003), phase changes and the fate of slabs (Christensen, 1995), and water in the mantle (Hirschmann, 2006). In an attempt to provide more than a summary of these reviews and recent papers, we will focus this chapter around a broad theme that emerges from recent observations and models of convection. Through a variety of observations (e.g., petrology, careful relocation of earthquake hypocenters, seismic tomography, shear-wave splitting; van der Hilst et al., 1997) it is clear that there is significant three-dimensional (3-D) structure and complexity within subduction zones. Not only do subducted slabs change dip along the length of an arc, studies increasingly recognize slab tears, slab windows, and other forms of slab deformation. While such complexity was noted in some of the earliest subduction zone studies (e.g., Isacks and Molnar, 1971), most slab thermal structure models assume no deformation of the downgoing slab. There is a continuing effort to document and understand changes in petrology along arcs as well, and in many cases these changes cannot be correlated with the traditional parameters (convergence rate and slab age) that control simple slab models. Complementary to the increasing complexity of the observations, convection modeling tools have increased numerical resolution and sophistication in terms of the complexity of rheology formulations, and physics of melting, dehydration, and flow. These models are now attempting to model specific subduction zones using constraints that were not possible to consider with earlier generations of models.

Our understanding of the geometry and composition of subducted slabs, as well as the physical process of subduction, comes from indirect observations including seismology (e.g., earthquake hypocenters, seismic structure, anisotropy, and attenuation; see Chapter 4.11), gravity, topography, geochemistry, and petrology of island arc lavas. Numerical calculations of the dynamics of subducted slabs and slab thermal models also play an important role, enabling us to test hypotheses. There have been many advances in all these areas in the past decade, and it would be impossible to do justice to all the important contributions that have been made, some of which are chapters of other parts of this volume. An attempt has been made in this chapter to point to other review papers along the way that highlight areas which cannot be given proper attention here.

The development of the worldwide seismographic network enabled the accurate location of earthquake hypocenters. Gutenberg and Richter (1939) first classified earthquakes by focal depth and compiled the distribution of global earthquake hypocenters clearly showing that intermediate- and deep-focus earthquakes occurred in narrow zones that are consistent with descending plate material beneath trenches (cf. Isacks et al., 1968). In addition, it was recognized that earthquake focal mechanisms of the deepest earthquakes are aligned with the axis of compression along the dip of the plane delineated by the hypocenters (cf. Isacks and Molinar, 1971; see Chapters 4.11 and 4.16) and this was taken to be evidence that the downgoing lithosphere meets some form of resistance with depth (e.g., Richter, 1973; Vassiliou et al., 1984). It was also recognized that the focal mechanisms of intermediate-focus earthquakes were aligned with the axis of extension along the dip of the plane delineated by the hypocenters in areas where there are no deep-focus earthquakes. The newly recognized process of seafloor spreading required the consumption of lithosphere to maintain a constant volume of crust (Morgan, 1968; Le Pichon, 1968) and Le Pichon (1968) calculated the rate of seafloor consumption based on the rate of formation of new seafloor. Le Pichon’s estimate agrees with the predicted rate of destruction of seafloor at trenches from the seismically determined slab geometry (e.g., Isacks et al., 1968).

The realization that cold plate was sinking into the mantle (e.g., Vening Meinesz, 1954) quickly led to the development of the first slab thermal models (McKenzie, 1969; Minear and Tóksoz, 1970), recognizing that the thermal structure of the descending slab would impact surface heat flow, gravity, and seismic-wave travel times in the area surrounding the trench. It was apparent from these models that the temperature of the slab was controlled by the rate of seafloor spreading and thermal properties of the slab (McKenzie, 1969). Some of these early models predicted that for the slowest moving plates (≈1 cm yr⁻¹), the slab would be in thermal equilibrium with the mantle by a depth of 600 km (Minear and Tóksoz, 1970). Turcotte and Schubert (1973) explained the spatial relationship between the newly discovered downgoing slabs and the linear island chains associated with trenches by frictional heating associated with the motion of the descending slab. Using a 1-D analytical treatment of narrow shear zones, Yuen et al. (1978) showed that the mantle deforms readily enough that shear zones are insufficient to produce melting. The relative importance of frictional heating has been debated (Turcotte and Schubert, 1973; Scholz, 1990; Molnar and England, 1990; Peacock, 1992; 1993; Tichelaar and Ruff, 1993; Hyndman and Wang, 1993), and we will return to this topic when we examine thermal models below.
7.08.1.1 Subducted Slabs and GPE Driven Plate Tectonics

We begin by focusing on the GPE associated with subducted slabs, the primary energy source driving mantle convection (e.g., Turcotte and Oxburgh, 1967; Forsyth and Uyeda, 1975; Richter, 1977; Davies and Richards, 1992). First, we estimate the thermal structure (hence buoyancy) that is coming into the trench as a result of the cooling of the oceanic plate. The thickness of an oceanic plate, \( d \), is proportional to the square root of the time the plate spends at the surface. The constant of proportionality is the square root of the thermal diffusivity, \( \kappa \). This time can be estimated by the length of the plate, \( L_p \), divided by the plate velocity \( v_p \). Hence

\[
d = \frac{\kappa L_p}{v_p} \tag{1}
\]

(cf. Chapter 7.04). As the plate subducts, the cold material is more dense than the warmer mantle surrounding it. Thus the density difference between the slab and mantle will be \( g \rho_o \Delta T \) where, \( \alpha \) is the coefficient of thermal expansion, \( \rho \) is the density, \( g \) is the gravitational acceleration, and \( \Delta T \) is the average temperature difference between the interior mantle and surface. This buoyancy must be balanced by resistance of the mantle surrounding the slab, which has the form \( \eta D \eta \), where \( \eta \) is the coefficient of dynamic viscosity and \( D \) is the deformation-rate tensor (see Chapter 7.02). A crude estimate of the deformation rate is the velocity gradient, which we will assume goes from \( v_p \) at the plate to zero at some point near the middle of the mantle. Assuming the length of the plate, \( L_p \), is approximately the same as the depth of the mantle, \( D \), and balancing the buoyancy force and viscous resistance, one obtains

\[
v = D \left[ \frac{g \rho_o \Delta T \kappa^{1/2}}{2 \eta} \right]^{2/3} \tag{2}
\]

With a reasonable choice of values for these parameters, one obtains a value for \( v_p \) on the order of 100 mm yr\(^{-1}\). This value is remarkably close to the velocity of oceanic plates with subduction zones (cf. Forsyth and Uyeda, 1975), especially when one considers the simplicity of the analysis.

For each of the major plates, Forsyth and Uyeda (1975) compiled plate velocity, area of plate, length of ridge, trench, and transform fault along the circumference of the plate boundary and calculated the relative importance of each contribution, concluding that the subducting slabs dominate the force balance. Taking their data (from table 1 of Forsyth and Uyeda, 1975), we plot the covariance matrix (Figure 1) illustrating that the percent of plate boundary dominated by trench (i.e., effective trench boundary — as trenches at opposite sides of the plate will presumably cancel their contribution) has the strongest correlation with plate velocity (correlation coefficient of 0.92). The solid colors indicate a strong correlation (green, positive; pink, negative) and the more washed out the color, the weaker the correlation. The percentage of continental area is next (correlation coefficient of 0.64) and the percent of effective ridge along the plate boundary is third (correlation coefficient of 0.46). The total area of the plate and the percentage of the boundary that is transform fault (as well as the other observations tabulated by Forsyth and Uyeda (1975)) are uncorrelated with plate velocity. (This figure is a compact representation of the graphs shown in figures 5–9 of Forsyth and Uyeda (1975).)

7.08.2 Slab Geometry

Historically, the geometry of subducted slabs has been constrained by earthquake hypocenters and focal mechanisms (e.g., Isacks et al., 1968; Isacks and Molnar, 1969). Figure 2 illustrates a global summary of the distribution of down-dip compressive stresses in subduction zones (Isacks and Molnar, 1971). While slab tomography is playing an increasingly important role in understanding slab structure, it does not provide information on the stress field. Slab geometry is variable, and significant attention has been given to correlating slab dip with a variety of observables related to subduction zones. A seminal compilation of 26 parameters measured or estimated for 39 modern subduction zones was undertaken by Jarrard (1986). Using these observations Jarrard performed a multistep correlation analysis, finding that three independent variables — the age of the oceanic plate at the trench, the convergence rate of the plates at the trench (based on global plate reconstructions, e.g., Chase, 1978; Minster and Jordan, 1978), and intermediate slab dip (the dip angle between the trench and the slab at 100 km depth) — correlate significantly with the other observations. There is an excellent correlation between the slab length and the product of the convergence rate and the age of the downgoing slab, consistent with the predictions of some slab thermal models (Molnar et al., 1979). This product
has become known as the slab thermal parameter (Kirby et al., 1991, 1996).

We summarize the subduction zone parameter correlations by plotting a covariance matrix with nine of the best-constrained observations from Jarrard’s compilation plus the slab thermal parameter (slab age times convergence velocity) in Figure 3. There is also a strong positive correlation between the age of the plate at the trench and the age calculated at the slab tip, which is expected if the age of the slab tip is a linear function of the age of the incoming plate. This would be the case if the slab was rigid and the fact that this correlation is high is one indicator that slab deformation is limited. The other striking correlation is that between the age of the arc and the intermediate slab dip (trench to 100 km depth) and the deep slab dip (the average of the slab dip between 100 and 400 km). Given the attention focused on trench migration, it is worth pointing out that the analysis shows at best a very weak correlation between slab dip and trench migration.

The relationship between trench rollback and slab age has been examined in a number of studies with conflicting results. Jarrard (1986) finds as many advancing as retreating trenches while Garfunkel et al. (1986) find that trench rollback, or retreat, predominates. Slab dynamics and back-arc deformation style have recently been re-examined (Heuret and Lallemand, 2005) using relocated hypocenters (Engdahl et al., 1998), updated plate ages (Mueller et al., 1997), and an updated global plate velocity model (Gripp and Gordon, 2002). After excluding regions where continental lithosphere, volcanic arc crust, or oceanic plateaus are being subducted, they divide the oceanic subduction zones into 159 segments with a uniform interval of two degrees of trench length. Heuret and Lallemand (2005) find as many advancing as retreating trenches with slab motion occurring in two distinct populations: one with essentially zero trench velocity and one with a significant (−50 mm yr⁻¹) component of retreat. As one might expect, they find back-arc extension when...
the overriding plate is retreating from the trench and back-arc compression when the overriding plate is advancing. The correlation between the absolute velocity of the overriding plate and the style of back-arc deformation from this new compilation is consistent with Jarrard (1986). However, when correlating slab motion and age, Heuret and Lallemand’s result is opposite of the result seen in many numerical models; on Earth older subducting plates are observed to be advancing, not retreating as is the case in many dynamic subduction calculations. Heuret and Lallemand (2005) suggest that slabs are partially anchored because of the slabs’ resistance to large-scale lateral flow (e.g., Gurnis and Hager, 1988; Griffiths et al., 1995; Scholz and Campos, 1995; Becker et al., 1999). The degree of variability in the data suggests that local flow effects, such as small-scale flow in the vicinity of retreating slab edges or tears or more regional-scale flow associated with the shrinking of the Pacific plate area (e.g., Garfinkel et al., 1986) may significantly impact slab–trench dynamics.

Following a slightly different approach, Cruciani et al. (2005) re-examine the global correlation of slab dip with subduction zone observations using the regionalized upper mantle (RUM) seismic model (Gudmundsson and Sambridge, 1998) to map out the slab geometry rather than using relocated earthquake hypocenters, as both Jarrard (1986) and Heuret and Lallemand (2005) did. Cruciani et al. (2005) construct 164 trench-perpendicular cross-sections through the RUM model using a digital seafloor age model for the age of the incoming slab (Mueller et al., 1997) and a global plate velocity model for the convergence velocity (DeMets et al., 1994). There is more scatter in the Cruciani et al. (2005) compilation than the original compilation by Jarrard (1986), possibly because of the use of a seismic model as opposed to the more direct focal mechanism data, but more likely reflecting the actual complexity of slab geometries. The product
of slab age and plate convergence rate (thermal parameter) correlates with slab dip better than either slab age or plate convergence rate individually, although the correlation is weak (correlation coefficient of 0.45), and suggests that forces beyond the local subduction environment may be important. Hager and O’Connell (1981) show that they could produce a reasonably good fit to slab dips simply with global plate motions and large-scale return mantle flow, without a detailed model of the deformation in the subduction zone, suggesting that global-scale mantle flow may play an important role in slab geometry.

As one example where deep mantle flow is almost certainly impacting the deformation of a slab, Gurnis et al. (2000a) show that the Tonga slab is being deformed by a large-scale upwelling associated with the Pacific superplume. There is no substantial aseismic penetration into the lower mantle beneath Tonga, consistent with initiation of subduction during the Eocene. Both the pattern and amount of deformation and the seismic energy release in the Tonga slab show that it is deforming faster and has accumulated more deformation than any other slab. Gurnis et al. (2000a) argue that the strong deformation of the Tonga–Kermadec slab in the transition zone is a result of the short subduction history (40 My) and the upward flow in the lower mantle from a broad upwelling associated with the Pacific superplume. The lack of deep aseismic extension of the Tonga slab into the lower mantle is consistent with the short subduction history and the deep mantle flow. While recognizing the importance of regional plate motions, especially the opening of the Lau basin, Gurnis et al. argue that the structure and geometry of the Tonga–Kermadec slab is best explained with controls from deep and shallow forces.

### 7.08.2.1 Slab Structure

In addition to studies of subducted slab dip, numerous studies illustrate that slabs bend, kink, and thicken. These include the location of Wadati–Benioff zones.
of a subducting slab changes over a narrow depth interval, from 10–20° in the interplate 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 represented 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–Aleutian, Sumatra–Java–Flores, Caribbean, Scotia, Ryukyu, and Mariana arcs undergo a minimum of 10% strain (Bevis, 1986; Giardini and Woodhouse, 1986).

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° vs 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. 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 overriding oceanic plates, it is not clear whether this correlation once again reflects the effect of trench migration or mechanical differences in the overriding plate.

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 slabs reach the top of the lower mantle, slab-dip angles become progressively shallower if there is oceanward trench migration because the deep slab becomes anchored in the higher-viscosity lower mantle. The numerical and laboratory experiments observe trench rollback, while compilations of slab geometry show that older slabs are advancing, not retreating (e.g., Heuret and Lallemand, 2005). It is interesting to note that the strongest and only significant correlation in the data set compiled by Jarrard (1986) is the correlation of the duration of subduction with dip of the deepest part of the slab, defined by Jarrard as 100–400 km. While it is not possible to isolate this from the other factors that 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.

### 7.08.2.2 Slab Structure Seen from Refracted and Scattered Waves

At high frequencies (0.5–10 Hz) seismic waves are particularly sensitive to the presence and state of subducted oceanic crust. Both S and P waves that travel long distances along the top of the slab appear dispersed (e.g., Abers and Sarker, 1996; Abers, 2000) and the dispersion is consistent with a constant-velocity waveguide of constant thickness. Observations in all North Pacific subduction zones suggest that crust remains distinct to depths of 150–250 km. The signals show waveguide behavior at the scale of a few kilometers: short-wavelength, high-frequency energy (≥3 Hz) is delayed 5–7% relative to that of low frequencies (≥1 Hz), systematically. Velocities in a low-velocity layer 1–7 km thick, likely subducted crust, must remain seismically slow relative to surrounding mantle at these depths to explain the observations. The inferred velocities are similar to those estimated for blueschists, suggesting that hydrous assemblages persist past the volcanic front.

Abers (2005) solves for traveltimes for waves traveling through a low-velocity channel with variable velocity with depth. Using both S and P waves from events deeper than 100 km recorded at Global Seismic Network (GSN) stations between 1992 and 2001, Abers compiles a data set of 210 P waves and 156 S waves for study and finds that the data from seven circum-Pacific arcs (Aleutian, Alaska, Hokkaido–S. Kurile, N. Honshu, Mariana, Nicaragua, and Kurile–Kamchatka) are consistent with a waveguide extending to greater than 150 km depth with a low-velocity channel 28 km thick with anomalies as large as dln $V_p = 14\%$ compared with surrounding mantle, although there is substantial variability with depth. Below 150 km the waveguide velocity is close to the surrounding mantle velocity. The low velocities are consistent with either low-temperature hydrated mafic rocks or metastable gabbros. The waveguide velocities also vary substantially from arc to arc, correlating with slab dip but not with other subduction-related parameters (e.g., plate age, convergence velocity, thermal parameter). Abers favors fluids or hydrous minerals as opposed to variable metastability of anhydrous gabbro as an explanation pointing out that the variability with dip is consistent with fluids from shallow slabs being driven into the mantle wedge while for steeper slabs the fluid preferentially migrates up the sediment layer.

### 7.08.2.3 Slab Structure Seen from Seismic Tomography

With an increasing number of tomographic studies in subduction zones (Zhou and Clayton, 1990; Van der Hilst et al., 1991, 1993; Chiu et al., 1991; Fukao et al., 1992; van der Hilst and Seno, 1993; Van der Hilst, 1995; Gudmundsson and Sambridge, 1998; Widiyantoro et al., 1999; Miller et al., 2004; 2005) it is becoming clear that subducted slabs are not 2-D tabular features. For example, slab dip varies along the length of many arcs. In the case of the Cocos slab, the dip angle changes along the Middle American Trench from about 80° at the northern end, with earthquakes occurring to depths of about 200 km, to 60° in the central part, with earthquakes occurring to depths of about 125 km, and even shallower (dip and maximum earthquakes) to the south (Colombo et al., 1997). The age of the Cocos Plate, the relative velocity between the Cocos and North/South American Plates, the heat flow, and even the seamount distribution change along the length of the arc and all of these are known to impact slab thermal structure.

Seismic tomography has been used to image the slab beneath the Mariana arc penetrating vertically into the lower mantle, whereas beneath the Izu–Bonin arc the slab appears to be deflected horizontally on top of the 660 km discontinuity (Zhou and Clayton, 1990; Chiu et al., 1991; Fukao et al., 1992; van der Hilst and Seno, 1993; Gudmundsson and Sambridge, 1998; Widiyantoro et al., 1999; Miller et al., 2005). Recent analysis of tomographic slab models (Miller et al., 2005) shows a distinct change in the seismic velocity in the Pacific slab beneath the Izu–Bonin arc at 300–450 km depth, in a position north and west of the Ogasawara Plateau at the southern end of the Izu–Bonin arc. The change in morphology of the slab could be interpreted as a ‘slab tear’. Beneath northern Tonga, a complex morphology is observed with a seismically fast anomaly lying subhorizontally above the 660 km discontinuity for almost 1000 km before (apparently) descending into the lower mantle (Van der Hilst, 1995). Flat-lying fast seismic-velocity anomalies at the base of the transition zone have been found beneath Japan, Java, and Izu–Bonin trenches (van der Hilst et al., 1991, 1993), and these are interpreted to be the extension (often aseismic) of ongoing subducting slabs.
The pattern of deep seismicity beneath Tonga provides yet another illustration of the complexity in subduction zones (Hamburger and Isacks, 1987; Fischer et al., 1991; van der Hilst, 1995; Chen and Brudzinski, 2001, 2003; Brudzinski and Chen, 2003a, 2003b; see Chapters 4.11 and 4.16). There is a group of well-located deep events that lies about 200 km above the deepest end of the Tonga slab, making it difficult to connect this cluster of earthquakes with the actively subducting lithosphere. If indeed this cluster of earthquakes represents a remnant slab, it is not clear that it is attached to the current Tonga slab. A possible explanation for the origin of a detached slab is subduction of the Pacific Plate along the fossil Vitiaz trench where subduction ceased about 5–8 Ma (Hamburger and Isacks, 1987; Fischer et al., 1991; Okal and Kirby, 1998; Chen and Brudzinski, 2001). Although another interpretation is that this cluster of earthquakes represents a subhorizontal extension of the Tonga slab along the subduction zone (Fukao et al., 1992; van der Hilst et al., 1993). The interaction of the Tonga slab with a mantle upwelling (Gurnis et al., 2000a) is likely to be a significant factor (perhaps in combination with the other events). With either explanation, it is clear that the interaction of subducted material with the transition zone can yield complex planforms of subducted material.

Another region of slab complexity is the South American slab, in the vicinity of the 1994 deep Bolivian earthquake (Engdahl et al., 1995). Tomographic images of the deep structure of the Andean subduction zone beneath western South America indicate that the Nazca slab is continuous both horizontally and with depth over most regions; in contrast, the Nazca slab penetrates the lower mantle beneath central South America, but is partly deflected to the south. There are two regions of deep earthquakes (a northern and southern zone) separated by the region containing the 1994 deep Bolivia main shock. The region between the north and south deep earthquake zones is modeled as a northwest-striking and steeply northeast-dipping slab structure.

A summary of tomographic cross-sections across Pacific subduction zones (Figure 4) from Karason (2002) illustrates both a variety of slab planforms and the general broadening of slabs in lower mantle. It is possible that the general appearance of thinning of the slab at the top of the transition zone is a result of the seismic properties of the mantle in this region, rather than reflecting a thinning of the slab itself. Without independently solving for the topography on the phase boundary itself, traveltime anomalies due to the deflection of the phase boundary will be mapped into volume perturbations in the 3-D seismic

Figure 4  Summary of seismic tomography P-wave cross-sections through Pacific subduction zones. Reproduced from Albarède F and van der Hilst RD (2002) Zoned mantle convection. Philosophical Transactions of the Royal Society of London A 360: 2569–2592, with permission from Royal Society.
model. Because the jump in velocity at the phase boundary is on the order of 5%, this is a significant contribution to the total traveltime anomaly. Furthermore, a difference between the reference model and the average velocity with depth will lead to the apparent thickening or thinning of vertical anomalies when plotting residual anomaly cross-sections. Thus, especially near phase boundaries, caution must be exercised when interpreting seismic cross-sections. It is also important to bear in mind that this kind of global compilation does not capture the along-strike variability within a single arc.

Deal and Nolet (1999) construct a high-resolution 3-D P-wave velocity model beneath the northwest Pacific. Assuming the positive velocity deviations in the subducting lithosphere are to first order due to a temperature anomaly, they construct a theoretical slab temperature profile using an approach based on the appendix of Davies and Stevenson (1992) and convert this to a synthetic slab velocity model using \( V_p = V_p \frac{C}{C_0} \). They then invert this model with the observed tomographic model, to estimate the slab thickness and mantle potential temperature, obtaining thickness estimates of 84–88 km ± 8 km for the Izu-Bonin, Japan, and Kuril slabs. The uncertainty in the thickness is dominated by the uncertainty in the mantle temperature and the estimated variability of 8 km is a result of their estimated 100\% uncertainty in mantle temperature. The initial tomography result is then modified to closely resemble the synthetic slab tomogram with a method that guarantees that the new tomography model will resemble the synthetic slab tomogram with a method that guarantees that the new tomography model will satisfy the original seismic delay times. The modified slab images show continuous and narrow slabs compared to the initial tomographic results.

Deal et al. (1999) use the same approach in Tonga and the theoretical temperature model gives an optimal slab thickness of 82 km for a region near 29° S in Tonga. The final image shows a very narrow and continuous slab with maximum velocity anomalies of the order of 6–7\%; many of the gaps within the slab, as well as artifacts around the slab which were present in the minimum-norm solution, are absent in the biased image. One can deduce from these experiments that some of the gaps seen in seismic tomography images are not required by the seismic data and that a more continuous slab model is equally acceptable. A different and slightly less rigorous approach was taken by Zhao (2001) who used a slab model \textit{a priori} as a part of the starting model. Because the final model retained the signature of the original slab starting model, one can conclude that the traveltime data were consistent with the starting model, or at least that there was insufficient information in the seismic data to require a modification to the starting model.

### 7.08.2.4 Shear-Wave Splitting in Subduction Zones

Perhaps the most challenging seismic observations to reconcile with flow models are the estimates of seismic anisotropy, as measured by shear-wave splitting observations (Savage, 1999). Seismic anisotropy in the mantle is generally assumed to be the result of lattice-preferred orientation (LPO) of mantle minerals such as olivine (see Chapter 2.16). In the absence of water, olivine \textit{a}-axes (the seismically fastest axis) are assumed to be aligned with the direction of flow in the dislocation creep regime (e.g., Karato and Wu, 1993). However, recent experimental studies have suggested that the olivine slip system changes under high stress in hydrated systems (Jung and Karato, 2001). Clearly, subduction zones are likely to be hydrated and these new laboratory results complicate the interpretation of the shear-wave splitting observations.

Shear-wave splitting observations include a fast polarization direction, reflecting the orientation of fabric, and a splitting time, reflecting the organization and strength of the fabric. As is the case with the other seismic observations, subduction zones exhibit a variety of shear-wave splitting orientations along a single arc. These have been summarized in global compilations, such as Figure 5 taken from Lassak et al. (2006).

A number of studies have used mantle flow models to predict the LPO development and resulting shear-wave splitting (e.g., Fischer et al., 2000; Fouch et al., 2000; Hall et al. 2000; Kaminski and Ribe, 2001; Lassak et al., 2006). These studies map the LPO for mantle mineralogies by either orienting crystallographic axes to the local flow direction or calculating finite-strain directions and using them as a guide for the orientation of crystallographic axes. Kaminski and Ribe (2001) have developed a model for the evolution of LPO in olivine aggregates that accounts for intercrystalline slip and dynamic recrystallization.

In the Tonga subduction zone (Bowman and Ando, 1987; Fischer et al., 2000; Smith et al., 2001), shear-wave splitting results vary from trench-parallel fast directions near the trench to convergence-parallel fast directions closer to the backarc. Jung and Karato (2001) suggest that hydration of olivine may influence the mantle close to the trench and modeling of anhydrous–hydrorous transition predicts a fast polarization direction pattern similar to the observed
range for Tonga, but are not able to match the amplitude (Lassak et al., 2006). The mantle wedge flow associated with the Tonga slab is complex, due to complexity in the slab that is interpreted as a slab tear in the north and the possible influence of the Samoa plume (Smith et al., 2001).

In the Japanese subduction zone, shear-wave splitting observations also range from trench–parallel near the trench to convergence-parallel in the back-arc (Hiramatsu and Ando, 1996). The Japan subduction system is complex with two triple junctions and multiple subducting slabs. In addition, crustal anisotropy may be responsible for some local variability of splitting variations (e.g., Fouch and Fischer, 1996); however, there is agreement that mantle anisotropy is required to explain the majority of the shear-wave splitting results across the region (e.g., Fouch and Fischer, 1996; Nakajima and Hasegawa, 2004; Long and van der Hilst, 2005). Near Hokkaido, Nakajima and Hasegawa (2004) observe splitting variations that are consistent with north–south shear in the overriding plate or the presence of a hydrated mantle wedge. Further south near Honshu and Ryukyu, Fouch and Fischer (1996) and Long and van der Hilst (2005) observe similar shear-wave splitting directions with fast directions that rotate from trench-parallel near the trench to trench-orthogonal into the back-arc. The splitting directions are more consistent than the splitting times, with no clear systematics from the splitting times. While most studies conclude that the Japan subduction zone shear-wave-splitting results are consistent with a hydrated wedge (Kneller et al., 2005), the modeling leaves plenty of room for alternatives.

In the Kamchatka subduction zones, shear-wave splitting results once again range from convergence-parallel near the trench to trench–parallel toward the back-arc (Peyton et al., 2001; Levin et al., 2004). Levin et al. (2004) interpret these measurements as the result of a complex flow regime with mantle flow moving around the northern bending edge of the downgoing slab due to slab loss/detachment, possibly combined with deformation of material in the wedge. Lassak et al. (2006) suggest that hydration of the wedge may also be a possible explanation. South America also exhibits a range of shear-wave splitting directions (Russo et al., 1994; Polet et al., 2000; Anderson et al., 2004).

Yang et al. (1995) inferred possible strain geometries in the mantle beneath the eastern Aleutians by

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**Figure 5** Summary of shear-wave splitting observations around the Pacific Rim illustrating the variety of observed fast directions near subduction zones from Lassak et al. (2006). Double-headed vectors indicate the average regional orientation of the fast polarization direction. Many regions exhibit fast direction populations with orientations both parallel and orthogonal to local trench strike. For details see Lassak et al. (2006). From Lassak TM, Fouch MJ, Hall CE, and Kamiknsi E (2006) Seismic characterization of mantle flow in subduction systems: Can we resolve a hydrated mantle wedge? Earth and Planetary Science Letters 243: 632–649.
mapping seismic anisotropy with shear-wave splitting. Their observed splitting parameters are well matched by predicted values for olivine-rich mantle wedge models with 1% SV anisotropy where the olivine $b$-axis is arc-orthogonal and the $a$-axis is vertical (no arc-parallel strain) or arc-parallel (no vertical strain). These results are consistent with mantle wedge strain models in which arc-normal compression is accompanied by arc-parallel or vertical shearing or extension. Mehl et al. (2003) describe residual mantle exposures in the accreted Talkeetna arc, Alaska, that provide the first rock analog for the arc-parallel flow that is inferred from seismic anisotropy at several modern arcs. Stretching lineations and olivine [100] slip directions are subparallel to the Talkeetna arc for over 200 km indicating that mantle flow was parallel to the arc axis. Slip occurred chiefly on the (001) [100] slip system, rarely the dominant slip system observed in olivine, and the alignment of the olivine [100] axes yields a calculated S-wave anisotropy with the fast polarization direction parallel to the arc.

The key point to be made here is that the shear-wave splitting observations, like many of the other seismic observations discussed above, appear to point to a flow environment in subduction zones that is more complex than the simple corner flow model that has been envisaged for the past 30 years. It is possible that some of the interpretation of the shear-wave splitting observations may actually be better explained by less complexity in the flow model and recognition that shape-preferred orientation, due to lenses or veins of melt (e.g., Spiegelman, 2003) or hydration (e.g., Jung and Karato, 2001; Lassak et al., 2006), can also give rise to shear-wave splitting.

Given the complexity of the slab geometry and the petrologic variations introduced by the phase transformations (and possible structure due to nonequilibrium phase transformations), the interpretation of these measurements is difficult. Seismic studies in subduction zones are extensively reviewed in Chapter 4.11 in this series.

### 7.08.3 Slab Modeling and Thermal Structure

Slab thermal modeling takes a complex region of the Earth where the downgoing plate descends into the mantle and attempts to capture this in a mathematically tractable manner. Typically slab thermal models assume a thermal structure for the incoming plate, some form of coupling between the subducting and overriding plates in the ‘seismogenic zone’, and some assumptions about the flow in the mantle wedge (Figure 6). In most slab thermal models the slab is assumed to be rigid and descends with a constant dip. Thermal models differ as to assumptions regarding flow in the mantle wedge; however, the largest difference comes from a variety of assumptions (sometimes not clearly stated) regarding shear heating and mechanical coupling at the slab and overriding-plate interface are used in slab thermal models. Because of the limited domain of slab thermal models (both horizontally and vertically), it is not possible to calculate geoid or dynamic topography anomalies from slab thermal models. This is one motivation for multiscale models, which use a coarse global model and a refined model near the subduction zone (e.g., Billen et al., 2003).

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There are two end-members for modeling subducted slabs: kinematic and dynamic subduction models. Most present-day kinematic models make the assumption that slabs are (largely) 2-D and rigid and are generally used for detailed slab thermal structure calculations. They are quite often used in seismic, petrologic, or mineral physics investigations because it is possible to control the slab geometry a priori (e.g., Minear and Toksoz, 1970; Molnar et al., 1979; Helffrich et al., 1989; Peacock, 1991, 1996; Staudigel and King, 1992; Peacock et al., 1994; Kirby et al., 1996; Bina, 1996; 1997; Marton et al., 1999, 2005). When using a kinematic model, the velocity of the slab is imposed and it is only the careful use of initial and boundary conditions that ensures that the slab velocity is consistent with the buoyancy that provides the driving force for slab motion. The majority of the P–T–t (pressure–temperature–time) curves used by petrologists are taken from kinetic slab models (e.g., Peacock 1996, 2003). While kinematic models have been used to predict the lateral variability of phase transformations in slabs, these models cannot address the dynamic impact of phase transformations (see Chapters 7.02 and 2.06) on slabs (e.g., do phase changes retard slab motion?) because the motion of the slab is prescribed. In addition, the deformation that occurs as a plate bends, breaks, and generally deforms as it passes through a subduction zone (e.g., Conrad and Hager, 1999; Buffet and Rowley, 2006) is not accounted for in kinematic models, except in so far as the model prescribes deformation by imposing a curved slab or variation in velocity. With the clear recognition of changes in slab dip, thickening of the slab, possible tears in the slab, and along-slab flow, it seems that there is an increasing need for models with the resolution of kinematic models and the ability for the slab to deform without a priori specification. One important result of subduction zone thermal structure studies is that the temperature structure that develops in the upper 50–100 km of the slab is one of the primary controls of the structure of the deep slab, where we take ‘deep slab’ to mean 100–400 km as in the definition used by Jarrard (1986); that is, whatever else happens to the slab as it subducts, the thermal structure that develops in the upper 100 km is the most important controlling factor. The assumed rate of shear heating has a profound effect on the slab thermal structure, with temperatures on the slab–wedge interface varying by several hundred degrees at 100 km depth (cf. Peacock, 1992, 1993, 1996).

Dynamic subduction zone models often use mantle convection codes, in which the thermal (and sometimes chemical) buoyancy of the slab drive the motion of the slab and the larger-scale mantle flow. These models can include viscous faults, weak zones, complex, nonlinear rheologies, and compositional variation within the slab. In dynamic slab calculations, 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 a specific subduction zone. Dynamic slab models are also generally lower resolution than kinematic models and cover a larger region of the mantle by necessity. The numerical resolution and the difficulty in reproducing specific subduction zone geometries have inhibited the use of dynamic slab models in seismic, petrologic, or mineral physics investigations where detailed slab thermal structure models are needed.

Dynamic slab models have been used to address the mechanisms of slab dip (Gurnis and Hager, 1988), the interaction of slabs with phase changes (Chrsitensen and Yuen, 1984; Zhong and Gurnis, 1994: Christensen, 1995; 1996; Ita and King, 1998), the geoid and topographic profiles over subduction zones (e.g., Davies, 1981, 1984, 1986; King and Hager, 1994; Zhong and Gurnis, 1995; Chen and King, 1998; Zhong and Davies, 1999), and the mass transfer between the upper and lower mantle (e.g., Davies 1995, Christensen, 1996; Kincaid and Olson, 1987; Kincaid and Sacks, 1997; King et al. 1997; Ita and King, 1998). There has been some effort to compare the thermal structures from buoyancy-driven slabs and kinematic slab models (Peacock, 1996; van Keken et al., in preparation).

7.08.3.1 Kinematic Models and Slab Thermal Structure

The first slab thermal structure model, presented by McKenzie (1969), solves for the thermal structure within the slab assuming a hot, uniform-temperature mantle surrounding the slab. Beginning with the conservation of energy (see Chapter 7.02), the temperature within the slab is given by

\[ \rho C_p \left( \frac{\partial T}{\partial t} + \mathbf{v} \cdot \nabla T \right) = \nabla (k \nabla T) + H \]  \hfill [3]

where \( \rho \) is the density of the slab, \( C_p \) is the coefficient of specific heat, \( T \) is the temperature, \( \mathbf{v} \) is the slab velocity, \( k \) is the thermal conductivity, and \( H \) is the rate of internal heat generation. Ignoring radioactivity in the slab and assuming that the slab is in thermal...
equilibrium with the surrounding mantel, eqn [1] reduces to

\[ \rho C_p v \cdot \nabla T = \nabla (k \nabla T) + H \]  

Substituting characteristic values for the temperature difference between the mantle and slab, \( T_{\infty} \) and the length scale of the problem, \( l \) (e.g., the width of the slab) (see Chapter 7.02)

\[ T = T_{\infty} + x', \quad z = z' \]  

then aligning the coordinate system such that the \( x \)-axis is along the down-dip slab direction and the \( z \)-axis is perpendicular, we can write eqn [4] as

\[ \frac{\partial^2 T'}{\partial x'^2} - 2R \frac{\partial T'}{\partial x'} + \frac{\partial^2 T'}{\partial z'^2} = 0 \]  

where \( R \) is the Thermal Reynolds number, given by

\[ R = (\rho C_p v l)/2k \]

Note that \( v \) becomes \( v_x \) in this coordinate system and \( v_z \) is zero because the assumption is that the slab is 2-D and does not deform. If we assume the temperature at the top and bottom of the slab maintained at a uniform temperature by the surrounding mantle, the solution can be simplified to

\[ T' \approx 1 - \frac{2}{\pi} \exp \left( -\frac{\pi x'}{2R} \right) \sin \pi z' \]  

There are several generalizations of kinematic slab thermal models that are illustrated quite clearly by this derivation. First, the mantle surrounding the slab is assumed to be at a uniform temperature and is not locally impacted by the slab. Second, the slab is assumed to descend at a uniform rate that is always parallel to the dip of the slab everywhere (e.g., no slab buckling or internal strain). Finally, the slab does not exchange mass with the surrounding mantle (e.g., no thermal erosion, no melting or dehydration).

Using this model, McKenzie concluded that the maximum depth of the deepest earthquakes in the Tonga–Fiji–Kermadec arc coincide with the depth of the 680° C isotherm in the model. While he avoided attaching a significance to the exact value of the isotherm, noting the simplicity of the calculations, he suggested a thermal control to the maximum depth extent of deep earthquakes, a suggestion which has been subsequently questioned (e.g., Chen et al., 2004). In addition, he uses this the thermal structure to calculate a buoyancy force due to the cold subducting slab which he estimated at

\[ f = 2.5 \sin \phi \text{ kbars} \]

Some of the limiting assumptions in the McKenzie slab model are addressed in the numerical model by Minear and Toksöz (1970). In the Minear and Toksöz model, the slab material moves with a fixed velocity relative to a stationary mantle and the computational scheme consists of translating the temperature field along the slab-dip direction and allowing the slab to equilibrate over a fixed time interval. As they state in the text, “in essence, we have assumed the dynamics and have computed the temperature field given this motion field.” An appealing aspect of the Minear and Toksöz model is that the slab geometry, age of the incoming ocean lithosphere, and slab velocity are specified, making it a fairly straightforward exercise to model observed subduction zone geometries. In addition, the effect of additional sources of energy such as adiabatic compression, phase transformations, and shear heating could be incorporated in the slab thermal structure through the imposed heat source term, \( H \), in eqn [3].

Based on their resolution analysis, Minear and Toksöz (1970) estimate that the temperatures resulting from these computations “are probably in error by less than 10%.” Because this model enables researchers to calculate the thermal structure of a subducting slab using geophysically observable constraints such as slab dip, incoming plate age, and plate velocity, the Minear and Toksöz model has been extensively used (e.g., Spiegelman and McKenzie, 1987; Helffrich et al., 1989; Iita and Stixrude, 1992; Stein and Stein, 1996; Kirby et al., 1996; Bina, 1996, 1997; Marton et al., 1999, 2005; Wiens, 2001).

Another kinematic approach to the subduction problem is to use the corner flow solution (e.g., Batchelor, 1967; Stevenson and Turner, 1977; Turcotte and Schubert, 1982; Peacock et al., 1994) for the velocity field of the subducting slab. This is sometimes referred to as the ‘Batchelor solution’ after the original reference (i.e., Batchelor, 1967). With the corner flow solution, the velocity field of the slab and the surrounding mantle, including the mantle wedge, satisfy the equations of motion for a constant-viscosity fluid. This solution uses the stream-function formulation (see Chapter 7.04) to solve the equations of motion assuming a uniform velocity, rigid slab and plate. As unlikely as it may seem, it is possible to write the stream function, \( \psi \), in the form

\[ \psi = Ax + Bz + (Cx + Dz) \arctan \left( \frac{Y}{X} \right) \]  

where \( A, B, C, \) and \( D \) are constants to be determined by the boundary conditions. Because of the geometry
of the descending slab, there are two distinct stream functions, that is, two sets of constants to be solved for in this problem: one above and one below the slab. The general problem leads to quite complicated integrals; however, a slab dip of \( \pi/4 \) allows easy evaluation of the integrals. Solving for the solution in the mantle wedge, we have the following boundary conditions

\[
  u = v = 0 \quad \text{on} \quad z = 0, \quad x > 0 \quad \text{or} \quad \arctan \frac{z}{x} = 0 \quad [9]
\]

\[
  u = v = \frac{U \sqrt{2}}{2} \quad \text{on} \quad z = x \quad \text{or} \quad \arctan \frac{z}{x} = \frac{\pi}{4} \quad [10]
\]

By straightforward substitution, it is possible to show that the pressure in the arc corner is

\[
  p = \frac{\mu U \sqrt{2(\pi x + (4 - \pi)z)}}{(2 - \pi^2/4)(x^2 + z^2)} \quad [11]
\]

We can evaluate this expression using the fact that \( x = z = \sqrt{2}/2 \) where \( r \) is the length along the slab and we find that

\[
  p = \frac{4\mu U}{(2 - \pi^2/4)r} \approx \frac{-8\mu U}{r} \quad [12]
\]

The negative pressure means the back-arc region flow tends to lift the slab against the force of gravity. Notice that the pressure varies as \( 1/r \) along the slab, with a singularity at the trench. The total lifting torque on the slab will be \( r \times P \), where \( P \) is a constant.

For the ocean side of the trench, the boundary conditions are

\[
  u = U, \quad v = 0 \quad \text{on} \quad z = 0, \quad x < 0 \quad \text{or} \quad \arctan \frac{z}{x} = \pi \quad [13]
\]

\[
  u = v = \frac{U \sqrt{2}}{2} \quad \text{on} \quad z = x \quad \text{or} \quad \arctan \frac{z}{x} = \frac{\pi}{4} \quad [14]
\]

Again by straightforward substitution, the pressure on the bottom side of the slab is given by

\[
  p = \frac{\mu U}{r} \left( \frac{3\pi \sqrt{2} - 4}{9\pi^2/4 - 2} \right) \approx 0.5\mu U \quad [15]
\]

Given the geometry and the signs of the pressures, this flow also exerts a lifting torque on the slab, that is about a factor of 20 smaller than the torque on the slab from the arc flow (Stevenson and Turner, 1977). The corner flow solution does not overcome the ridge slab approximation nor does it allow for a slab velocity that is consistent with the slab driving force. As is shown below, the effect of a temperature-dependent viscosity in the mantle wedge part of the flow has a significant effect on the flow near the corner and a significant effect on the slab thermal structure. This has reduced the difference between estimates of slab temperature from flow calculations and petrology.

While the stream-function solution solves the equations of motion, a solution to the energy equation is also needed and several different solution strategies have been employed (Peacock, 1996). Because of the discontinuity in the velocity boundary conditions at the corner of the wedge, there is a singularity in the pressure field that presents problems for fluid/melt flow models (e.g., Spiegelman and McKenzie, 1987).

Figures 7 and 8 illustrate results of two different kinematic slab thermal models: the first based on the method described in Minear and Toksöz (1970) and the analytic corner flow solution described above plus the Streamline Upwind Petrov–Galerkin solver from ConMan (King et al., 1990) for the temperature (energy) equation. The problem is a 50-Myr-old slab descending 45° at 5 cm yr⁻¹ and the complete list of parameters for these calculations is provided in Table 1. There are two reasons for the significant difference between the solutions shown in Figures 7 and 8. First, in the Minear and Toksöz (1970) thermal model the slab curves (bends) as it enters the subduction zone, whereas for the corne flow solution, the slab simply descends at 45° at the left edge of the box. This accounts for the difference between the two solutions seen in the top 50 km. Second, the Minear and Toksöz (1970) thermal model does not have flow in the wedge, whereas this is a significant component of the corner flow solution with the ConMan result. This accounts for the majority of the difference in the region below 50 km. In the Minear and Toksöz (1970) thermal model, perpendicular to the slab heat is transferred by conduction only. (There is an induced flow approximation in the Minear and Toksöz code, but this does not improve the agreement between the solutions.) The importance of the induced wedge flow on the slab thermal structure will become more significant in the discussion of temperature-dependent rheology. Returning to the Minear and Toksöz claim that the solution is in error by less than 10% (due to the grid) it is important to recognize that this is a statement of the numerical solution on the grid which they used in their paper relative to the solution with the same basic underlying physical assumptions in their model on a more highly refined grid. From Figure 7, the difference between the Minear and Toksöz solution and the Batchelor solution plus advection is much more...
than 10% for the problem described above and as we will see when we move to temperature-dependent rheology the difference can be even greater still. In the ConMan solution, we use a grid spacing on the order of 1 km in the corner of the wedge. We find that grid spacing on this order is necessary to resolve the flow near the pressure singularity near the corner. In the real Earth, it is likely that because of temperature-dependent rheology and serpentinization of the region near the corner of the wedge (e.g., Bostock et al., 2002) that this singularity does not present the problem it does in trying to match the corner flow solution. However, for this problem, the resolution is important because the temperature of the slab is controlled by what happens in the upper 50–100 km.

Figure 7  A comparison of slab thermal structures from a 50-Myr-old, 45° dipping slab descending at 5 cm yr\(^{-1}\) with kinematic slab approximations (left) (Minear and Toksoz, 1970) and (right) the analytic corner flow solution for flow and ConMan’s temperature solver. Below 50 km depth, the Minear and Toksoz slab is colder than the corner flow plus ConMan solution.

Figure 8  Temperature along the line \(x = z\) (i.e., the slab interface) for the kinematic slab thermal models shown above. The difference of approximately 150 °C is largely due to the lack of corner flow in the Minear and Toksoz model.

Table 1 Parameters used in subduction zone thermal models shown in Figures 7 and 8.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Symbol</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Slab velocity</td>
<td>(v)</td>
<td>5 cm yr(^{-1})</td>
</tr>
<tr>
<td>Slab dip</td>
<td>(\Theta)</td>
<td>45°</td>
</tr>
<tr>
<td>Box depth</td>
<td>(d)</td>
<td>600 km</td>
</tr>
<tr>
<td>Box width</td>
<td>(w)</td>
<td>660 km</td>
</tr>
<tr>
<td>Slab age</td>
<td>(T_{slab})</td>
<td>50 My</td>
</tr>
<tr>
<td>Thickness of overriding plate</td>
<td>(d_{plate})</td>
<td>50 km</td>
</tr>
<tr>
<td>Density</td>
<td>(\rho)</td>
<td>3300 kg m(^{-3})</td>
</tr>
<tr>
<td>Thermal conductivity</td>
<td>(k)</td>
<td>3.0 W m(^{-1}) K(^{-1})</td>
</tr>
<tr>
<td>Specific heat</td>
<td>(c_p)</td>
<td>1250 J kg(^{-1}) K(^{-1})</td>
</tr>
<tr>
<td>Thermal diffusivity</td>
<td>(\kappa = k/(\rho c_p))</td>
<td>(7.272 \times 10^{-6} \text{ m}^2 \text{s}^{-1})</td>
</tr>
<tr>
<td>Reference temperature</td>
<td>(T_o)</td>
<td>1300 °C</td>
</tr>
</tbody>
</table>
A series of calculations described by Peacock et al. (1994) and Peacock (1996) solve the energy equation (eqn [3]) for the thermal structure of a subduction zone using a finite-difference formulation described in Peacock (1991). Peacock focuses on the petrologic implications of the slab thermal structure, as opposed to slab deformation and driving mechanisms of plate motions. The velocity of the slab is imposed and the analytic ‘Batchelor solution’ (Batchelor, 1967) is used for the corner flow in the mantle wedge. Beneath the volcanic arc, the thermal models predict temperatures in the oceanic crust in the range of 500–700°C. Peacock (1996) concludes that with the exception of very rare circumstances, the slab does not melt; however, dehydration of Na-amphibole, lawsonite, and chlorite in the subducted oceanic crust can trigger partial melting in the slab wedge. We will return to the petrology of slabs as well as the role of water later.

Kinematic slab models have been used to map the mineralogy of the slab (e.g., Ita and Stixrude, 1992; Bina, 1996, 1997; Marton et al., 1999, 2005). These mineralogical models produce significant differences in slab density, and changes in slab density lead to significant differences in the plate and slab velocities (e.g., King and Ita, 1995; Ita and King, 1998); however, the change in slab density is not accounted for in the flow field in the kinematic models because buoyancy forces do not drive slab flow in the kinematic slab approximation. While kinematic models have provided an estimate of where the transitions in mineralogy might occur, they do not have the predictive ability to follow the implications of the mineralogical models as the slabs evolve dynamically (e.g., Ita and King, 1998). We would like to be able, for example, to calculate whether metastable olivine would have an observable impact on subduction velocity, dip, and slab deformation. This kind of study has been carried out in larger-scale fluid models (e.g., Gurnis and Hager, 1988; Zhong and Gurnis, 1994; Christensen, 1996; Zhong and Gurnis, 1997; Ita and King, 1998). However, those calculations are not able to resolve detailed slab thermal structure with the same resolution as the kinematic models. In most dynamic models, the grid resolution is on the order of 10 km and the mechanics of the deformation near the trench are often grossly approximated by a fault zone (sometimes extending to depths of 100 km or more) or large, prescribed mechanically weak zone.

A kinematic description of slab motion remains popular in part because the input parameters are easily observed properties of subduction zones (i.e., plate age, plate velocity, and slab dip) and because dynamic slab models require a careful choice of boundary conditions in order to incorporate the effects of global mantle flow (cf. Billen and Gurnis, 2001; Billen et al., 2003). Until recently (Billen and Gurnis, 2001; Billen et al., 2003), convection models have not attempted to reproduce the major geophysical characteristics (e.g., geoid, topography, heat flow) of specific subduction zones, so kinematic models have been the only tools available to researchers. Physical experiments have shed new light on slab morphology and dynamics (e.g., Kincaid and Olson, 1987; Griffiths et al., 1995; Guillou-Frottier et al., 1995; Becker et al., 1999); however, this kind of modeling cannot provide detailed thermal structure information. A new hybrid form of thermal structure calculations (e.g., Peacock and Wang, 1999; van Keken et al., 2002; Peacock et al., 2005) has done an excellent job of matching the geometry and heat flow from specific subducting slab environments; however, these calculations are not able to make use of geoid or dynamic topography observations because they lack large-scale dynamic flow.

### 7.08.3.2 Hybrid Kinematic–Dynamic Slab–Wedge Models

A relatively new approach to subduction zone modeling is a combination of the kinematic and dynamic flow models. In this approach, the slab velocity is imposed (kinematic) and the energy equation (eqn [3]) as well as the equations for incompressible viscous flow are solved numerically (e.g., Davies and Stevenson, 1992; Furukawa, 1993; van Keken et al., 2002; Conder et al., 2002; Kelemen et al., 2003; Arcay et al., 2005; Manea et al., 2005; Peacock et al., 2005). While the slab motion is still imposed, this formulation enables greater flexibility in modeling the flow in the mantle wedge, specifically including the effect of temperature-dependent and/or stress-dependent rheology. Even with this new more dynamic approach, the coupling of the overriding plate with the subducting plate (i.e., the seismogenic zone) remains problematic.

Of particular importance has been the recognition that with the effect of temperature-dependent rheology, flow in the wedge corner is significantly stronger than that predicted by the isoviscous corner flow model, leading to warmer temperatures at the slab–wedge interface in the shallow slab (e.g., Furukawa, 1993; van Keken et al., 2002; Kelemen et al., 2003; Arcay et al., 2005; Manea et al., 2005; Peacock et al., 2005). Figure 9 illustrates the temperature field from...
two slab calculations using the convection code ConMan (King et al., 1990). The mesh is nonuniform with significant refinement near the corner of the wedge with a grid spacing of approximately 1 km. The grid extends to 660 km × 600 km and the boundary conditions along the inflow/outflow sides of the wedge are natural boundary conditions (i.e., normal and tangential components of total stress are zero). Both calculations have a slab descending at 45° at a fixed velocity of 5 cm yr⁻¹. The incoming plate material is 50 My old and the velocity of the overriding plate is set to zero to a depth of 50 km. The left image is a constant-viscosity mantle wedge and the right image is a temperature-dependent viscosity wedge, with an activation energy of 335 kJ mol⁻¹. The flow in the wedge in the temperature-dependent viscosity case extends upward along the slab to a depth of almost 50 km (the depth of the imposed zero-velocity plate) while the constant-viscosity flow has subhorizontal isothermal below the overriding plate. This difference has a significant effect on the thermal structure of the downgoing slab and the heat flow near the trench. The heat flow near the trench is elevated by 10% in the temperature-dependent case. Compared with the constant-viscosity calculation. The temperature in the downgoing slab differs by 98° at the slab–wedge interface at 150 km depth.

The higher temperatures near the slab–wedge interface warm the sediments and oceanic crust at the top of the slab and this may explain the conflicting geochemical evidence for high temperatures at the top of the slab based on sediment melting (Johnson and Plank, 1999) and low temperatures required by the Boron measurements in arc lavas (Leeman, 1996) as discussed in van Keken et al. (2002). The low viscosities in the mantle wedge (due to temperature, volatile content, and high strain rates) suggest that buoyancy in the mantle wedge may be dynamically important (e.g., Kelemen et al., 2003), although in the calculations above, we have not included the effect of buoyancy on flow in the mantle wedge. Buoyancy in the mantle wedge presents a challenge because of the nature of the stress-free boundary conditions at the wedge boundary and the stability of the overriding plate.

Because of our limited understanding of the processes in the seismogenic zone (see Chapter 4.11) and the variety of approximations used in the seismogenic zone, it is useful to compare the temperature along the seismogenic zone from the hybrid kinematic–dynamic models with other estimates of the temperature in this zone. Molnar and England (1990) develop analytic expressions for the steady-state temperature in the top 50 km of the subduction zone. Their expressions (eqns [16] and [23]) are

\[
T = \frac{(Q_0 + Q_{SH})z_t/k}{S}
\]  

[16]

where \(Q_0\) is the basal heatflux (in W m⁻²), \(Q_{SH}\) is the rate of shear heating (in W m⁻²), \(z_t\) is the depth to the
fault (m), \( k \) is the thermal conductivity (in W m\(^{-1}\) K) and \( S \) is given by

\[
S = 1 + b \sqrt{V z \sin \delta / \kappa}
\]

where \( b \) is a constant of order 1, \( V \) is the convergence velocity, \( \delta \) is the angle of subduction, and \( \kappa \) is the thermal diffusivity (m\(^2\) s\(^{-1}\)). Using this expression with values from Table 1 and \( Q_0 \) as 35 mW m\(^{-2}\), and plotting this with the temperatures from the temperature-dependent ConMan result in Figure 10, there is a good agreement with the analytic result with a shear stress of 40 MPa. The two ConMan results are for calculations with a ‘fault’ at the seismogenic zone, that is, velocities from the overriding plate and the slab are explicitly decoupled (black line), and a calculation where the velocity across the seismogenic zone varies continuously with the finite-element grid (red line). The difference is less than 10\(^{\circ}\), showing that the seismogenic zone formulation in this calculation does not have an overwhelming influence on the solution. To match the Molnar and England solution with the ConMan result without including shear heating requires an unrealistically high \( Q_0 \) value of 95 mW m\(^{-2}\). It is worth emphasizing that the ConMan formulation does not have an explicit shear heating term in the energy equation, so the need for a shear heating term in the Molnar and England (1990) formulation to match the ConMan solution illustrates that there is a difference between the solutions that has a similar effect as shear heating.

Manea et al. (2005) investigate the Guerrero subduction zone in Mexico, characterized by flat subduction with a distant volcanic arc. Using a temperature-dependent rheology to model the temperature structure in the slab and wedge, they find significantly higher temperatures below the volcanic arc compared with previous studies using an isoviscous mantle wedge (e.g., Currie et al., 2001). In Manea et al.’s model, temperatures at the slab–wedge interface are also sufficient to allow for a slab contribution to the arc volcanism from the melting of hydrated sediment and oceanic crust, in agreement with geochemical observations of magmas in the Central Mexican Volcanic belt. Manea et al. (2005) also investigate the effect of buoyant melt in the wedge using the predicted wedge flow structures. Using the location of the Popocatepetl volcano they estimate that magmatic blobs need to be larger than 10 km in diameter to have reasonable trajectories in the mantle wedge with realistic estimates for viscosity.

Peacock et al. (2005) construct high-resolution 2-D finite-element models using the geometry along four profiles perpendicular to the Central American subduction zone in Nicaragua and Costa Rica. Realistic stress- and temperature-dependent rheology in the mantle wedge is necessary to have sufficiently high temperature in the mantle wedge below the volcanic arc to produce melting. This subduction zone is characterized by rapid convergence of young oceanic lithosphere with minor changes in the convergence speed and age of the lithosphere along strike but significant variations in the depth of the seismicity and the geochemistry of arc volcanism. The small along-strike variations in the convergence speed (increasing to the southwest from 79 to 88 mm yr\(^{-1}\)) and age (decreasing from 24 to 15 My) nearly offset and the predicted slab surface temperatures are remarkably similar throughout the subduction zone. Comparing the thermal structures predicted with phase diagrams for mafic (crust) and ultramafic (uppermost mantle) compositions (Hacker et al., 2003) suggests that the predicted temperature in the uppermost 500 m of the subducting oceanic crust is high enough to generate partial melting if the oceanic crust and sediments are hydrated. The thermal models predict that dehydration is nearly complete beneath southern Costa Rica, but that a significant

![Figure 10](image-url)
amount of water can be transported into the deep mantle below Nicaragua and NW Costa Rica. However, the predicted variations in the thermal structure are insufficient to explain the large along-strike variations in seismicity and arc geochemistry, leading to the conclusion that these variations are due to regional variations in sediment subduction, crustal structure, and the distribution of hydrous minerals in the incoming lithosphere (e.g., Rüpke et al., 2004).

One of the remaining trouble spots for slab thermal modeling is the coupling of the subducting and overriding plates. There are two traditional approaches to model the decoupling. The first assumes that both the overriding plate and wedge have viscous rheology and decoupling occurs through a weak zone, or a slipping fault. The second approach defines the overriding plate as a rigid zone below which the subducting slab couples with the viscously defined mantle wedge. The depth of decoupling, type of weak-zone or slipping-fault parameters are not well constrained by observations and this can result in significant variations in the predicted temperatures in the wedge and subducting slab. Conder (2005) suggests an approach that incorporates new rheological criteria in which ambient temperature and strain rate define the brittle–ductile transition, which in turn defines the depth at which the boundary condition between slab and overriding plate changes from fully decoupled to fully coupled. The slab surface temperatures in Conder’s models are generally higher than observed in other models, and may provide an explanation for the geochemical signatures of sediment melting in some subduction zone.


7.08.3.3 Dynamic Models and Mantle Flow

There are many questions that cannot be addressed with subduction models with imposed slab velocities, chief among them the question of the driving forces of plate motions. Dynamic subduction models are generally an attempt to produce a subduction-like geometry within a mantle convection calculation. Some of the earliest calculations have downwellings along the edge of the domain (e.g., Christensen and Yuen, 1984) These calculations address the effect of phase transformations on subduction. In tank experiments with fluids stratified by density and frozen ‘plates’ atop the fluid, Kincaid and Olson (1984) produce slabs with morphologies similar to those in Christensen and Yuen (1984) (see also Chapter 7.03 for more on tank experiments).

Attempts to produce asymmetric downwellings in dynamic calculations include velocity boundary conditions (e.g., Stevenson and Turner, 1977; Davies, 1989; Christensen, 1996), a priori specified mechanically weak zones (e.g., Kopitzke, 1979; Gurnis and Hager, 1988; King and Hager, 1994; Chen and King, 1998), and/or dipping viscous faults (e.g., Zhong and Gurnis, 1992, 1994, 1995; Toth and Gurnis, 1998; Ita and King, 1998). King et al. (1992), Zhong et al. (1998), Gurnis et al. (1999), Bercovici et al. (2000), and Bercovici (2003) and references therein discuss and compare various plate-generation methods. The influence of the vertical sidewalls in 2-D is a significant limitation to subduction models because the 2-D geometry requires a strong ‘return flow’ which has a tendency for the slab to steepen with depth, and in many cases dipping more than 90° something not observed in subduction zones on Earth. Attempts to circumvent this problem 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).

It has been recognized that there are limitations to the planform of convection in 2-D and there was a hope that subduction-like geometries might be easier to achieve in 3-D calculations. Yet the model geometry alone is insufficient to produce subduction zone geometries; 3-D spherical geometries fail to produce the slab geometries observed in modern subduction zones (e.g., Bercovici et al., 1989; Tackley et al., 1993; 1994; Ratcliff et al., 1995, 1996; Bunge et al., 1996, 1997). In 3-D constant-viscosity calculations (e.g., Bercovici et al., 1989) the global planform of the flow could be described as linear, sheet-like downwellings that break off into isolated drips and cylindrical, plume-like upwellings. Although these features do not resemble plumes or slabs in any detail, the results were encouraging. When temperature-dependent rheology is included in 3-D spherical calculations, the planform changes, that is, cylindrical downwellings at the pole and sheet-like upwellings (Ratcliff et al., 1995, 1996) or stagnant-lid convection (Moresi and Solomatov, 1995; Solomatov and Moresi, 1996).

Weak zones and faults are difficult to implement numerically, require additional implementation challenges if the fault or weak zone move and/or evolve
with the flow, and implicitly assume some heterogeneity (either in strength or composition) pre-exists before subduction begins. This might be entirely realistic; the initiation of subduction may require some level of pre-existing heterogeneity (cf. Toth and Gurnis, 1998; Gurnis et al., 2000b). Some investigators have attempted to produce dynamic, mobile plates using complex constitutive equations (e.g., Bercovici, 1994, 1995, 1999; Tackley, 1998; Trompert and Hansen 1998). A constitutive equation is a relationship between two or more variables that is needed to close the equations of motion. Constitutive equations are generally not based on conservation laws, but are empirically derived. For mantle convection (i.e., creeping viscous flow), this is a relationship between stress and strain rate. The simplest relationship is a linear relationship between stress and strain rate; in this case, the constant of proportionality is the kinematic viscosity, which can be a constant or function of temperature, pressure, or composition (see Chapters 2.14 and 7.02). If the relationship between stress and strain rate is linear, this is referred to as a Newtonian viscosity. If the relationship between stress and strain rate is nonlinear, it is called a non-Newtonian or power-law fluid.

A series of investigations of convection with temperature-dependent Newtonian and/or non-Newtonian rheologies documented a transition from the free-slip planform found in constant-viscosity calculations through a sluggish-lid planform to a stagnant-lid planform of convection as the stiffness of the boundary layer increases (Christensen, 1985; Solomatov, 1995; Moresi and Solomatov, 1995; Solomatov and Moresi, 1996). In stagnant- and sluggish-lid convection, most of the top boundary layer is rigid and remains at the surface and only the weakest bottom 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 and is quite different from our picture of subduction, where most, if not all, of the subducting plate descends into the interior of the fluid. In the stagnant-lid planform, a significant amount of the negative buoyancy in the top thermal boundary layer (i.e., the plate) does not participate in the downwelling (Moresi and Solomatov, 1995; Conrad and Hager, 1999). This has important implications for the plate driving force due to subducted slabs and for the calculation of gravity anomalies over subduction zones because the thermal structure is related to the density structure through the coefficient of thermal expansion. Furthermore, 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, which has a significant effect on the cooling of the Earth through time (i.e., thermal history calculations, cf. Gurnis, 1989). The sluggish-lid and stagnant-lid convective planforms are inconsistent with piecewise-uniform surface velocities, and to date constitutive theory models have yet to produce realistic subduction zone geometries.

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 (van den Berg et al., 1991; Weinstein and Olson, 1992; King et al., 1992); 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. Rheologies based solely on the properties of mantle minerals require the imposition of a yield stress (Tackley, 1998, 2000a, 2000b) or damage theory (Bercovici et al., 2001; Bercovici, 2003; Bercovici and Richard, 2005) and have not been used for subduction modeling experiments. Some researchers argue that subduction forms at pre-existing zones of weakness in the lithosphere (e.g., Kemp and Stevenson, 1996; Schubert and Zhang, 1997; Toth and Gurnis, 1998; Regenauer-Lieb and Yuen, 2000; Regenauer-Lieb et al., 2001; Regenauer-Lieb and Yuen, 2003); hence, heterogeneous properties of the lithosphere are important for dynamic subduction zone models.

A criticism of the calculations discussed above is that they are limited, for the most part, to 2-D geometries, or 3-D geometries with reflecting sidewall boundary conditions. Symmetric, near-90° dipping downwellings are also the observed planform in 3-D 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 2-D eliminates the effect of the sidewalls and 2-D 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 (roll-back) has been one of the most studied mechanisms for producing dipping slabs
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 RD (1986) Relations among subduction parameters. Reviews of Geophysics 24: 217-284.

Figure 11

(compositional buoyancy leads to separation of the slab components. Furthermore, these calculations demonstrate that a significant amount of slab deformation is expected in the transition zone. This is explored in tank experiments by Guilloufrottier et al. (1995) showing a wide range of slab deformation styles (e.g., sinking slab, stagnant slab, spreading slab, sinking pile, and stagnant pile) when slight changes in the parameters are imposed. These deformation modes are governed by two velocity ratios, which characterize the horizontal and the vertical components of the slab velocity near the interface.

There are several cases where 3-D 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, and a significant amount of trench migration has been documented. It is possible to reconcile the apparent lack of correlation of the global trench migration observations with slab geometry and the success of the regional subduction studies by admitting that subduction is a time-dependent phenomenon and that 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). It is important to keep in mind that the geometries of present-day subduction zones provide a single snapshot of a time-dependent phenomenon and that 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). The difference between Chilean style subduction zones and Mariana style subduction zones has been recognized for some time (e.g., Uyeda and Kanamori, 1979). It is difficult to separate the effects of dip direction from the fact that most of the eastward-dipping slabs are being overridden by the North and South American Plates (i.e., continents) while many of the slabs in the Pacific are being overridden 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 2D are not practical to implement in a complete sphere in 3-D calculations. Furthermore, the grid resolution that is required to study subduction problems exceeds what can be practically achieved in a 3-D spherical shell at present.

The area where the asymmetry of subduction is most apparent is the shallow-dipping slabs, such as the subduction of the Pacific Plate beneath Alaska, the Juan de Fuca Plate under Washington and Oregon, and the Cocos and Nazca Plates under Central and South America. The subduction of young lithosphere (in most cases the slab is significantly younger than the slabs in the western Pacific) is not driven by the negative buoyancy of the slab (cf. Van Hunen et al., 2000), 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; Vaal, 1983; Van Hunen et al., 2000, 2001, 2002, 2004). 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. However, Van Hunen et al. (2000, 2001) find that weak crustal rheology, a ‘subduction fault’ between the two plates, nonlinear rheology, and viscous shear heating all play a role in weakening the shallow crust and mantle and enabling shallow subduction. The subduction of a buoyant oceanic plateau has been suggested as a possible cause for shallow subduction; however, this does not appear to be a significant factor in the case of Peru (Van Hunen et al., 2004).

**7.08.3.4 3-D Numerical Models of Regional Subduction**

Billen and Gurnis (2001) and Billen et al. (2003) examine 3-D, dynamic subduction calculations using a slab geometry appropriate for the Tonga–Kermadec and Aleutian slabs. By varying the slab and wedge rheology as well as the slab density structure, they are able to match the observed geoid, topography, and the state of stress in the slab. There are several important restrictions to note in the details of their model development: (1) these are instantaneous snapshots of the flow field (i.e., the slab morphology, plate/slab velocities, and slab thermal structure do not evolve with the flow field); (2) the element size ranges from 2.5 km within the wedge to 100 km at the size boundaries (compare this with the results from the 2-D slab thermal models discussed above); and (3) the thermal structure of the slab (i.e., the initial density structure) is derived from a kinematic thermal structure model. Nonetheless, this work represents an important and significant step forward in understanding subduction zone dynamics.

Although the work of Billen and colleagues (Billen and Gurnis, 2001; Billen et al., 2003) does not focus on slab thermal structure, one of their results motivates the need for more detailed studies of this kind. They find that they need to increase the buoyancy in their slabs by 30% more than predicted by a reasonable equation-of-state model in order to drive the plate and slab at the appropriate observed velocity and match the geoid and topographic profiles. It is reasonable to ask whether the problem is the result of buoyancy estimate from the equation of state, the dissipation of energy in the ‘trench region’ due to their fault and/or rheology (cf. Conrad and Hager, 2001), or whether the slab thermal structure which they use as the initial condition for their model is wrong. Billen and Gurnis choose to vary the buoyancy; however, previous work has shown that there is a tradeoff between slab buoyancy and slab strength. For example, Griffiths et al. (1995) and Houseman and Gubbins (1997) have shown that the deformation of a slab depends on the slab stiffness and slab buoyancy. Griffiths et al. define a slab stiffness (the viscosity of the slab divided by the viscosity of the mantle) and find a relationship between the slab deformation and slab stiffness. There are additional observational constraints that can be used to address this question. For example, Bevis (1986, 1988) used slab geometry and earthquake sources to estimate slab strength. The distribution of earthquakes in Benioff zones (Oliver and Isacks, 1967; Lundgren and Giardini, 1992), regional tomography (van der Hilst, 1995), and scattering experiments all constrain the morphology of the slab in the upper mantle.

**7.08.3.5 Phase Transformations: Dynamics**

The eventual fate of deep slabs has been a topic of debate for several decades (see Lay, 1994; Christensen, 1995; see Chapter 7.10). One of the original arguments for a barrier to convection at 670 km depth was the cessation of earthquakes at this depth (see Chapter 4.11). The amount of slab penetration into the lower mantle controls the rate of heat and mass transfer.
between the upper and the lower mantle (e.g., Lay, 1994; Silver et al., 1988) and understanding slab penetration is critical to understanding thermal convection and chemical mixing of the mantle (e.g., Albarède and van der Hilst, 1999; Kellogg et al., 1999). Proposed hindrances to slab penetration into the lower mantle include the phase transformation of wadsleyite to perovskite plus magnesiowustite (Schubert and Turcotte, 1971; Schubert et al., 1975; Ringwood and Irifune, 1988; see Chapter 2.06), a change in chemical composition (e.g., Richter, 1977; Anderson, 1979), and an increase in viscosity (e.g., Hager, 1984). A compositional barrier to convection at around 660 km is not considered likely by most and the topic of viscosity will be covered in a later section. Here we focus on the effect of phase transformations.

A phase change, in and of itself, will not necessarily impact convective flow. Because convection is driven by horizontal density gradients, the idealized phase transformation that happens at one pressure, \( P \), for all temperatures will, to first order, have no impact on the flow. There is an effect due to the latent heat release due to the phase change and the small effect of dynamic pressure, which varies horizontally, but is generally on the order of 1% of the hydrostatic pressure. Phase transformations primarily affect convective flow in the mantle because of the temperature variation of the pressure of the phase transformation. This is expressed by the Clausius–Clapeyron relationship,

\[
\frac{\partial \rho}{\partial T} = \gamma = -\frac{\Delta S}{\Delta V} = \frac{\Delta S \rho^2}{\Delta \rho} = \frac{Q_h \rho^2}{\Delta \rho T} \tag{18}
\]

where \( T \) is the absolute temperature, \( \Delta S \) is the entropy change, \( \Delta V \) is the specific volume change, \( Q_h \) is the latent heat, and \( \rho \) is the mean density of the two phases (see Chapter 7.02). A positive Clapeyron slope, such as the transformation of olivine to wadsleyite, means the denser phase will occur at shallower depth in the slab, with the effect of increasing the negative buoyancy in the slab, enhancing convection. A negative Clapeyron slope, such as the transformation of spinel (ringwoodite) to perovskite plus ferropericlase, means the lighter phase will be present to greater depths in the cold slab, as compared to the surrounding mantle, decreasing the net negative buoyancy of the slab and retarding convective motion (see Figure 12).

A simple comparison of densities is sufficient to illustrate the importance of phase transformations on buoyancy. The coefficient of thermal expansion for mantle minerals is on the order of \( 2.0 \times 10^{-5} \text{ K}^{-1} \). If we assume a 400 km column of slab material, \( D \), that is 500°C colder than its surrounding material, \( \Delta T \), this slab will have an integrated buoyancy anomaly, \( \rho g \Delta T D \), of 132 N m\(^{-2}\). Comparing this with a phase transformation with a change in density of 6% that has a topography on the phase boundary of 30 km, the phase-change topography will have a buoyancy anomaly of 0.06 \( \times \rho g b_{pc} \) of 59.4 N m\(^{-2}\). Thus, a 30 km deflection of a phase boundary can have a comparable buoyancy anomaly to a 400 km column of 500°C anomalously cold slab.

It is important to recognize that the mantle is not a single-component system. Because of the nature of the transformations from olivine to wadsleyite and ringwoodite (spinel) to perovskite plus ferropericlase and the close association of the pressure of these phase transformations to the 410- and 660-km seismic discontinuities, most studies of convection with phase transformations have included only the olivine system in their calculations. For a simplified phase diagram see Figure 13. It is generally assumed, because the pyroxene-to–garnet transitions are spread out over a larger pressure range, that the effect of these phase transitions will be small and the overall effect of the combine pyrolite system will be that the dynamic effect of the olivine-to-wadsleyite and ringwoodite-to-perovskite plus ferropericlase phase changes will be weaker than in a pure olivine system, because the pyroxene/garnet component makes up about 40% of the composition.

Because deep earthquakes are restricted to subduction zones (see Chapter 4.11), the mechanisms of deep earthquakes are assumed to be related to the
This has led to a variety of hypotheses on what hinders the cold Tonga slab, including subhorizontal deflection of the slab (e.g., van der Hilst et al., 1995) or the collection of slab material at the base of the transition zone to form a large megolith (Ringwood and Irifune, 1988; Irifune, 1993). The complexity of the Tonga subduction zone and the absence of a deep fast anomaly beneath Tonga point to some form of hindrance to slabs at the base of the transition zone.

It is important to recognize that while the over-riding assumption made in almost all subduction zone studies is that buoyancy within a slab is dominated by the effect of temperature, phase transformations have significant effects locally (e.g., Daessler and Yuen, 1996; Bina 1996; 1997; Christensen, 1996b, 1985; 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 descent 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. In addition, the pattern of stresses induced by the differential buoyancy are consistent with the pattern of deep earthquakes (Bina, 1997).

Early attempts to model subduction zones using dynamic mantle convection models 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; see also Chapter 7.02). The tank experiments induce an asymmetric 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 is introduced with a shallow-dipping bend at its leading edge; this provides the instability that allows this plate to subduct under the other plate. The agreement between the tank and numerical experiments (Figure 14) suggests that the three

![Figure 13](image-url)
The planforms discussed above are robust features of cold downwellings interacting with phase transformations and/or chemical boundaries. Subsequent numerical (Zhong and Gurnis, 1994; 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.

There is a general consensus that has emerged from both 2-D and 3-D calculations in both Cartesian and spherical geometries. The evolution of the flow can be described by three distinct stages: (1) the initial impingement of the leading edge of the subducting slab on the spinel to perovskite plus ferropericlase phase boundary; (2) a period of increased trench rollback and draping of the slab on the phase boundary; and (3) virtual cessation of trench rollback as the slab penetrates into the lower mantle.

Studies have clearly demonstrated that the effect of a phase transformation on the pattern of flow in a convection fluid is sensitive to the Rayleigh number, initial conditions, boundary conditions, and the equation-of-state approximation (e.g., Ita and King, 1994, 1998; King et al., 1997). For example, Ita and King (1998) show that a depth-dependent coefficient of thermal expansion approximation develops a stagnant slab and layered convection while the slab penetrates into the lower mantle with a calculation with a self-consistent equation of state based on Ita and Stixrude (1992) and otherwise identical parameters. King et al. (1997) show that calculations including an exothermic olivine-to-spinel phase transformation decreases layering. Ita and King (1994) show that at Rayleigh numbers of order $10^5$, the initial flow pattern imposed on the system can persist for 30–40 transit times or more and reflecting sidewall boundary conditions used in a Cartesian, box-like system tend to produce layered flows more easily than those with flow-through conditions. Ita and King (1994) go on to show a decrease in layering as the thermodynamic consistency is increased (i.e., Boussinesq calculations are more likely to be layered than compressible calculations with a self-consistent equation of state). The effect of latent heat on the topography of phase boundaries can be important (cf. Christensen, 1998) and is not well accounted for in most convection calculations.

7.08.3.6 Shear Heating

There is a large degree of uncertainty regarding the relative importance of shear heating in the thermal structure of a subduction zone, in large part because of a disagreement on the magnitude of the subduction zone shear stress with some estimates ranging from 100 MPa (e.g., Scholz, 1990; Molnar and England, 1990) and others an order of magnitude lower (e.g., Peacock, 1992; Tichelaar and Ruff, 1993; Hyndman and Wang, 1993). The rate of shear heating is given by the product of the shear stress, $\tau$, and the convergence velocity, $v$ (see Chapter 7.06),

$$Q_{sh} = \tau v$$  \[19\]

While the convergence rate is well constrained (e.g., Jarrard, 1986), the magnitude of the shear stress in subduction zones is estimated based on surface heat flow measurements and/or petrologic arguments and this quantity is at best loosely constrained. In a fully compressible convection formulation, the shear stress can be calculated consistently in the viscous flow formulation (e.g., Yuen et al., 1978). However, even
in such a formulation there are uncertainties because the deformation mechanism within the slab is poorly understood and perhaps more importantly, because of the uncertainty and difficulty in representing the deformation between the slab and overriding plate. The fore-arc heat flow places a strong constraint on the amount of shear heating that can be added in the upper 50 km. Early kinematic subduction zone thermal models required a high shear stress in order to match the observed heat flow (e.g., Turcotte and Schubert, 1973). With temperature-dependent rheology, the wedge corner flow brings more hot material close to the surface (and the slab) and a lower rate of shear heating is needed (e.g., van Keken et al., 2002). The uncertainty in the amount of shear heating, particularly below 50 km depth, leads to several hundred degrees difference in the temperature of the slab (Peacock, 1996).

7.08.3.7 Elastic/Plastic Subduction Calculations

The discussion thus far has focused on calculations with either imposed velocities or viscous flow; however, there is another important group of calculations that include the effect of elastic strength and brittle faults (e.g., Kemp and Stevenson, 1996; Schubert and Zhang, 1997; Toth and Gurnis, 1998; Gurnis et al., 2000a, 2004; Regenauer-Lieb and Yuen, 2000; Regenauer-Lieb et al., 2001; Regenauer-Lieb, 2003; Hall et al. 2003). The work of Vassiliou et al. (1984) showed that viscous flow was able to reproduce the patterns of stress attributed to elastic bending of the plate as it entered the trench; hence, viscous flow has been thought to capture the broad pattern of deformation at a subduction zone. While this is probably true, the recognition that the deformation in the seismogenic zone (the upper 50 km or so of the boundary between the subducting and overriding plate) impacts the deeper slab structure by altering the slab thermal structure, more work in this area is needed. A useful overview of the role of pre-existing faults and subduction can be found in Gurnis et al. (2000a).

7.08.4 Petrology, Geochemistry, and Arc Volcanics

Perhaps one of the most interesting subduction zone puzzles is that subduction zones are where cold surface material returns to the mantle, yet island arcs, a significant source of volcanic activity, are associated with subduction zones. Subduction zones are cool, relative to the average upper-mantle temperature, so why is there any volcanic activity associated with subduction zones? The early postplate tectonic view is that island arc volcanism was the result of slab melting (e.g., Ringwood, 1975) which required temperatures approaching the melting point of basalt. The only way to achieve such high temperatures in a subduction environment was with large rates of frictional heating (e.g., Oxburgh and Turcotte, 1968). However, other constraints on the rate of frictional heating (cf. Peacock, 1996), such as the fore-arc heat flow, produce slabs that are significantly cooler than necessary for basalt melting (e.g., Peacock, 1991; Davies and Stevenson, 1992).

There are several ways to resolve this apparent inconsistency: volatiles, mainly water from the oceanic sediments and crust, reduce the melting point of mantle minerals leading to melting at lower temperatures; low melting point sediments could be the major source of arc volcanics; frictional shear heating could significantly increase the temperature near the slab–wedge interface; and induced flow from the down-going slab could advect additional heat into the mantle wedge, warming the wedge relative to average upper mantle. Each of these mechanisms will be examined below and in reality some combination of these factors is likely to be important.

While most of the subducting slab is not melted and may eventually wind up at the core–mantle boundary, subduction zones are the primary source of continental crust formation and the major mechanism for recycling volatiles into the mantle. The subducting slab is comprised of layers of compositionally distinct components: sediments, oceanic crust, and oceanic lithosphere. Often these components are considered as a single entity, the ‘slab’, and most convection models do not have the resolution to track the individual components separately. Given that oceanic lithosphere makes up more than 90% of the mass (volume) of the subducting slab, it is not surprising that this controls the mechanics of subduction. The mass (volume) of oceanic crust and sediment will be insufficient to play the major role in the buoyancy or strength of the subducting slab. Possible examples of where oceanic crust may play a significant role include the subduction of oceanic plateaus. Another caveat is that sediments, crust, and/or volatiles coming off the slab may play an influential role in the mechanics of the seismogenic zone. Even though the mechanics of subduction is
dominated by the oceanic lithosphere, it is important to keep in mind that the crust and sediments play important geochemical roles in subduction zone processes.

The emerging view is that the geochemistry points to both volatile-induced melting and sediment melting (Elliott, 2003). There are extensive reviews by Elliott (2003) and Gaetani and Grove (2003) that cover work that is beyond the scope of this review. While there is a vast amount of work on arc petrology and geochemistry, much of this work has only begun to be integrated in subduction zone thermal modeling (e.g., van Keken et al., 2002; Kelemen et al., 2003; Peacock, 2003). There are several reasons for this, the primary is that until the ‘rediscovery’ of Furukawa’s (1993) work on the impact of temperature-dependent rheology on flow (and hence temperature) in the slab corner, there was an inconsistency in slab thermal models – the only way to achieve the temperatures needed for sediment melting and significant dehydration was a high rate of shear heating. However, the high shear stress required was inconsistent with fore-arc heat flow (cf. Peacock, 1996). Another reason that the geochemical and petrological constraints have not been more closely linked with subduction zone models is that the scale of subduction zone models, particularly dynamic models, has been too coarse to study the region of interest. This is changing with the emergence of the new hybrid kinematic–dynamic models (e.g., van Keken et al., 2002; Kelemen et al., 2003; Peacock et al., 2005).

7.08.4.1 Subduction and Volatiles

Oceanic crust contains a large fraction of the water carried into subduction zones (Wallmann, 2001; see Chapter 2.04). Mid-ocean ridge basalt (MORB) is depleted in incompatible trace elements and contains 50–200 ppm water (Dixon et al., 2002). Therefore, water, CO₂, and incompatible trace elements in the crust are the result of hydrothermal alteration and seafloor weathering. Staudigel et al. (1996) infer that most of the water, CO₂, and incompatible trace elements are concentrated in the upper 500 m of basalts. Formation of amphibolites in the lower crust stores water but little else (Carlson, 2001). Serpentine may be the most important mineral for water transport because it is stable to 7 GPa and can carry an order of magnitude more water than hydrated crust (Pawley and Holloway, 1993).

The role of water in melting at arcs has been studied extensively, dating back to Green and Ringwood (1967) and recent reviews have covered the topic in a depth not allowed in this chapter (Poli and Schmidt, 2002; Eiler, 2003; Elliott, 2003; Hirschmann, 2006). Water plays several important roles in arc geochemistry and dynamics by reducing the melting temperature (Hirth and Kohlstedt, 1996; Asimow et al., 2004) and impacting the rheology (Hirth and Kohlstedt, 2003).

The subducted slab plays an important role in the water cycle, connecting the surface to deep mantle. The extent to which water (and other volatiles) may recycle back to the deep mantle is an intriguing problem. Peacock (2001) suggests that the lower line of earthquakes in double Benioff zones may be related to dehydration. The water-storage capacity for olivine increases significantly with depth (Kohlstedt et al., 1996) and the transition zone phases of wadsleyite and ringwoodite have large water-storage capacities (Kohlstedt et al., 1996; Smyth et al., 1997). It is possible that if water can get to the transition zone, the transition zone could contain as much water as is present on the surface (Smyth, 1994; Bercovici and Karato, 2003). While subducted oceanic crust is thought to be dehydrated before reaching the transition zone (e.g., Schmidt and Poli, 1998), peridotite could carry water into the transition zone, particularly by water that has percolated to depth in the slab along extensional faults formed at the outer rise (e.g., Peacock, 2001; Rüpe et al., 2004). It is possible that hydrated phases may only be stable to the transition zone along unusually cold adiabats (e.g., Staudigel and King, 1992), requiring old ocean lithosphere or very high convergence rates, or both.

Arcay et al. (2005) model subduction zone thermal structure with a dynamic mantle convection code and include dehydration reactions and a rheology that depends on pressure, temperature, strain rate (i.e., non-Newtonian), and water content. Consistent with other recent studies (e.g., van Keken et al., 2002; Conder et al., 2002, Kelemen et al., 2003) they find that non-Newtonian rheology leads to a hotter slab surface temperature and greater erosion of the overlying plate than in isoviscous corner flow models. At high convergence rates, serpentinite can be transformed into hydrated phase A, leading to recycling of water to significant depth. Even in cases where a significant amount of water is carried to depth, the mantle wedge is significantly hydrated within 250 km of the trench, effectively broadening the low-viscosity zone near the tip of the slab wedge. The hydrated weak wedge region may contribute to back-arc deformation.
7.08.4.2 Melting of Sedimentary Material

The most compelling evidence for a ‘slab’ component in arc magmas comes from the trace elements, B, Be, Th, and Pb (e.g., Morris et al., 1990; Hawkesworth, et al., 1993; Plank and Langmuir, 1993; Ishikawa and Nakamura, 1994; Smith et al., 1995; Elliott et al., 1997; Wunder et al., 2005; George et al., 2005; Plank, 2005). The presence of $^{10}$Be, an isotope of Be only produced in the upper atmosphere with a half-life of 1.6 My, is strong evidence that some oceanic sediments are recycled within subduction zones, travel through the mantle wedge, and wind up in the source region of arc lavas (Morris et al., 1990). Furthermore, the short half-life places a tight constraint on the transport of material through the subduction/arc system.

The volume and composition of sediment subducted into various trenches varies considerably (e.g., von Huene and Scholl, 1991). This might represent an along-trench variation apart from slab geometry (dip and convergence velocity) that could explain along-arc variations in subduction zones. There is pretty clear evidence now that the geochemistry identified two distinct components (e.g., Elliott, 2003): a melting of sedimentary material component and an aqueous fluid-derived component.

7.08.4.3 Phase Transformation in Slabs

Phase transformations within slabs can have a significant effect on subduction dynamics (e.g., Daessler and Yuen, 1993, 1996; Bina, 1996, 1997; Christensen, 1996b, 1998; 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-wadsleyite phase transformation is kinetically hindered in cold slabs (see Kirby et al. (1996) and references therein). The presence of metastable olivine may reduce slab descent 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).

Marton et al. (2005) use a thermokinetic model of subduction zone thermal structure with a thermal conductivity that depends on pressure, temperature, and mineralogy. These calculations do not consider variable rheology. The thermal conductivity decreases exponentially with increasing temperature and the effect of pressure is much smaller, especially over the depth range of the upper mantle. Considering the effect of radiative heat transfer, they conclude that thermal conductivity of the slab could increase by more than 50%, leading to a 50–100° increase in the temperature of the interior of the slab compared to that in a standard model with uniform thermal conductivity, independent of the slab geometry. Marton et al. (2005) use these models to address the question of olivine metastability and the ability of a metastable wedge to explain the pattern of deep seismicity. They conclude that the deepest earthquakes occur in regions that have already transformed to wadsleyite or ringwoodite, even when taking into account all of the uncertainties in their parametrization. In the case of Tonga the deepest earthquakes extend 140 km beyond the olivine metastability region.

7.08.5 Geoid and Topography

The association between local maxima in the long-wavelength component of the Earth’s gravitational potential field (geoid) and subduction zones has been recognized by many authors (Runcorn, 1967; Kaula, 1972; McKenzie, 1977; Chase, 1979; Crough and Jurdy, 1980; Davies, 1981; Hager, 1984; see Chapter 3.02). The geoid is the surface of constant potential (i.e., gravitational energy plus centrifugal potential energy) that coincides (almost) with mean sea level over the oceans. (Mean sea level is not quite a surface of constant potential, due to dynamic processes within the ocean; however, for illustrative purposes it will suffice.) The geoid is actually the height of the potential surface referenced to an oblate spheroid resulting from a hydrostatic spinning Earth (Nakiboglu, 1982).

In a static mantle, with no dynamic flow, the geoid over a positive mass anomaly is positive (Figure 15, left panel). In an isoviscous fluid (Figure 15, middle panel), the gravitational potential (or geoid) over a positive mass anomaly, such as a subducting slab, is negative (Richards and Hager, 1984). This is because the flow driven by the mass anomaly (blue in Figure 15) deforms the surface (and core mantle boundary) creating the red, negative mass anomalies (i.e., long-wavelength depressions). The sum of the contributions from the positive and negative mass anomalies in the isoviscous fluid is negative (and small) (cf. Richards and Hager, 1984; Ricard et al., 1984). With a layered viscosity increase with depth (Figure 15, right panel), the amount of surface deformation from the same positive mass anomaly is
reduced (i.e., the mass anomaly is supported by the stiff layer below) and the resulting anomaly can change sign depending on the viscosity structure (cf. Richards and Hager, 1984; Ricard et al., 1984; Chapters 7.02 and 7.04).

The long-wavelength geoid associated with subduction zones requires that subducting slabs encounter a resistance to flow, modeled as an increase in effective viscosity by a factor of 30, or more, at a depth of 670 km (Hager, 1984; Hager and Richards, 1989; Ricard et al., 1989; Hager and Clayton, 1989; Zhong and Gurnis, 1992; King and Hager, 1994). Because the equation for the gravitational potential is linear, we can decompose the problem; the total geoid anomaly is the sum of the contribution from each density anomaly within the Earth. This includes density anomalies that drive mantle convection, such as dense slabs or buoyant plumes, and density anomalies that result from deformed boundaries as a result of convection. This problem was first addressed by Pekeris (1935) but has been expanded on by others (e.g., Runcorn, 1967; Morgan, 1965; McKenzie, 1977; Richards and Hager, 1984; Ricard et al., 1984). Focusing on subduction zones for purposes of illustration, the mass excess due to the dense, sinking slab contributes a positive term to the total geoid while the mass deficit, due to the down-warping of the surface above the slab, contributes a negative term to the total geoid. These contributions are similar in magnitude and opposite in sign. The resulting total geoid is sensitive to variations in either the density structure of the slab or the deformation of the surface. The magnitude of the down-warping of the surface is a strong function of the viscosity structure of the medium (cf. Richards and Hager, 1984; Ricard et al., 1984). (It should be noted that the boundary between the mantle and core is also deformed as a result of the flow driven by the sinking slab; however, at the wavelengths appropriate to subduction zone problems, this is not a significant contribution even though it is included in most geoid calculations (Richards and Hager, 1984).)

While the density structure of a subducting slab is a complex function of temperature, chemical variations, and phase transformations, there is little doubt that subducting slabs are more dense than surrounding mantle. In order to create a positive geoid anomaly over a subducting slab in an isoviscous fluid, it would require that estimates of slab thermal structure underestimate the density anomaly due to the slab by approximately a factor of 4 (Davies, 1981) or that approximately 300 km of slab material is piled in the transition zone (Hager, 1984). This is much greater than the combined uncertainties in the thermal structure, thermal properties of the slab, and the errors made by neglecting the compositional

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**Figure 15** Schematic representation of geoid anomalies over a positive mass anomaly for a static (left), isoviscous (center) and layered viscosity (right) mantle. The blue line is the component of the anomaly due to the internal mass, the red line is the component of the anomaly due to the surface deformation, and the black line is the sum of the components.
component of the slab. Thus, some form of resistance to the downward motion of the slab is necessary to reduce the associated surface deformation. In the majority of studies on subduction, it has been assumed that the resistance to the slab is provided by an increase in viscosity at some depth (usually taken to be between the upper and lower mantle) and King (2002) has shown that the phase transformation from ringwoodite to perovskite plus ferropericlase is insufficient to change the sign of the geoid. With a sufficient reduction in the surface deformation, provided by the increase in viscosity of the lower mantle, the total geoid anomaly over the downwelling slab can be positive. In this way, the association of local maxima in the geoid with subduction zones provides a constraint on the viscosity of the mantle.

Based on the analysis described above, an increase in mantle viscosity with depth by a factor of 30–100 is required to produce local geoid maxima over subducting slabs. The low end of this range is consistent with studies of other geophysical observations that constrain mantle rheology including postglacial rebound (cf. Mitrovica, 1996; Mitrovica and Forte, 1997; Forte and Mitrovica, 2001), plate velocities (cf. Lithgow-Bertelloni and Richards, 1995) and rotation. The high end of the viscosity increase required by the subduction zone geoid observation is more difficult to reconcile with most mantle viscosity models from postglacial rebound studies; although a few postglacial rebound studies have predicted a viscosity increase this large (e.g., Lambeck et al., 1990). It is also possible that both observations are compatible because the depth to which the postglacial rebound observation and the subduction zone observation are sensitive are different. For example, some models of mantle viscosity predict an increasing viscosity with depth throughout much of the lower mantle (e.g., Ricard and Wuming, 1991). If these viscosity models are correct, then it is possible to reconcile a higher value of lower-mantle viscosity predicted by the subduction zone observation because the depth to which the subducting slab is sensitive is greater than the depth to which the postglacial rebound observation is sensitive.

The power spectrum of the geoid is dominated by the longest wavelengths and the power decreases at approximately the fourth power of the spherical harmonic degree (Kaula, 1968). This observation has become known as Kaula’s rule. As a result, the analysis of all but the longest components of the geoid requires spatial filtering. Figures 16 and 17 compare the global geoid from the NIMA/NASA EGS96 geopotential model (Lemoine et al., 1997) that has

Figure 16 Global geoid from EGM96 (Lemoine et al., 1997) band-pass-filtered passing wavelengths smaller than 15,000 km and cutting wavelengths greater than 5000 km. These parameters are comparable to the degree 4–9 ‘slab’ geoid used by Hager (1984). Routines in the GMT software package were used to filter the data (Wessel and Smith, 1991). Reproduced from King SD (2002) Geoid and topography over subduction zones: The effect of phase transformations. Journal of Geophysical Research 107 (doi:10.1029/2000JB000141).
been band-pass filtered using two bands. The first image (Figure 16) cuts wavelengths greater than 15,000 km and smaller than 4000 km and passes wavelengths between 10,000 and 5000 km. The second image (Figure 17) cuts wavelengths greater than 5000 km and passes wavelengths smaller than 4000 km. Figure 16 is comparable to the spherical harmonic degree 4–9 geoid that has been shown to correlate well with subduction zones (Hager, 1984) and the correlation between geoid highs and subduction zones is clearly evident in with the Aleutian arc being a notable exception. Short-wavelength geoid anomalies, like those presented in Figure 17, have been presented in trench-perpendicular profiles (e.g., Davies, 1981; Zhong and Gurnis, 1992; King and Hager, 1994; Billen et al., 2003) and the geoid and topography at this wavelength has been used to constrain dynamic subduction zone models (Zhong and Gurnis, 1992, 1994, 1995; King and Hager, 1994; Chen and King, 1998; King, 2002). Richards and Engebretson (2002) point out that the long wavelength geoid correlates with the general pattern of fast seismic velocities in the Pacific.

Focusing on short-wavelength geoid (Figure 17), there is significant structure in the geoid that appears to be related to subduction zones but is obscured by the longer-wavelength components. Specifically, there is a narrow trough in the geoid (a local geoid minimum) that is that approximately 200 km in width and spatially coincident with the trench (Figure 17). There are local geoid maxima in the back-arc region of almost every subduction zone in the short-wavelength geoid, and the geometry of the geoid maxima matches the geometry of the associated arcs. At this scale, there are geoid local maxima associated with subduction zones even at locations where the association fails at the longer wavelengths (e.g., the Aleutian, Kamchatkan, and Middle-America subduction zones.)

### 7.08.6 Discussion

Having reviewed the major observations, we turn to several outstanding questions that remain unresolved in subduction zone dynamics.

#### 7.08.6.1 Are Slabs Strong or Weak?

While convection models can reproduce many subduction zone observations with both strong and weak slabs (Tao and O’Connell, 1993), 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 throughout 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 $2.5 \times 10^{22}$ Pa s, no more than 200 times the upper-mantle viscosity and maybe only a factor of 2 greater than the lower-mantle viscosity (cf. King, 1995).

The regional gravitational potential, or geoid, high over subduction zones has an important constraint on mantle rheology (Hager, 1984). 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 (Richards and Hager, 1989; Ritzert and Jacoby, 1992; Zhong and Gurnis, 1992; King and Hager, 1994; Moresi and Gurnis, 1996; Chen and King, 1998; 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. King (2002) shows that dynamic support of the slab from an endothermic phase transformation is not enough to explain the geoid highs over subduction zones.

Creager et al. (1995) show that the concave oceanward bend in the Bolivian Orocline (South American slab) forces along-strike compression in the subducting slab. Their calculations of membrane strain rates in an assumed continuous slab demonstrate that this can be accommodated by either 10% along-strike compressive strain or geometric buckling. They present evidence for complex deformation in the deep slab that could contribute to the conditions for very large earthquakes by dramatically increasing deformation rate, anomalous advective thickening of a proposed overdriven olivine wedge. This provides a mechanism for local thickening of the olivine wedge and allows transformation faulting to occur over a much larger volume than for normal subduction.

As discussed in the slab structure section, the rate of accumulation of seismic moment in WBZs can be used to estimate the average down-dip strain rate in subducting slabs (Bevis, 1986; Fischer and Jordan, 1991; Holt, 1995), assuming 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. 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, 1978), leading to the rather surprising suggestion that the difference in deformation rates between the asthenosphere and subducting slabs is no more than a factor of 3. Taking the average descent rate of slabs this deformation 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.

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 recrystallization 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 remember that assumptions regarding slab viscosity are built into some slab thermal models. For example, the kinematic velocity field used to advect the temperatures is uniform except in the corner. 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 comparisons between kinematic models, which do not include slab deformation and dynamic models, which can include slab deformation, it is impossible to make a quantitative statement.

A consequence of the kinematic slab approximation is that there is no guarantee that the resulting
thermal structure models are dynamically consistent. The results of Billen and Gurnis (2001) illustrate the problem, even though high-resolution thermal slab structure was not the focus of their work. In order to produce the plate/slab velocities that give rise to the initial slab thermal structure, they have to modify the buoyancy in their slab in a way that is inconsistent with their assumed equation of state. (This is not meant to take away from their work, which is an important step forward; rather, their study illustrates the potential that is available to include more realistic equations of state and attempts to model realistic subduction geometries.) This has rather significant implications for studies that have used kinematic models to map out the mineralogy of the slab (e.g., Kirby et al., 1991, 1996; Ita and Stixrude, 1992; Bina, 1996; Marton et al., 1999). These mineralogical models produce significant differences in slab density. Changes in slab density lead to significant differences in the plate and slab velocities (e.g., Christensen, 1996b; Ita and King, 1998); however, the change in slab density is not accounted for in the flow field because buoyancy forces do not drive slab flow in the kinematic slab approximation. While kinematic models have provided an estimate of where the transitions in mineralogy might occur, they do not have the predictive ability to follow the implications of the mineralogical models as the slabs evolve dynamically (e.g., Ita and King, 1998). For example, we would like to be able to calculate whether metastable olivine and/or an increase in rheology would have an observable impact on subduction velocity, dip and slab deformation. Similar studies have been carried out in dynamic models (e.g., Gurnis and Hager, 1988; Zhong and Gurnis, 1992, 1997; King and Hager, 1994; Christensen, 1996b; Kincaid and Sacks, 1997; Ita and King, 1998; Chen and King, 1998). However, none of those calculations were able to resolve slab thermal structure with the resolution of the kinematic models. It is now possible to use dynamic models with the kind of resolution needed for thermal structure calculations (e.g., van Keken et al., 2002; van Keken, 2003).

### 7.08.6.2 Are Deep Earthquakes the Result of a Metastable Transformation of Olivine to Wadsleyite?

Transformational faulting of metastable olivine into wadsleyite or ringwoodite arose as an intriguing possible mechanism for the source of deep earthquakes, following the Green and Kirby laboratory experiments (cf. Green and Burnley, 1989; Kirby et al., 1991, 1996; Stein and Rubie, 1999). However, the rupture areas of the two big, deep 1994 events in Bolivia and Fiji seem to be quite large, based on finite fault modeling and aftershock studies (Wiens et al., 1994; Silver et al., 1995) while thermokinetic mineralogical models of subducting slabs (Daessler and Yuen, 1993; 1996; Daessler et al., 1996; Devaux et al., 2000) imply relatively thin ‘wedges’ of olivine at the depths of the two large deep earthquakes. The combination of the rupture paths and the thermokinetic modeling appears to rule out transformational faulting as a mechanism for deep earthquakes (cf. Wiens, 2001; Chapter 4.11). However it is important to note that the work of Daessler and colleagues has been performed with kinematic slab models and the Tonga–Fiji region is a complex zone with multiple interpretations (e.g., Gurnis et al., 2000a; Chen and Brudzinski, 2001, 2003; Burdzinski and Chen, 2003a, 2003b), exactly the kind or region where a kinematic slab approximation is most suspect.

The seismic observations supporting metastable phase transformations are weak at best. In addition to the large rupture zones of the deep earthquakes of 1994 and 1995 (Wiens et al., 1994; Silver et al., 1995), Collier et al. (2001) find no evidence for nonequilibrium phase transformations based on the elevation of the 410 km discontinuity at subduction zones, while Chen and Brudzinski (2003) find a pattern of mantle anisotropy beneath Tonga–Fiji that could be consistent with a metastable olivine wedge. Sandvol and Ni (1997) combined shear-wave splitting parameters of local and teleseismic S waves from intermediate and deep earthquakes in the southern Kurile and Japan subduction zones with splitting parameters obtained from SKS and SKKS waves to determine depth variation in shear-wave splitting both above and below the earthquake. The shear-wave splitting lag times indicate significant variation in azimuthal anisotropy below 350 km depth. This inference is consistent with the source-side splitting of 0.08 s lag time observed from deep teleseismic S waves that traverse the upper-mantle/lower-mantle boundary. Metastable olivine in the flattened and broadened southern Japan slab would be consistent with these observations; although, the presence of an anisotropic layer composed primarily of highly anisotropic γ-spinel at the base of the 410 km discontinuity would also be consistent with the splitting observations. Koper et al. (1998) present the results of a 3-month deployment of a 1000 km seismic line of 23 ocean-bottom seismometers (OBSs) and island broadband

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7.08.6.3 How Much Water Is Carried into the Transition Zone?

Subduction zones play a key role in the recycling of water into the mantle. By simple mass balance, oceanic crust contains a large fraction of the water carried into subduction zones (Wallmann, 2001; see Chapter 2.04). Olivine is an important host of hydrogen in the Earth’s upper mantle, and the OH⁻ abundance in olivine determines many important physical properties of the planet’s interior. The water-storage capacity for olivine increases significantly with depth (Kohlstedt et al., 1996) and the transition zone phases of wadsleyite and ringwoodite have large water-storage capacities (Kohlstedt et al., 1996; Smyth et al., 1997). Natural and experimentally hydrated olivines have been typically analyzed using uncalibrated spectroscopic methods with large uncertainties in accuracy. Bell et al. (2003) determined the hydrogen content of natural olivines using a new technique and indicate that an upward revision of some previous determinations by factors of between 2 and 4 is necessary. This could impact the calculations that have been performed to date. It is possible that if water can be carried into the transition zone, the transition zone could contain as much water as is present on the surface (Smyth, 1994; Bercovici and Karato, 2003). The question remains, “how much water can be carried into the transition zone?” Equally interesting, this raises the question, “how much water can leave through the bottom of the transition zone?” (see for instance, Karato et al., 2006).

There are several studies that address this question; however, the results are not yet clear. Davies and Stevenson (1992) suggest that a significant amount of water can be carried into the wedge via a mechanism of lateral transport via solid matrix flow and vertical transport of the aqueous phase. However, because their calculations assumed a constant-viscosity wedge, more recent results suggest that the P–T–t paths will be quite different from temperature-dependent wedge results. Arcay et al. (2005) show that recycling of water to significant depth is possible in their subduction calculations. Even in cases where a significant amount of water is carried to depth, the mantle wedge is significantly hydrated within 250 km of the trench, effectively broadening the low-viscosity zone near the tip of the slab wedge, consistent with the geoid and topography observations, that require a low-viscosity wedge (e.g., Billen et al., 2003). The idea that volatiles may only be preserved in the slab at high convergence rates (and perhaps only in unusually high convergence rates) was put forward by Staudigel and King (1992) to explain isolated geochemical anomalies such as theDupal anomaly.

Serpentine may be the most important mineral for water transport (Pawley and Holloway, 1993) and is the most abundant of the hydrous minerals in ultramafic rocks under 500°C (Hacker et al., 2003). There have been a number of slab models that suggest that the mantle wedge is significantly (up to 50%) serpentinized (Peacock and Hyndman, 1999; Peacock, 2001; Hyndman and Peacock, 2003). Hyndman and Peacock (2003) find a low-density, low-viscosity wedge that is 20–50% serpentenized, which again
would be consistent with the geoid and topography observations (e.g., Billen et al., 2003).

In contrast to the studies above, Iwamori (1998) uses a corner flow (constant-viscosity) model with a porous flow model for the migration of the aqueous fluid. He concludes that the fluid coming off the slab is absorbed in a serpentinite layer nearly parallel to the slab, rather than the entire wedge. This serpentinite layer may extend to 150 km below the depth of the dehydration, advected by the induced mantle wedge flow. Bostock et al. (2002) interpret their seismic image of the southern Cascadia subduction zone as a highly hydrated and serpentinized fore-arc region, which they point out is consistent with thermal and petrological models of the fore-arc mantle wedge. Bostock et al. (2002) further recognize that serpentinized material is thought to have low strength and go on to suggest that this may control the large-scale flow in the mantle wedge.

While the thermal models above all suggest the mantle wedge is highly serpentinized, they provide little constraint on how much water gets into the transition zone. Subducted oceanic crust may become dehydrated before reaching the transition zone (e.g., Schmidt and Poli, 1998); however, peridotite could carry water into the transition zone, particularly by water that has percolated to depth in the slab along extensional faults formed at the outer rise (e.g., Peacock, 2001; Rüpke et al., 2004). So perhaps the best argument to be put forward is based on direct observations. Typical MORB contains 50–200 ppm water (Dixon et al., 2002). Wood (1995) shows that the seismically determined width of the 410 km phase change is consistent with \( \leq 200 \text{ ppm water at the transition zone, suggesting that the upper bound might be that the transition zone is at most as wet as the MORB source region. Measurement of water in the transition zone via electrical conductivity could provide another constraint (Huang et al., 2005).}

We have said little about water leaving the transition zone including whether water can be trapped in the transition zone (cf. Bercovici and Karato, 2003; Karato et al., 2006; Dasgupta et al., 2005; Hirshmann, 2006). Without attempting to evaluate the transition zone water filter model, it is important to recognize that the transport of water and the fate of the subducted slab may be significantly decoupled (Ricard et al., 2006). While many slabs appear to penetrate into the lower mantle, some slabs may stagnate in the transition zone for a significant period of time. In addition, there is also evidence that components of the slab may separate, depending on the rheology of the slab (e.g., Christensen, 1998; Richards and Davies, 1989; Gaherty and Hager, 1994; van Keken et al., 1996).

We will leave the fate of water and the fate of slabs beyond the transition zone to other reviews (cf. Chapters 2.04 and 7.10).

### 7.08.6.4 How Can We Resolve the Nature of the Deformation in the Seismogenic Zone?

New geophysical measurements are helping to provide additional needed constraints on the deformation of the upper 50–100 km of subduction zones. For example, residual mantle exposures in the accreted Talkeetna arc, Alaska provide rock analogs for the arc-parallel flow that is inferred from seismic anisotropy at several modern arcs (Mehl et al., 2003).

Bevis et al. (2001) interpret the interseismic crustal velocity field of the central Andes using a simple three-plate model in which the Andean mountain belt is treated as a rigid microplate located between the Nazca and South American Plates. They obtain our best fit to the geodetic velocities if the main plate boundary is fully (100%) located between depths of 10 and 50 km and 8.5% of Nazca–South American Plate convergence is achieved in the back arc (by underthrusting of the Brazilian Shield beneath the Subandean zone).

Cassidy and Bostock (1996) observe shear-wave splitting in three-component broadband recordings of local earthquakes near southern Vancouver Island, British Columbia. These measurements constraint shear-wave anisotropy in the continental crust above the subducting Juan de Fuca Plate and support weak coupling between the downgoing Juan de Fuca Plate and the overlying North America Plate, because the principal stress is perpendicular to the direction of subduction, and the S-wave splitting above the subducting plate is nearly perpendicular to the SKS splitting direction in the upper mantle beneath the subducting plate.

### 7.08.7 Summary

Subduction zones are complex regions of the Earth with challenging physical and chemical processes that impact global geodynamics, geochemistry, and thermal evolution of the planet. While we have a fairly good general understanding about the processes occurring in subduction zones, many details...
(and a few major pieces) remain elusive. It is increasingly evident that future progress requires collaboration and communication among a variety of specialists. So perhaps topping any list of future directions or outstanding questions is the need for better communication between researchers of various backgrounds and expertise. While this has become a common statement in our scientific lexicon, when it comes to understanding subduction zones, it is probably also true. Aside from this, we provide our own list of most interesting outstanding questions:

1. Why does Earth have plate tectonics while the other terrestrial planets do not? While this has not really been a topic discussed in this review, it is one of the foremost questions related to our understanding of global planetary geodynamics. It is hard to imagine that we really understand the process of subduction if we cannot explain why Earth is the only terrestrial planet with active subduction.

2. What is the nature of deformation in the seismicogenic zone? Are there observations in addition to fore-arc heat flow that can be used to constrain the deformation and shear heating in this region? Our understanding of the deformation and heat generation in the top 50–100 km of a subduction zone is limited. Unfortunately, the thermal structure acquired in the top 100 km is largely translated down slab dip, so our ignorance of the processes in this region is propagated along with the slab.

3. How strong are slabs and does slab deformation significantly effect the thermal structure of the slab? For many problems of interest, slab strength may not be the major controlling factor, but for understanding the interaction of slabs with phase transformations in the transition zone, understanding the role of slab buoyancy in plate dynamics, and understanding whether slab components are able to separate (which could be significant in the chemical evolution of the mantle) slab rheology is almost certainly critical.

4. How much water is carried into (and out of) the transition zone by slabs? The recognition that wadsleyite has a large water storage capacity has led to the interesting possibility that a large amount of water could reside in the transition zone. With the evidence for a weak mantle wedge thought to be due to dehydration of the slab and/or serpentinization of the mantle wedge, it is by no means clear that slabs retain any water by the time they reach the transition zone. The fate of water at the base of the transition zone is even more obscure. It is important to remember that this may not be a steady-state process. Water may only be carried into the transition zone during relatively unusual periods of fast subduction.

5. How does subduction begin? While there have been some interesting developments in this area related to preexisting regions of weakness and/or lithospheric discontinuities, there is a lot more to understand. It seems the easiest way to make a new subduction zone assume preexisting subduction. This begs the question as to how the first subduction zone started. When subduction first began and what subduction looked like in the early Earth are also open questions.

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References


Creager KC, Chiao L-Y, Winchester JP, Jr., and Engdahl ER (1995) Membrane strain rates in the subducting plate...


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Polet J, Silver PG, Beck SL, et al. (2000) Shear wave anisotropy beneath the Andes from the BANJO, SEDA, and PISCO


