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THE FATE OF DESCENDING SLABS

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INTRODUCTION

Oceanic lithosphere, the stiff plate formed by the cooling of chemically differentiated material that rises and melts under mid-ocean ridges and spreads laterally in the process called sea-floor spreading, has an ephemeral residence at the surface of the Earth. Typically, within 100–200 Ma, densification of the old oceanic lithosphere leads to a dynamic instability causing it to sink into the interior, descending as a layered slab with highly anomalous thermal and chemical characteristics relative to the surrounding ambient mantle. The slab heats as it sinks, but the warming is sluggish, slow to overcome the effect of more than 100 Ma of cooling at the surface, and the negative buoyancy of the slab continues to drive its descent.

Increasing pressure and temperature with depth drive various chemical reactions in the slab, including expulsion of water and other volatiles accumulated at the surface, and destabilization of associated hydrous phases. Suites of mineralogical phase transformations occur within the slab's crustal and depleted-mantle components. Because the slab has been processed through the melting zone at the ridge, partially hydrated, and cooled significantly, the phase equilibria within the slab differ from those of the surrounding mantle. Near 400 km depth an exothermic phase transformation occurs, with $(Mg,Fe)_2SiO_4$ olivine in the slab converting to the high-pressure modified (β) spinel structure. The spinel transition helps to drive the slab onward, although a central core of metastable olivine may persist to greater depths, with eventual transformational faulting

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producing deep earthquakes. Other phase transitions occur within the transition zone (400–720 km deep). The slab continues to sink rather readily, with the leading edge reaching a depth near 660 km about 10 Ma after leaving the surface. But now the slab encounters resistance to further penetration, sometimes resulting in distortion of the slab, and there is a build up of internal compressional stresses oriented in the down-dip direction of the flow. The slab at 660 km depth is still colder than the surrounding mantle by many hundreds of degrees, and appears to have sufficiently high viscosity to retain its integrity as a stress guide. An abrupt cessation of all earthquake activity occurs by 700 km depth. Any slab material that penetrates deeper will undergo an endothermic dissociative transformation of spinel-structured $(Mg,Fe)_2SiO_4$ into perovskite-structured $(Mg,Fe)SiO_3$ and (Mg,Fe)O. Pyroxenes and garnets will also transform to the perovskite structure, which is the predominant mineral form in the lower mantle.

There is no major disagreement on any aspect of this scenario up to this point, but what happens next? Does the slab continue to descend, penetrating as deep as the core-mantle boundary and thus defining the downwelling in a whole-mantle convection system? Is it deflected and retained within the upper mantle of a strongly stratified convective system, isolated from the deeper mantle? Or is there some intermediate convective configuration, involving either variable penetration of the lower mantle or piling up and catastrophic overturn of the accumulated mass? These questions are followed by the issue of how slab material—which eventually gets thermally assimilated but retains its anomalous chemical signature—ultimately cycles through the mantle?

This issue of the fate of subducting slabs has aroused great interest in the earth sciences, as it is intimately linked to the chemical cycling, and thermal and dynamical evolution of the planet (Silver et al 1988, Jordan et al 1989, Olson et al 1990). Fundamental questions about the causes and persistence of geochemical heterogeneity in the mantle, the possible existence of a thermal boundary layer at the transition zone, and episodicity of tectonic motions at the surface are all linked to the issue of whether slabs do or do not sink into the lower mantle. Observations from numerous disciplines have been brought to bear on this question, and periodically, seemingly compelling arguments have been articulated both for and against deep mantle penetration by the subducting downwellings, only to be softened by subsequent recognition of overlooked complexities. Slab penetration debates have often ended in a frustrating draw, and it is fair to say that this remains one of the foremost unresolved questions in the earth sciences. Research activity continues at an intense pace, with new perspectives on the problem emerging from unexpected directions almost annually. The recent developments in this arena will be highlighted here, and the current weight of evidence will be assessed, although any conclusions found here may prove as ephemeral as the slabs themselves.

SEISMOLOGICAL CONSTRAINTS ON DEEP SLAB STRUCTURE

While there are many indirect lines of evidence invoked in favor of layered or whole-mantle convection scenarios based on observed geochemical heterogeneities or notions of early mantle melting and chemical fractionation (e.g. Silver et al 1988, Anderson 1987b), direct detection of subducted oceanic lithosphere is still the most attractive way to assess the fate of slabs. This places a primacy on seismological observations, which are the most precise means available to determine the deep three-dimensional structure within the Earth. For the deep slab issue, seismological information comes in three forms: 1. earthquakes occur within the anomalous thermal/chemical environment of the subducted slab, illuminating the flow by their very presence, at least to their maximum depth extent; 2. the strain release in deep earthquakes reflects the deep slab deformational environment; and most importantly, 3. the seismic waves radiated from the deep events propagate through the surrounding medium, conveying information about the deep structure to surface sensors. Recent seismological investigations of each of these aspects are shedding light on the fate of deep slabs.

Seismicity Tracers of Deep Slabs

The recognition that intermediate and deep focus earthquakes occur in regions where cold oceanic lithosphere has sunk into the mantle played a major role in the development of the theory of plate tectonics in the 1960s (e.g. Isacks et al 1968). The maximum depth to which seismic events occur is about 700 km (Stark & Frohlich 1985), although many Benioff-Wadati seismicity zones do not extend to even this deep. Earthquakes below 100 km are only found in regions of current or recent subduction; anomalously low temperatures are required for any form of abrupt strain energy release. While pore-fluid assisted brittle failure or frictional sliding are probably responsible for most earthquake activity down to depths of 300 km, earthquake activity at greater depths appears to require a different mechanism. Recent demonstrations that, in the presence of deviatoric stresses, some important upper mantle and slab component mineral phase transformations involve shear faulting-type instabilities (e.g. Green & Burnley 1989, Green et al 1990, Kirby et al 1991, Burnley et al 1991, Meade & Jeanloz

1991) may provide a mechanism for the occurrence of earthquakes at transition zone depths (400–660 km).

Particularly important phase transitions exhibiting transformational faulting involve olivine, as it transforms to high-pressure modified spinel or spincl structures (Green et al 1992a,b). Olivine is expected to be a significant component of the oceanic lithosphere below the basaltic crust. Experimental work has shown that kinetic effects may inhibit the olivine-spinel phase transitions, causing them to occur below their equilibrium depths for a normal mantle geotherm (near 400 km for olivine-modified spinel and 520 km for modified spinel-spinel) (e.g. Sung 1979, Sung & Burns 1976). A viable explanation for deep earthquake occurrence is that transformational faulting takes place in a kinetically-overdriven tongue of metastable olivine within the colder central portions of the downwelling slab. Travel time patterns from earthquakes in the Japan slab provide some evidence for the postulated low-velocity wedge in the core of the slab at transition zone depths (Iidaka & Suetsugu 1992), although studies in other regions have not revealed such a structure. From this perspective, deep earthquake activity can be viewed as tracking the coldest regions of the slab, and is expected to terminate as the slab heats up sufficiently to overcome the kinetic barriers to transformation or as it goes through higher pressure phase transitions that do not involve faulting instabilities, such as the spinel-perovskite transition in olivine expected near a depth of 660 km (e.g. Ito & Yamada 1982, Ito & Takahashi 1989).

While seismicity distributions intrinsically only illuminate the slab where physical conditions allow earthquakes to occur, analysis of the three-dimensional spatial configuration of the slab in the seismogenic region (see reviews by Burbach & Frohlich 1986, Yamaoka et al 1986, Fukao et al 1987, Chiu et al 1991) can constrain deformations above the 660 km mantle discontinuity. While some slabs, such as Tonga, appear to display buckling and imbrication (e.g. Giardini & Woodhouse 1984, Fischer & Jordan 1991), others, such as the Japan slab show surprisingly straight orientations all the way to 660 km depth. Of particular interest are regions of isolated deep earthquake activity, displaced laterally from the well-defined deep seismic zones in several arcs. Figure 1 (Okino et al 1989) shows an isolated event offset from the deep seismicity in the Izu slab, similar to events found under the northern Kurile arc (Glennon & Chen 1993), Spain, Peru, and the New Hebrides (Ekström et al 1990, Lundgren & Giardini 1992b). Given the special environment required for deep earthquake occurrence, these rare events are usually taken as evidence for horizontal deflection of the deep slab, although the possibility that independent detached slab remnants are responsible for the anomalous events complicates such arguments (Hamburger & Isacks 1987). While the spatial distribution of seismicity intrinsically cannot resolve the maximum depth extent of subduction, the shape of the downwelling revealed by the seismicity does provide a boundary condition for modeling slab kinematics for the aseismic extension of the slab (e.g. Creager & Boyd 1991).

Strain Orientations in Deep Slabs

Seismic waves from deep earthquakes not only reveal the occurrence of the strain release, but also indicate the strain orientation at the source (e.g. Isacks and Molnar 1971, Vassiliou 1984, Apperson & Frohlich 1987) and the total energy release (Richter 1979). A robust feature of deep earthquake focal mechanisms is their strong tendency to have roughly down-dip compressional strain axes, usually interprctcd as indicating resistance to penetration into the lower mantle. With the accumulation of increasing numbers of earthquake focal mechanisms, recent efforts have been directed at assessing the overall slab deformation associated with the deep seismicity. The underlying assumption is that the deep earthquake strain tensors are indicative of the overall slab deformation and can be interpreted in the context of viscous loading of coherent slab structures (e.g. Vassiliou et al 1984, Vassiliou & Hager 1988).

Giardini & Woodhouse (1986) related focal mechanisms in the Tonga subduction zone to the contorted seismicity distribution in the region to infer systematic along-strike shear deformation of the entire slab induced by horizontal flow in the surrounding mantle. The structure of shear zones within the slab that accommodate such distortion can be revealed by seeking alignments of seismicity with focal mechanism nodal planes (Giardini & Woodhouse 1984, Lundgren & Giardini 1992a). Zhou (1990b) conducted a systematic study of focal mechanisms for earthquakes in the northwest Pacific and Tonga-Kermadec, and found evidence for flattening of the compressional strain axes near 660 km depth beneath the Izu-Bonin and Tonga arc, but no clear evidence for this in the Kurile, Japan, or Marianas slabs.

The strain rate of the seismogenic portion of a deep slab can be estimated by summing the contributions from all earthquake mechanisms in the slab. Fischer & Jordan (1991) find that along-arc strain rate decreases relative to the cross-strike strain rate from north to south in the Tonga slab—a result generally compatible with the slab shearing model of Giardini & Woodhouse (1986). A 50% thickening of the seismogenic core of the slab can be accounted for by the seismic moment release. While this indicates that slab thickening can occur due to the resistance to lower mantle penetration, it appears that the main volume of subducted material reaching 660 km depth is aseismic, thus seismic strain release cannot quantify the associated deformation. Another complexity is that thermal and phase

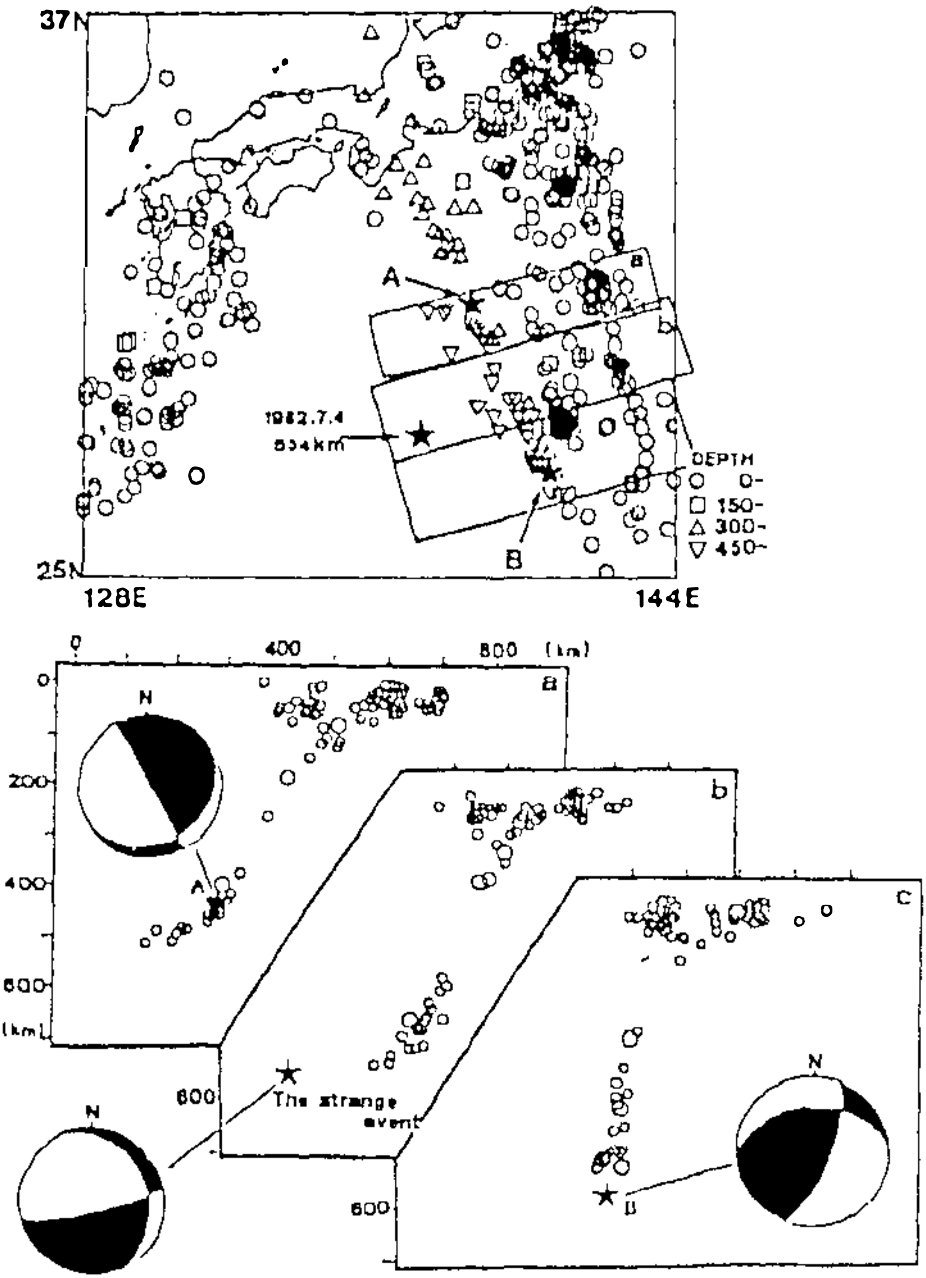


Figure 1 Example of a rare, isolated deep focus earthquake. The July 4, 1982 event occurred at a depth of 534 km, offset about 200 km west from the primary zone of deep activity associated with subduction of the Izu (Pacific) slab (from Okino et al 1989).

transformation induced stresses may overprint any viscous deformational stresses caused by resistance to subduction (e.g. Goto et al 1983, 1985, 1987; Kirby et al 1991). In this case, stresses generated within the slab may account for the down-dip alignment of compressional strain axes.

Seismic Velocity Heterogeneity

Huge volumes of slab have descended in the current subduction zones, greatly exceeding the volumes that are presently seismically active (e.g. Richards & Engebretson 1992, Scrivner & Anderson 1992); thus seismological techniques for imaging the aseismic slab material must be used to assess where the slab material has gone. Since slabs are thermally and chemically distinct from the surrounding mantle, seismic waves transmitted through the slab incur amplitude and travel-time anomalies. For 25 years seismologists have been analyzing wave propagation characteristics of descending slabs, addressing both the seismogenic regions and the deep extension of the slab (Lay 1993). Current approaches to the problem of resolving the fate of slabs include: 1. modeling of travel-time and amplitude anomalies from individual events; 2. tomographic imaging of the mantle around deep earthquakes; 3. waveform modeling of distortions produced by gradients in seismic velocity structure near the deep slab; and 4. tomographic imaging of lower mantle structures and mantle discontinuity topography on both large and small scales.

The basic idea underlying travel-time modeling of deep slabs is that the cold slab should have higher elastic velocities than the ambient mantle, with the tabular geometry of the slab (at least above 660 km) giving rise to simple symmetries in the patterns of relatively early and late arrivals (the former have longer path lengths in the slab material). Numerous studies indicate that in the transition zone slabs do have about 3–5% faster seismic velocities than the surrounding mantle. Thermal models, combined with low-pressure values for temperature derivatives of seismic velocity, may account for this level of heterogeneity (e.g. Creager & Jordan 1986); stronger velocity gradients are expected on the crustal side of the slab. There is, unfortunately, still substantial uncertainty in seismic velocity dependence on temperature at high pressures, complicating predictions of deep slab velocity heterogeneity (Anderson 1987a, 1988). The chemically layered nature of the slab adds detailed velocity structure, such as a high-velocity eclogitic crustal layer, the possible presence of low temperature phases such as $MgSiO_3$-ilmenite, as well as significant ($\pm 5\%$) variations associated with either elevated or depressed equilibrium phase boundaries and/or kinetically suppressed boundaries. Anisotropic characteristics of the slab and the surrounding flow may also contribute to the seismic velocity signature of a deep slab (Anderson 1987b, Kendall & Thomson

1993). Finally, there is the major limitation of seismology: Seismic wave heterogeneity may have multiple causes. For example, it may be very difficult to distinguish aseismic slab extensions into the lower mantle from induced downwellings in a thermally-coupled convection system. Consideration of dynamical models when interpreting seismic images is always crucial.

Most studies of travel-time patterns associated with deep slabs utilize deep earthquakes. The greatest challenge in isolating the slab signature in the seismic wave arrival times is that the location of the seismic source is not known independently. Standard earthquake location procedures project as much as possible of the observed travel-time deviations from the reference Earth model into shifting the source location and origin time. Routine earthquake locations are performed with no slab structure in the Earth model, or at best a very simplified slab model, leading to biased locations of deep events. Patterns of anomalies with respect to the biased locations are expected to be quite different from patterns relative to the actual location (e.g. Creager & Boyd 1992), thus one must account for this effect in any modeling effort.

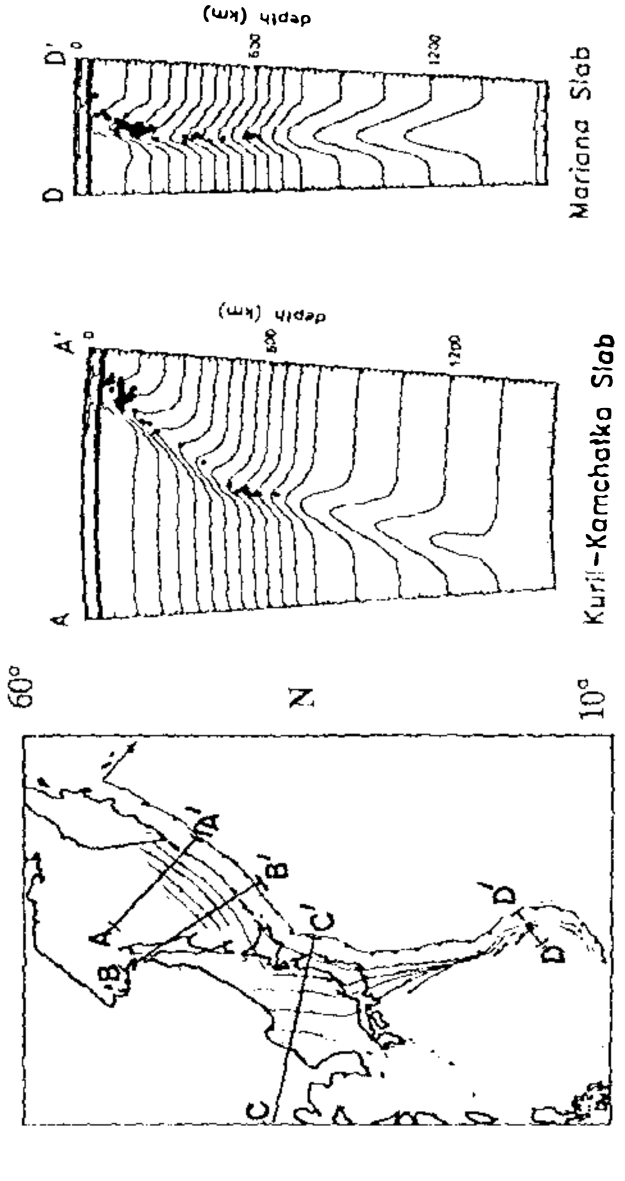
Numerous studies have used the residual sphere method introduced by Davies & McKenzie (1969) to explore systematic patterns in deep earthquake travel times. Residual spheres are projections of travel-time anomalies onto a source focal sphere, indicating the azimuth and take-off angle of each ray path corresponding to observed or calculated anomalies. Jordan (1977), Creager & Jordan (1984, 1986), Fischer et al (1988, 1991) and Boyd & Creager (1991) have conducted a series of residual-sphere modeling efforts, all of which favor tabular high-velocity (3–4%) extensions of slabs four to five hundred kilometers below the deepest seismicity in the Kurile, Japan, Marianas, Tonga, and Aleutian arcs (Figure 2). The smoothed residual patterns for relocated sources, corrected for station statics and low-degree aspherical velocity models of the lower mantle (e.g. Dziewonski 1984), were compared with three-dimensional ray-tracing calculations for thermal slab models embedded in a homogeneous mantle, similarly relocated and filtered. Excellent fits to the data are achieved in most cases, using very simple models. In all of these studies the preferred slab models steepen in dip in the lower mantle, but are otherwise not strongly distorted by passing through the 660 km boundary (Figure 2). Sensitivity tests (Fischer et al 1988) indicate that broadening of the slab by up to a factor of three is allowable, without seriously degrading the fit to the data, but the preferred models do not broaden.

Various concerns have been raised about the surprisingly clean results obtained in this group of residual sphere studies, involving details of the processing (e.g. the heavy filtering used to enhance the smooth slowly-varying patterns in the data, which are assumed to be induced by near-source structure), the data selection (e.g. the use of "noisy" catalog travel-time measurements from nonstandard stations for only down-going P arrivals which span a very limited portion of the focal sphere), the model parameterization (e.g. neglect of possible slab anisotropy), and the heterogeneous Earth correction procedures (e.g. use of low resolution Earth models which probably underestimate actual deep heterogeneity to correct for distant path effects). Several recent studies have attempted to overcome some of the possible difficulties.

The possibility of systematic bias in catalog data sets associated with variations in instrument magnification was raised by Grand (1990), who found that low gain stations tend to report late arrivals. A greater concern about the catalog travel times is the intrinsically large scatter in the data. Gudmundsson et al (1990) analyzed ISC travel-time statistics, separating spatially coherent and incoherent variance, and found low signal to noise ratios for P waves at teleseismic distances. Efforts to manually measure arrival times (e.g. Takei & Suetsugu 1989, Ding & Grand 1992) tend to substantially reduce the range of anomalies in residual spheres for deep events relative to the catalog values.

To isolate the near-source signature of deep slab heterogeneity on travel-times, it is important to suppress contributions from mantle heterogeneity far removed from the source region. Given the predominance of large-scale heterogeneity in both the upper and lower mantles (e.g. Su & Dziewonski 1991, 1992), one cannot safely assume that any slowly varying pattern in a residual sphere must be caused by near-source structure. Zhou & Anderson (1989b) showed that residual spheres for shallow earthquakes in eastern China actually resemble those for deep events in the Kurile slab, suggesting a common component of lower mantle or near-receiver heterogeneity on paths to teleseismic stations. Lay (1983), Schwartz et al (1991a,b), and Gaherty et al (1991) have presented clear evidence for lower mantle and near-receiver contributions to deep earthquake travel-time residuals being underpredicted by standard station corrections or integration through low resolution aspherical Earth models. These distant effects can produce smooth patterns in residual spheres that can bias deep slab modeling. Models with substantially broadened slabs, involving relatively weak lower-mantle heterogeneity, are indicated by these studies.

Confronted with the difficulty of isolating near-source travel-time anomalies, there are two different procedures. As they become available, one can use higher resolution aspherical Earth models (e.g. Inoue et al 1990) to make deep path corrections, or one can adopt an empirical calibration strategy, using data from nearby events to establish improved path corrections, free from the damping and smoothing effects of tomographic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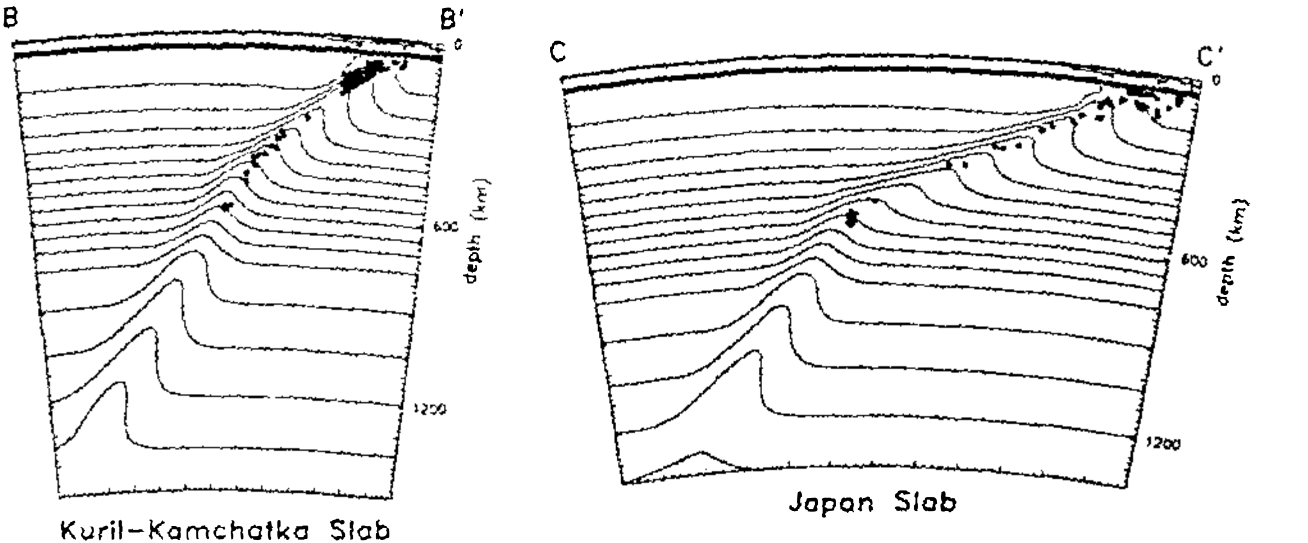

Figure 2 Cross sections through three-dimensional slab velocity models derived by residual sphere analysis of several intermediate and deep focus earthquakes in the western Pacific subduction zones. The seismic zone in the map is contoured in 100-km intervals. In all cases, slab models extend to 1350 km depth, although penetration depths below 1000–1200 km cannot be resolved by the data. Contour interval is 0.25 km/s (from Creager & Jordan 1986).

inversions. Schwartz et al (1991b) applied corrections for a high-resolution lower-mantle shear velocity model beneath North America, finding that it strongly reduced relative S wave travel-time patterns for Kurile slab sources. Zhou & Anderson (1989a) and Zhou et al (1990) used whole-mantle and regional northwest Pacific tomographic models to correct residual spheres for many events in the western Pacific. The latter authors extended the ray parameter coverage of the focal spherc substantially by including close-in observations as well. Consistently, many of the coherent features in residual spheres that give rise to the simple slab structures shown in Figure 2 are substantially reduced, weakening the evidence for simple slab penetration models. There is also evidence for horizontal flattening of some slabs, revealed only in the extended ray parameter coverage.

Since all tomographic models are potentially flawed in terms of providing detailed corrections for specific paths, empirical corrections have also been attempted. Zhou & Anderson (1989a) found that removing average path anomalies computed for many events in the western Pacific weakened slab-like signals in individual event residual spheres, especially for very deep events. Gaherty et al (1991) and Schwartz et al (1991a,b) found that using surface reflected phases to make distant path corrections also greatly reduced the patterns for deep events.

Differential residual spheres, first introduced by Toksöz et al (1971), can be computed for events at different positions in the same slab to cancel distant path effects. While the travel-time anomalies that result are substantially reduced relative to those found by the early residual sphere analyses, these studies have tended to favor the presence of high-velocity material in the lower mantle along portions of the down-dip extension of the Japan and Kurile slabs (Takei & Suetsugu 1989, Okano & Suetsugu 1992, Ding & Grand 1992). For the Marianas, the differential residual sphere method appears to eliminate the high-velocity region below the deepest events (Okano & Suetsugu 1992). The most complete analysis of this type has been for the Kurile slab, where Ding & Grand (1992) favor a model with substantial advective thickening of the plate and lower mantle penetration in the northern part of the arc, and flattening of the plate to subhorizontal in the southern arc. This is consistent with most of the tomographic results discussed later.

While most of these procedures for improving corrections for distant propagation effects lack the comprehensive three-dimensional ray tracing and careful handling of relocation artefacts that characterize the early work on residual spheres (Ding & Grand 1992 is an exception), it is clear that reliance on low resolution models for corrections is inadequate. Interpretation of differential patterns is somewhat more complicated than for individual events, but the confidence gained in the isolation of near-source contributions is substantial. Of course, there is a danger in differential residual sphere and other empirical path calibration procedures of eliminating the desired near-source signal if the path separation is not sufficient. Given the complex three-dimensional geometry of all subduction zones, this is a serious consideration.

If aseismic extensions of slabs have simple geometries, the residual-sphere modeling approach is quite attractive. However, if the slabs are strongly distorted so that no simple slab parameterization can be justified, it is preferable to image the slab heterogeneity by seismic tomography. Numerous tomographic studies of subducted slabs have been built on basic strategies introduced by Aki & Lee (1976) and Aki et al (1977), expanded to handle vastly increasing data sets and model sizes (e.g. Spakman & Nolet 1988). While early models (e.g. Hirahara 1977) addressed the upper mantle structure alone, many studies include the lower mantle below the seismogenic regions. Accounting for source mislocations and deep mantle and near-receiver effects are still major problems, as is the vertical smearing in tomographic images associated with limited crossing ray coverage below the deepest sources (Spakman et al 1989, Zhou 1988). While limited efforts have been made to solve the simultaneous location and velocity determination problem (e.g. Engdahl & Gubbins 1987), most tomographic studies utilize an iterative procedure: The initial catalog locations are relocated in a homogeneous model; then, at least in some studies, they are relocated in a first-step tomographic model using simplified ray tracing. A major problem is that much of the slab signal may be lost in the initial location process, unless the ray path coverage is extensive enough to eliminate the trade-off between slowly varying slab anomalies and slowly varying relocation effects. In the future, inversions starting with an initial slab structure may overcome much of this difficulty.

Tomographic images have been produced recently for the northwest Pacific subduction zones (Kamiya et al 1988, 1989; Suetsugu 1989; Spakman et al 1989; Zhou & Clayton 1990; Nenbai 1990; van der Hilst et al 1991, 1993; Fukao et al 1992; Yamanaka et al 1992), Java (Fukao et al 1992, Puspito et al 1993), Tonga and the New Hebrides (Zhou 1990a), Central America and the Caribbean (van der Hilst & Spakman 1987, 1989), and the Mediterranean (e.g. Spakman et al 1988, 1993; Ligdas et al 1991; Spakman 1991). Strategies for eliminating distant effects have included the omission of teleseismic ray paths altogether (Zhou & Clayton 1990), correction for mean station residuals at teleseismic distances (e.g. Kamiya et al 1988, van der Hilst et al 1991), and simultaneous inversion for the complete global structure using a variable block size for the near-source and deep mantle portions of the model (Nenbai 1990, Fukao et al

1992). Improvements in the source location problem have been achieved by including both regional and teleseismic data, as well as surface reflections (pP, PP), which enhance the focal sphere coverage and reduce errors in source depth (van der Hilst & Engdahl 1991, 1992).

The resulting tomographic images of deep slabs are relatively consistent in detecting fast velocity slab structures encompassing the deep earthquake locations. However, the general characteristics of deep slabs are significantly more complex than the models shown in Figure 2. The most recent models (Figure 3), which utilize the most reasonable strategies for controlling mislocation effects and distant contributions, are suggestive of a vertical extension of the central Kurile slab to depths of about 1100 km under the Sea of Okhotsk (van der Hilst et al 1991, 1993), flattening to subhorizontal for the southern Kurile slab (van der Hilst et al 1991, 1993; Fukao et al 1992), and flattening and thickening of the Japan slab (van der Hilst et al 1991, 1993; Fukao et al 1992). Much of the flattening in the southern Kuriles and Japan appears to occur above 660 km depth, but there is evidence for some high-velocity material extending below this depth in the model of Fukao et al (1992), which could be consistent with earlier studies by Kamiya et al (1989), Creager & Jordan (1986), and Takei & Suetsugu (1989), but this proves to be a minor feature relative to the horizontally deflected material. The central Izu-Bonin slab appears to deflect horizontally above the 660 km discontinuity, extending across the entire width of the Philippine plate to the Ryukyu arc (van der Hilst et al 1991, 1993; Fukao et al 1992), while a broadened high-velocity anomaly extends vertically below the Marianas subduction zone to depths of about 1200 km. The Middle American slab appears to extend well below the cutoff of seismicity (van der Hilst et al 1993), possibly connecting up to a broad fast velocity region more than 1000 km deep beneath the Caribbean imaged in earlier work (Jordan & Lynn 1974, Lay 1983, Grand 1987). Large-scale tomographic images of the entire mantle, while lacking resolution of small structures such as slabs, do suggest generally fast lower-mantle velocities below the circum-Pacific (e.g. Dziewonski & Woodhouse 1987, Su & Dziewonski 1991), but a direct connection of these very large-scale features with slab anomalies is not yet established. Distinguishing continuous downwellings from thermally-coupled layered systems is particularly difficult for large-scale tomographic models.

In addition to seismic wave travel times, amplitude and waveform anomalies have been analyzed to detect any effects of deep slab velocity heterogeneity. Ray path deflections caused by slab structure have been analyzed for more than 20 years (Sleep 1973), but new broadband data offer improved resolution of wavefield effects. Theoretical calculations predict that the high-velocity heterogeneity of slabs should cause complex diffrac-

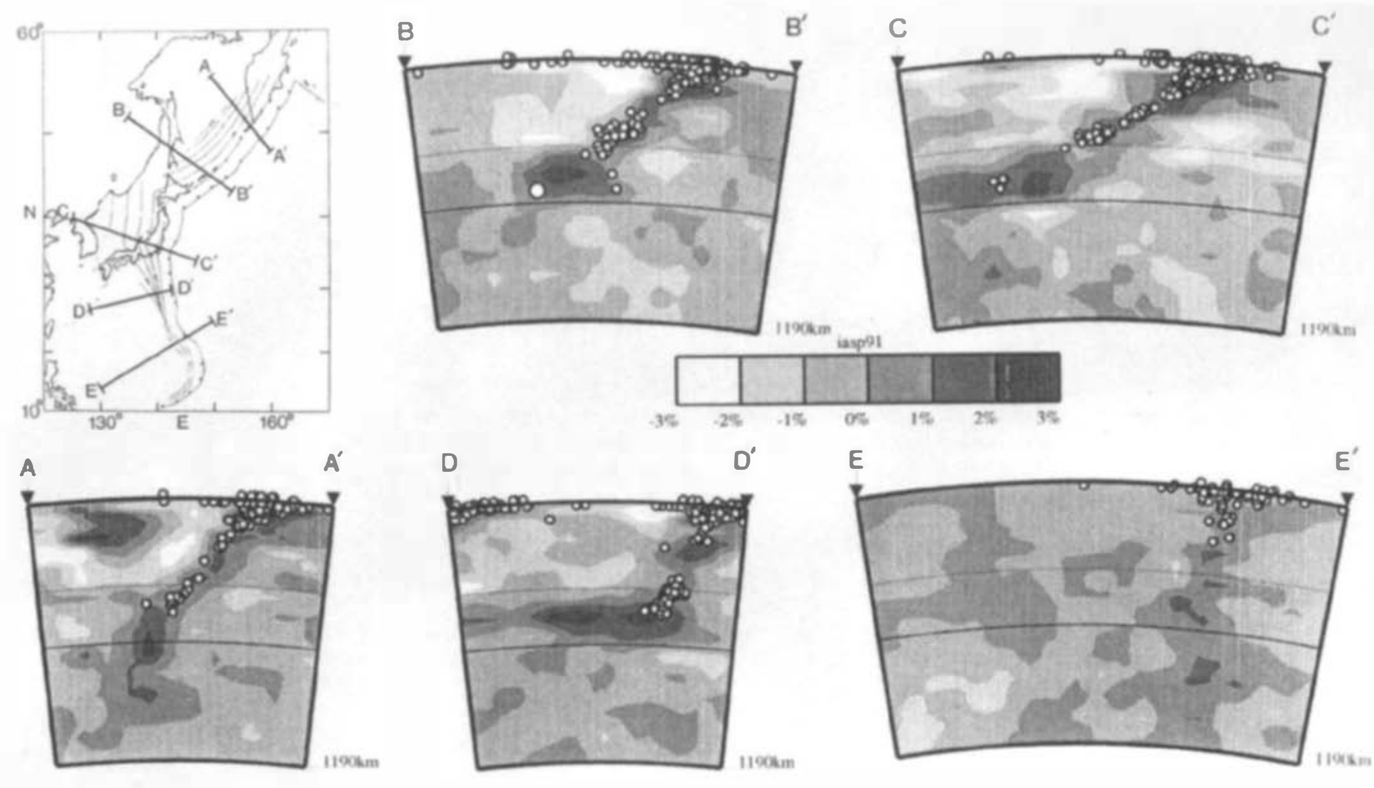

Figure 3 Cross sections through three-dimensional tomographic P wave velocity models for western Pacific subduction zones. The seismic zone in the map is contoured in 100-km intervals. Faster velocity material (relative to the reference iasp91 model) is darker, with shading in 1% velocity variations. Dots are earthquake hypocenters along each of the cross sections. Mantle discontinuities at depths of 400 and 660 km are shown. Note that high velocity material correlates with the deep seismic zones, and shows evidence for both penetration (AA′, EE′) and flattening (CC′, DD′) at the top of the lower mantle (from van der Hilst et al 1991, 1993).

tion, defocusing, and multipathing effects (Vidale 1987, Cormier 1989, Witte 1989, Weber 1990, Sekiguchi 1992, Vidale et al 1991). Some observations of waveform distortions for deep earthquakes do indicate the presence of complex velocity gradients along the ray paths (Beck & Lay 1986, Choy & Cormier 1986, Silver & Chan 1986, Lay & Young 1989, Schwartz et al 1991a, Fischer 1990), but there is still no convincing demonstration that these effects constrain the fate of deep slabs. The deep mantle has many heterogeneities that can distort waveforms (Lay 1983, Vidale & Garcia-Gonzalez 1988), and it appears that many waveform effects are superimposed to produce the scattered behavior that is observed. While there is strong evidence that shallow slab structures do affect teleseismic waveforms (Engdahl & Kind 1986, Engdahl et al 1988, Gubbins & Snieder 1991), it appears that deep slab velocity gradients are not as strong as indicated by the early thermal models for deep slabs. In particular, defocusing effects, which are predicted to be readily observable for the simple tabular thermal models, have proven difficult to detect (e.g. Vidale 1987, Gaherty et al 1991, Fischer 1990), although there have been some attempts to include amplitude information in the construction of deep slab models (Suetsugu 1989).

One additional area in which seismology is contributing to our understanding of deep slabs is in the mapping of topography on the 660 km discontinuity. While this seismic discontinuity is most likely the result of the spinel-perovskite phase transformation, rather than a chemical boundary, its undulations do provide constraints on the lateral temperature variations near the boundary. Several studies have attempted to detect perturbations of the boundary in the vicinity of an impinging slab, using converted and reflected phases. Depressions of the phase transition by tens of kilometers have been detected under Tonga (Bock & Ha 1984, Richards & Wicks 1990, Wicks & Richards 1991, Vidale & Benz 1992), the Izu-Bonin/Marianas (Vidale & Benz 1992, Wicks & Richards 1993), the southwest Pacific (Revenaugh & Jordan 1991), South America (Vidale & Benz 1992) and the Kuriles (Shearer 1991, Shearer & Masters 1992). While these studies support the evidence for a negative Clapeyron slope for this phase transition (Ito et al 1990), they cannot be used to argue against chemical stratification of the mantle which could prevent the slabs from penetrating. This is because postulated chemical differences between the upper and lower mantle that are needed to account for differences in the average densities (Jeanloz & Knittle 1989, Stixrude et al 1992) may not have a strong seismic velocity contrast (Jeanloz 1991). The primary application of boundary topography is instead as an indicator of cumulative thermal anomaly near the boundary. For example, Shearer & Masters (1992) show a 1500 km broad depression of the 660 km discontinuity landward from the southern Kurile arc. This is consistent with the cooling effect of a horizontally deflected slab, as imaged in the tomographic and differential residual sphere analyses mentioned above. If all recently subducted slabs are piled up in the transition zone, one would expect broad depressions of the 660 km discontinuity around all margins of the Pacific ocean, which is not generally observed (Shearer & Masters 1992), although current topographic models are very limited in resolution.

DYNAMICAL FACTORS INFLUENCING SLAB FATE

While seismology provides tantalizing images of heterogeneous elastic properties of the interior, quantification of the actual subduction process requires insights from numerical modeling. There have been great advances in our ability to simulate viscous flow in the mantle for increasingly realistic parameter ranges, and a wide variety of possible influences on the mantle system have now been explored. One of the surprising perspectives to emerge is that there are many factors that should act to inhibit deep mantle penetration by the downwelling slab, even if the mantle is not chemically layered. Some of the recent developments will be discussed below.

Effects of Chemical and Viscous Stratification

Given sufficient intrinsic chemical density contrast between the upper and lower mantle, it is not difficult to develop rigorously stratified mantle convection. Christensen & Yuen (1984) estimate that a 5% contrast is sufficient to deflect the slab, preventing mixing between the upper and lower mantle. This value is at the upper end of estimates of a bulk chemical density contrast between the upper and lower mantle based on mineral physics experiments on silicate perovskite and magnesiowüstite (e.g. Knittle et al 1986, Jeanloz & Knittle 1989, Stixrude et al 1992). The critical physical parameter underlying these arguments is the thermal expansion coefficient for silicate perovskite at high pressure, critical to estimating the density predicted for various lower mantle compositions. While there is controversy over the experimental determination of perovskite thermal expansion (Chopelas & Boeler 1989), most measurements have been made either at atmospheric pressure (Knittle et al 1986, Ross & Hazen 1989, Parise et al 1990), or at pressures below the stability field of perovskite (Mao et al 1991, Wang et al 1991). New measurements at simultaneous high pressure and high temperature that are just beginning to emerge (Funamori & Yagi 1993) indicate less decrease in thermal expansion with pressure than previously thought. Confirmation of these results will be critical for resolving the argument for a component of chemical contrast between the upper and lower mantles.

Density contrasts lower than 5% allow partial penetration of the downwelling slab, and the slab can sink all the way through the lower mantle for density contrasts less than 2%, which is at the lower range of the estimated chemical density contrast. The relative viscosity of the subducting slab influences its behavior, with higher viscosity allowing it to rctain a concentrated negativc buoyancy that abets penetration. Experimental work (Kincaid & Olson 1987) indicates that slab geometry also influences slab penetration, with steeper dipping slabs penetrating more effectively. Silver et al (1988) argue that penetrative convection, in which the compositionally-induced density increase across the transition zone is 2% or lcss, is a viablc cxplanation for a variety of dccp earth constraints. Continued efforts to quantify any bulk chemistry difference, between the upper and lower mantle are critical for assessing the role of any chemical layering, as are efforts to detect any seismic boundary that can be attributed to a compositional contrast. Qualitatively, the evidence for thickened and flattened slab structures presented above is consistent with thc range of behavior expected for moderate chemical stratification, but it is now clear that other effects can also produce such slab distortions.

Viscous stratification of the mantle provides an effective mechanism for deforming subducting slabs, and can readily lead to broadened downwellings in the deep mantle. An increase in viscosity across the 660 km boundary by a factor of 10 to 100 has been invoked in several efforts to model the geoid (Hager et al 1985, Hager & Richards 1989, Hager & Clayton 1989), although less abrupt viscosity increases have also been proposed (e.g. Forte et al 1992). An abrupt increase in viscosity can strongly deform the subducting slab, causing it to fold ovcr on itself in thc uppcrmost part of the lower mantle (Gurnis & Hager 1988). The viscous flow models can produce many features similar to those seen in tomographic images. Thus, viscous stratification is a viable mechanism for distorting the deep slab and advectively thickening it; but alone it can not prevent eventual sinking of the subducted mass unless extreme viscosity contrasts are invoked. A combination of a vicosity increase and a chemical contrast can be reconciled with the geoid, but requires a substantial accumulation of subducted material at the top of the upper mantle, and a deeply (several hundred kilometer) depressed compositional discontinuity. This could be reconciled with the broad high-velocity regions seen tomographically beneath the Marianas, Java, and Japan.

Effects of Chemical Buoyancy of the Slab

The distinct phase equilibria of slabs must also be considered when assessing their dynamics. Ringwood & Irifune (1988) and Anderson (1987b) evaluate the intrinsic density of the enriched, eclogitic crustal and depleted, harzburgitic mantle portions of subducted slabs as a function of depth. If thermally equilibrated with the surrounding mantle, these components could be dynamically stabilized in the transition zone. The intrinsic chemical buoyancy can affect subduction, and perhaps lead to separation of the slab components. However, numerical calculations of viscous slabs including chemical buoyancy effects indicate that thermal buoyancy predominates, and even if an increase in viscosity with depth leads to strong folding and deformation of the slab, the chemical buoyancy effects are secondary in the overall dynamics of the slab (Richards & Davies 1989, Gaherty & Hager 1993). These calculations indicate that unless the slab material can be held stagnant in the transition zone for sufficient time so that thermal equilibration is achieved, chemical buoyancy effects will not play a major role in determining the fate of the slab. It is important to note that if the slab material does penetrate below the transition zone, the basaltic component may again become denser than the surrounding mantle, adding negative buoyancy to the slab (Ringwood & Irifune 1988). Continued work on the high-pressure density of slab materials is needed to address this issue.

Effects of Phase Transitions in the Slab

There has been extensive work recently on the dynamic effects of the endothermic and exothermic phase transformations affecting upwellings and downwellings in the transition zone. Early work on the 660 km endothermic transition (e.g. Christensen 1982, Christensen & Yuen 1984, 1985) indicated that a large negative Clapeyron slope (-4 to -8 MPa/K) would be required to induce stratified convection, which was viewed as unlikely. Recent estimates of the Clapeyron slope vary from -2.8 to -4.0 MPa/K (Ito & Takahashi 1989, Ito et al 1990), and updated computations at higher Rayleigh number (*Ra*) have incorporated these values. Calculations with $Ra = 10^6$–5×10^7 have been conducted in a two-dimensional Cartesian geometry to assess the effects of the 660 and 400 km phase transitions (Liu et al 1991, Zhao et al 1992, Steinbach & Yuen 1992). Increasing Rayleigh number tends to promote layering of the system, as does the coexistence of two phase transitions (Figure 4). Spherical axisymmetric calculations indicate similar phenomena (Machetel & Weber 1991, Peltier & Solheim 1992), as well as a tendency for time-dependent behavior involving catastrophic flushing out of the accumulated downwellings in the upper mantle. Catastrophic overturn is observed in spherical axisymmetric (Machetel & Weber 1991, Solheim & Peltier 1993), two-dimensional Cartesian (Weinstein 1993), and three-dimensional Cartesian (Honda et al 1993) geometries, but the degree of episodicity of these instabilities is reduced in fully three-dimensional spherical flow regimes

(Tackley et al 1993), where flushing events are more localized phenomena. The latter calculations are for a realistic mix of bottom and internal heating, with a volume-averaged Rayleigh number from internal heating and superadiabaticity of 1.9×10^7. Thc occurrence of intermittent overturning events is an efficient mechanism for producing a heterogeneity spectrum dominated by long-wavelength features, as observed for the Earth. However, it is not clear that upper and lower mantle heterogeneities are as decorrelated as expected for the quasi-stratified flow regimes that have been computed (Jordan et al 1993). At this time, the effects of high Rayleigh number flow with strongly temperature-dependent viscosity and phase transitions have yet to be established. It is likely that the stiffness of the slab will strongly influcnce the ability of the phase transitions to impede their downward motions, so it is premature to base too many conclusions on these calculations.

Historical Slab Accumulations

The history of subduction places important constraints on the interpretation of seismic models for deep slabs. Estimates of the cumulative subducted material in circum-Pacific regions have been found to correlate with long-wavelength seismic velocity heterogeneity in both the transition zone (Scrivner & Anderson 1992) and the radially-averaged lower mantle (Richards & Engebretson 1992). Given the long time required for thermal equilibration of the slab material (e.g. Shiono & Sugi 1985), one expects large volumes of fast-velocity material near or below subduction zones, unless the slabs readily sink deep in the lower mantle. Degree 2 harmonic terms of aspherical velocity heterogeneity and accumulated slab volume do correlate rather well in the transition zone and uppermost part of the lower mantle (Scrivner & Anderson 1992), which can be used to argue either for or against slab penetration into the lower mantle, as long as broadening and thickening of the slab is allowed. Of course, degree 2 is extremely large scale, even relative to the volumes of accumulated slab material, and higher degree components (admittedly, less well resolved in the current seismic models) show far less correlation. Efforts to relate specific deep slab structures to the history of subduction are just beginning (Engebretson & Kirby 1992, Grand & Engebretson 1992, van der Hilst et al 1992), but this promises to be a fruitful area of research.

SUMMARY AND SYNTHESIS

There is as yet no unambiguous resolution of the fate of deep slabs. The collective seismological, geodynamical, and mineralogical evidence summarized above can perhaps best be reconciled with a model of the

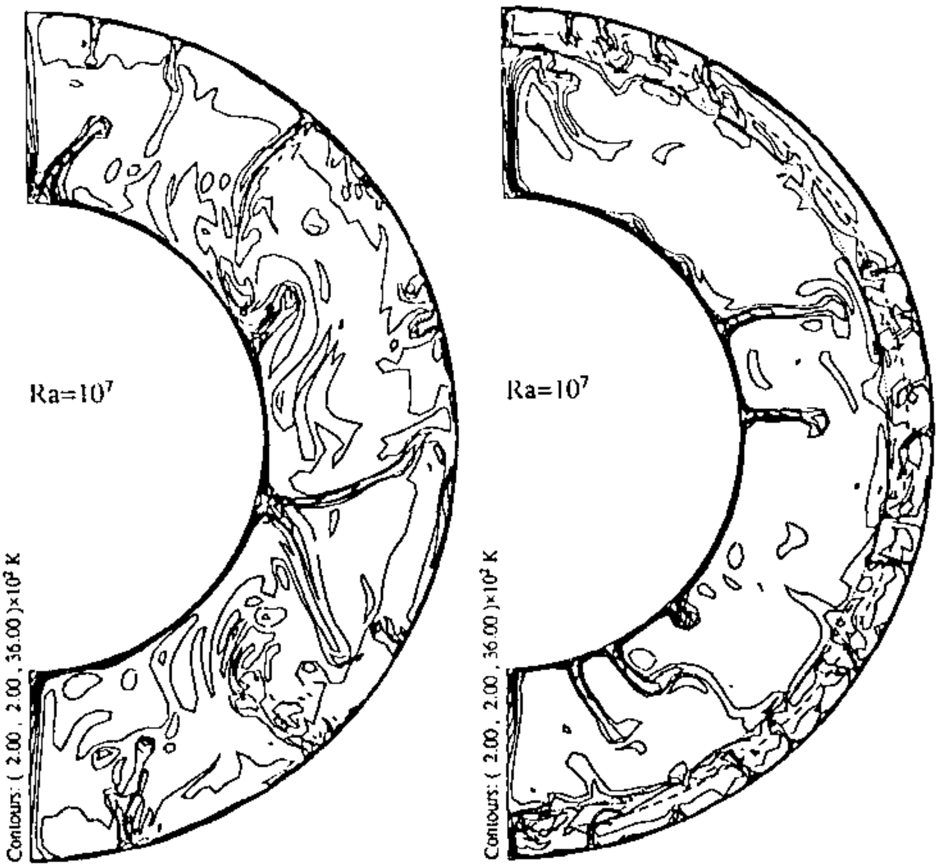

Figure 4 Instantaneous temperature fields from two axisymmetric simulations of mantle convection at a Rayleigh number of 10^7. Heating is entirely from below. The calculations on the left do not include any phase transitions, while those on the right include both a model olivine-β spinel transition and a model spinel-perovskite transition, with 50 km thick phase loops. Note that neither hot ascending plumes nor cold descending plumes appear to penetrate the 660 km discontinuity on the right (from Peltier & Solheim 1992).

Earth in which descending slabs encounter resistance to lower mantle penetration, as a result of increasing viscosity with depth, intrinsic resistance from the endothermic perovskite phase transitions, and possibly a few percent chemical density contrast between the upper and lower mantles. Apparently, the termination of seismicity near 660 km depth can be attributed to combined effects of thermal assimilation and completion of phase transformations in the slab (recall that seismicity in many slabs does not even penetrate this far), and, for seismogenic slabs that exceed this depth, silicate material transforming into perovskite, extinguishing the mechanisms causing deep earthquakes. The aseismic portion of the slab broadens and deflects at and below the phase boundary. Slabs with rapid trench motions, such as the Izu slab, may flatten above the boundary.

Because the thermal inertia of the slab is still high at 660 km depth, the intrinsic chemical buoyancy of the slab components, which would tend to retain the slab in the transition zone if thermal equilibration were achieved, is inadequate to keep the slab from sinking. The lower mantle downwelling appears to be slowed relative to the upper mantle rates, with large slab accumulations and entrained material producing long-wavelength heterogeneities in the lower mantle below active subduction zones. This heterogeneity is imaged in residual sphere and tomographic models. Broadening, and perhaps buckling of the slab, reduces the lower-mantle velocity gradients involved, making the deep slab anomalies inefficient diffractors of seismic waves. The variability of tabular extensions of deep slabs, some lying horizontal and others penetrating vertically, can be attributed to the variable tectonic history and specific geometry of particular zones, as material piles up and deforms near the top of the lower mantle. While the slab material may tend to flush out catastrophically, the images of accumulations above the 660 km discontinuity do not show the massive volumes expected for the total history of subduction. Instead, continuous piling up of material just above, and within the uppermost portion of the lower mantle is suggested by the collective seismic images. Of course, some flushing events may be ongoing, while in other places accumulations are just beginning.

Other interpretations can of course be made, but it seems that most of them have serious flaws. Undistorted slab penetration into the lower mantle is implied by some residual sphere models, but this is hard to reconcile with the abundant evidence for significantly broadened anomalies below 660 km depth, as well as the clear evidence for horizontal deflection of some slabs. A penetrative convection scenario, with relatively little deformation of the slab can be invoked to accommodate the duality of slab behavior (e.g. Silver et al 1988), but this requires that the Earth system resides within a fairly restrictive range of conditions. If the evidence for a bulk difference in chemistry between the upper and lower mantle holds up, a modified version of the penetrative convection scenario—in which a significant broadening of deep slab material is included—may still be the best option. Horizontal deflections of all slabs in a rigorously stratified model is difficult to reconcile with the evidence for tabular lower-mantle velocity anomalies down-dip of several slab trajectories. While several-hundred-kilometer deflections of a chemical discontinuity are expected, there is evidence for much deeper slab-like structures which cannot be readily explained, unless one invokes downwellings in a thermally coupled convection system. Horizontally deflected slabs are intrinsically more difficult to detect than steeply dipping extensions, but given the large cumulative volumes of subducted material, it is surprising that more dramatic transition zone anomalies near subduction zones are not apparent if this scenario is correct. Improved tomography of the transition zone and mapping of mantle discontinuity topography can contribute significantly to this problem.

Clearly, improved seismic images of deep slab structure, additional experimental work on chemical contrasts between the upper and lower mantles, and three-dimensional spherical convection calculations with phase transitions and temperature-dependent viscosity (to better incorporate slab properties) are all needed before there will be any consensus on the fate of descending slabs.

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