Seismic Constraints on the Morphology of Deep Slabs

KAREN M. FISCHER AND THOMAS H. JORDAN

Department of Earth, Atmospheric and Planetary Sciences
Massachusetts Institute of Technology, Cambridge

KENNETH C. CREAGER

Geophysics Program, University of Washington, Seattle

Residual sphere images from deep earthquakes not only detect the presence of slab-associated velocity anomalies but also lend insight into the flow and deformation of lithosphere subducted into the lower mantle. We have compared travel times from deep events in the Kuril and Mariana arcs with the seismic velocity anomalies implied by kinematical models that thicken the slab perpendicular to its plane by reducing the vertical velocity of the flow with depth. We assume that the details of the deformation (whether the slab buckles, imbricates, fragments, etc.) are averaged out along the ray paths, and hence our models constrain the scale, not the mode, of slab thickening. The deep event travel times are best fit by undeformed models, but the ability of the residual sphere method to resolve slab thickness is limited by ray bending effects. Although the Mariana times are consistent with advective thickening factors of 5 or more, factors larger than 3 are ruled out by the Kuril data. For all models examined, the data require that slab material extends to depths of 900-1000 km.

Global tomographic models and regional studies which delineate high-velocity anomalies in the lower mantle beneath zones of Cenozoic subduction are consistent with our results, as is recent work on pulse distortion by aseismic slabs. Comparison of observed and predicted rates of seismic moment release suggests that if substantial advective thickening does occur, it is largely aseismic.

INTRODUCTION

From a seismological point of view the easiest place to map deep mantle flow is in subduction zones, where the lateral gradients in seismic velocities are large and are frequently illuminated by deep focus sources. Studies of travel times from subduction zone earthquakes are providing an increasingly detailed picture of the morphology and extent of the downgoing lithosphere. In a series of papers [Jordan, 1977; Creager and Jordan, 1984, 1986; Fischer et al., 1986] we have used residual sphere images from Kuril-Kamchatka, Japan, Mariana, and Tonga-Kermadec earthquakes to constrain the strike, dip, and extent of slab material thrust down along the western margin of the Pacific plate. We have found that in all deep focus zones examined thus far, lithospheric slabs must penetrate to depths of at least 1000 km in order to match the amplitude and pattern of travel time anomalies. This large flux of lithosphere into the lower mantle, greater than 100 km$^3$/yr in the western Pacific alone, is presumably balanced by mass transport from the deep mantle by upwelling plumes. Therefore the detection of aseismic slabs below deep focus zones argues for deep circulation as a component of plate tectonic return flow and against the hypothesis that the mantle is stratified into two convecting systems with a sharp chemical boundary near the 650-km discontinuity [Jordan et al., 1988].

Because the primary focus of our previous studies was to establish the existence of slab material below this discontinuity, we parameterized the downgoing lithosphere as a simple detached thermal boundary layer which undergoes minimal deformation en route to the deep mantle. In the case of the Tonga-Kermadec slab, residual sphere images suggest a thickening of the slab perpendicular to its strike [Fischer et al., 1986], but the data in the northwest Pacific can be satisfied by models in which the particle velocities are restricted to the slab plane, with the only significant cross-strike thickening of the slab due to thermal conduction, as opposed to lateral advection [Creager and Jordan, 1986]. As noted by Creager [1984] and Creager and Jordan [1986], however, the ability of the residual sphere method to resolve slab width is limited by ray bending effects. They have demonstrated that the trade-off between slab length and slab width is weak and that slab structures considerably wider than simple thermal plate models can satisfy the residual sphere data for the Marianas.

Considerable evidence suggests that slabs encounter resistance to subduction with depth, particularly in the vicinity of the 650-km discontinuity. Isacks and Molnar [1969, 1971] documented that focal mechanisms from earthquakes occurring below 400 km require downdip compression, and their conclusion has been confirmed in a number of subsequent studies [e.g., Vassiliou, 1984; Giardini, 1984]. The data are particularly persuasive in the case of the Tonga deep seismic zone, with its large release of seismic energy [Richter, 1979] and moment tensor solutions, suggesting the lateral displacement of lithospheric material perpendicular to the arc [Giardini and Woodhouse, 1984]. Of course, these results do not preclude slab penetration into the lower mantle beneath Tonga, and the detection of high-velocity material in the slab plane below 650 km by residual sphere methods indicates that this penetration probably does occur [Fischer et al., 1986]. A model in which the vertical component of flow velocity decreases in the transition zone, causing the lithosphere to thicken by the advection of material away from the slab plane, would reconcile the moment tensor and seismicity data with the residual sphere results.

An increase in viscosity between the upper and lower mantle is one mechanism that could cause the slab to slow and deform while passing through the 650-km discontinuity. Hager et al. 4773
FISCHER ET AL.: MORPHOLOGY OF DEEP SLABS

TABLE 1. Source Parameters of Events Used in This Study

<table>
<thead>
<tr>
<th>Event</th>
<th>Date</th>
<th>Time, UT</th>
<th>Latitude, °N</th>
<th>Longitude, °E</th>
<th>Depth, km</th>
<th>mb</th>
<th>Region</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Jan. 29, 1971</td>
<td>2158:03.2</td>
<td>51.69</td>
<td>150.97</td>
<td>515</td>
<td>6.0</td>
<td>Kuril-Kamchatka</td>
</tr>
<tr>
<td>2</td>
<td>Aug. 30, 1970</td>
<td>1746:08.9</td>
<td>52.36</td>
<td>151.64</td>
<td>643</td>
<td>6.5</td>
<td>Kuril-Kamchatka</td>
</tr>
<tr>
<td>3</td>
<td>Oct. 30, 1979</td>
<td>0137:06.6</td>
<td>18.77</td>
<td>145.16</td>
<td>585</td>
<td>5.6</td>
<td>Mariana</td>
</tr>
<tr>
<td>4</td>
<td>July 5, 1972</td>
<td>0119:27.3</td>
<td>18.70</td>
<td>145.10</td>
<td>625</td>
<td>5.5</td>
<td>Mariana</td>
</tr>
</tbody>
</table>

[1983] and Vassiliou et al. [1984] show that small increases in viscosity may induce downdip compression in the deeper regions of the slab, while allowing the slab to penetrate the 650-km discontinuity without much distortion of dip and width, whereas viscosity jumps of several orders of magnitude confine slab flow to the upper mantle. Intermediate viscosity increases allow the slab to penetrate but significantly shorten and thicken it [Hager, 1986]. Most authors agree there are large viscosity variations in the uppermost mantle, but the question of how viscosity changes in the deep interior is the subject of a long-standing debate. Although an essentially isoviscous mantle is favored by some authors [Pelletier, 1974, 1981; Pelletier and Wu, 1982; Cathles, 1975], a moderate viscosity increase in the transition zone does not appear to be precluded by either the postglacial rebound data [McConnell, 1968; Walcott, 1973] or the systematics of activation energies and volumes [Sammis et al., 1977]. Recent support for viscous stratification has been obtained by Hager [1984] and Hager et al. [1985], who call for a viscosity increase by a factor of 10–30 at 650 km in order to satisfy the long-wavelength geoid.

Residual sphere images of deep slab penetration and the various hypotheses regarding deformation of slab material are motivating the development of new techniques for mapping subduction flow. A high-velocity slab acts as an "antiwaveguide," reducing the amplitudes and broadening the waveforms of the seismic waves traveling along it [Sleep, 1973; Silver and Chan, 1986; Vidale, 1987; Cormier, 1986]. It should be possible to combine this information with travel time data to better constrain the details of slab structure. Although waveform distortions apparently due to deep slab structure have been observed [Silver and Chan, 1986; Beck and Lay, 1986; Cormier, 1986], Vidale [1987] has argued that the lack of significant broadening in SH-polarized pulses propagating downdip from Kuril-Kamchatka deep focus events is evidence for a slab somewhat thicker than the simple thermal plate models of Creager and Jordan [1986]. The problem of interpretation is difficult, however. Accurately computing the propagation of seismic waves through three-dimensional slabs is a nontrivial numerical exercise, especially at angles oblique to the downdip direction where the effects of diffraction and multipathing should be most pronounced [Silver and Chan, 1986; Cormier, 1986]. Moreover, the inverse problem for amplitudes and waveforms is very nonlinear, and the possible effects of heterogeneities elsewhere along the wave path make it highly nonunique.

Travel times are less sensitive than waveforms and amplitudes to the details of slab deformation but, by the same token, are easier to interpret. This paper uses residual sphere images from deep earthquakes in the western Pacific to establish bounds on the advective thickening of the slab as it penetrates through the transition zone into the lower mantle. The forward modeling is similar to our previous studies: a three-dimensional velocity structure is obtained from a temperature field parameterized in terms of a simple kinematical model of subduction flow, which is required to be consistent with the observed seismicity. In this study we generalize the flow model to include the thickening of the slab perpendicular to its plane.

RESIDUAL SPHERE IMAGES

Four deep focus earthquakes were selected for this study, two from the Kuril-Kamchatka arc and two from the Mariana arc (Table 1). Figure 1 shows their epicenters on a map of the seismic zones and their hypocenters projected on cross sections through the three-dimensional slab models derived by Creager and Jordan [1986]. The slab models were constructed to satisfy the residual sphere data from these and other events. Our method for constructing residual sphere images is essentially identical to that employed by Creager and Jordan [1986], and we give only an abbreviated description here. The data are P wave arrival times compiled by the International Seismological Centre (ISC) with ray paths bottoming between 800 km (A > 30°) and the core-mantle boundary (A ≤ 100°). Residuals relative to the Herrin et al. [1968] radial Earth model are corrected for three types of large-scale aspherical heterogeneity: hydrostatic ellipticity [Dziewonski and Gilbert, 1976], mean station anomalies [Dziewonski and Anderson, 1983], and the L0.256 lower mantle structure [Dziewonski, 1984]. Each observation is assigned a variance that is the sum of two parts, the square of a reading error and Dziewonski and Anderson's [1983] single-observation variance about the mean station anomaly. Theoretical times corresponding to the same station set are generated by tracing rays through a three-dimensional slab model, assuming that the hypocenter is located at the ISC depth in the cold thermal core of the slab.

The residual sphere is a plot of residuals, either the observed or theoretical, on the lower focal hemisphere [Davies and McKenzie, 1969]. Two operations are applied in an identical manner to both the observed and theoretical residual spheres. The first eliminates any travel time anomalies that can be parameterized as a hypocentral relocation or origin time shift; formally, this relocation procedure is a projection of the N-dimensional residual vectors into an (N−4)-dimensional subspace orthogonal to the discretized and weighted residual sphere harmonics of zeroth and first degree. The second filters the residual spheres by applying an operator whose spectrum tapers smoothly to zero for harmonics of high angular degree. Applying this smoothing operation to the observed times increases the signal-to-noise ratio of near-source heterogeneity by preferentially damping random observational errors and heterogeneities far from the source, whose spectra on the residual sphere are shifted to high wave numbers. Applying it to the theoretical times accounts for the bias in the spectrum of near-source heterogeneity introduced by smoothing the data.

As a measure of the fit of the model to the observations, we compute the variance reduction factor

$$ R = 1 - \frac{|| \Delta t_d - \Delta t_m ||^2}{|| \Delta t_d ||^2} $$

where $\Delta t_d$ is the error-weighted, relocated, and smoothed data
Fig. 1. Map on left is a mercator projection of the northwest Pacific showing the epicenters (solid dots) of events 1–4 modeled in this study. The Pacific plate boundary (barbed line), seismicity contours in kilometers (light lines), and absolute Pacific plate velocity (arrows) are also shown. Slab models on right are for the northern Kuril (top) and Mariana (bottom) arcs [from Creager and Jordan, 1986]. P wave velocities are contoured at 0.25 km/s, and ambient values are the radial model of Herrin et al. [1968]. Circles are projections of well-recorded ISC hypocenters in the vicinity of the profile; the stars indicate the locations of events 1–4.

residual vector and At_m is the corresponding model vector. The procedure for inverting the data is to maximize R by adjusting the parameters of the slab model, which include the local strike, dip, and depth extent, as well as the thermal coefficient dP/dT. The latter specifies the (linear) relationship between temperature and compressional velocity; in the study by Creager and Jordan [1986], it was determined to be -0.5 ± 0.1 m/s/°K by inverting the data for intermediate focus earthquakes, which are insensitive to how far the slab extends below the seismicity cutoff. This value is consistent with the lower mantle estimates of Hager et al. [1985] and various laboratory studies [Sumino, 1979; Sumino et al., 1983; Suzuki and Anderson, 1983; Suzuki et al., 1983].

Figure 2 compares the observed residual spheres for the four deep focus earthquakes used here with the theoretical residual spheres calculated from the best fitting slab models of Creager and Jordan [1986]. The agreement is very good, both in the details of the pattern and the overall amplitude. The best variance reduction (R = 88%) is obtained for event 1, an earthquake of intermediate magnitude (m_l = 6.0) having sharp, clean arrivals [Jordan, 1977]. The worst (R = 66%) is for the smaller magnitude (m_l = 5.6) event 3. The three-dimensional models used in the ray tracing were obtained by rotating the cross sections shown in Figure 1 about arc poles at 60°N, 130°E for Kuril-Kamchatka and at 20°N, 90°E for the Marianas. The Mariana slab is essentially vertical, while the Kuril-Kamchatka slab displays a change of dip from about δ = 52° above 500 km depth to δ = 72° below 600 km [Veith, 1974; Jordan, 1977; Creager and Jordan, 1986]. Both slab models have velocity anomalies persisting to about 1400 km depth. However, as discussed in detail by Creager and Jordan [1984, 1986], the residual spheres for the deepest foci "saturate"
C) The theoretical patterns, so that extensions below about 1000 km cannot be resolved by the data. Decreasing slab length above this depth does significantly degrade the data model fit, leading us to conclude that the velocity anomaly must continue to at least 1000 km.

MODELING SLAB DEFORMATION

Creager and Jordan [1986] generated the slab structures shown in Figure 1 from a two-dimensional, entrained flow model where particle paths are constrained to be parallel to a specified surface and the shear normal to this surface is minimized. An error function thermal profile corresponding to the age of the lithosphere is used to initialize the finite difference algorithm, which computes the temperature structure of the descending lithosphere by stepwise alternating between mass transport and conductive heating [Toksoz et al., 1973]. In such a model the downdip velocity of the slab is constant at the plate tectonic rate, and the temperature anomaly is transported outward from the cold core of the slab by conduction alone; hence lateral thickening occurs very slowly. During the 10 m.y. required for the 150 Ma lithosphere to descend from 200 to 1200 km depth in the Mariana model, for example, thermal diffusion increases slab width by only 20%. This sort of flow obviously cannot be maintained to arbitrary depth, so at some point the lateral advection of thermal anomalies will become important. The form this deformation takes is unclear. Davies [1980] catalogues a number of scenarios: buckling, breaking, and imbricating, fragmenting into a shower of small pieces, and so forth. We do not choose among any of these mechanisms; instead, we assume that the details of the deformation are averaged out along the ray paths, so that the travel time data resolve the slab as a smoothly broadening and coherent whole.

We retain the description of subduction as a two-dimensional steady state flow in the plane perpendicular to the slab, and we solve the thermal conduction equation as before. To broaden the slab, however, we construct a new kinematic flow model where the vertical component of particle velocity decreases with depth throughout a given depth range. The particular deformation model adopted in this study is specified by three parameters: \( z_1 \), the depth at which the broadening begins; \( \zeta \), the depth interval over which it occurs; and \( \gamma \), the ratio of the vertical particle velocity \( v_1 \) at \( z_1 \) to the vertical velocity \( v_2 \) at \( z_2 = z_1 + \zeta \). Mass is conserved and the boundary conditions are such that the ratio of the horizontal velocity \( u \) to the vertical velocity \( v \) is the same at \( z_1 \) and \( z_2 \); that is, the slab dip \( \delta = \arctan(v/u) \) is the same above and below the region of deformation. The particle paths are lines of constant \( F(x,z) \), where

\[
2F(x,z) = -[(1 - \gamma^2) \cos(n(z-z_1) + \gamma - 1 + 1) v_1 x + [(1 - \gamma)(2z_1 - z)/(\zeta + 2)] u_1 z \tag{2}
\]

The horizontal and vertical components of particle velocity are given by the spatial derivatives of this stream function:

\[
u = \partial F/\partial x \quad v = -\partial F/\partial x
\]

This flow field has the property that the slab width smoothly increases (and \( v \) decreases) by a factor of \( \gamma \) over the depth range \( \zeta \), so that \( \gamma = 1 \) gives no distortion and \( \gamma = 2 \) yields a slab twice as

---

Fig. 2. Observed (left) and theoretical (right) residual spheres for events 1–4. Residual spheres are equal-area, lower hemisphere projections. Outer perimeter corresponds to the locus of rays with takeoff angles of 60° from vertical; data outside this range are not shown. Circles represent negative residuals; crosses are positive; the size of the symbol is proportional to residual, with 1-s residuals shown in upper box. Slab models are shown in Figure 1 and correspond to \( \gamma = 1 \) (no advective thickening). Bold lines running through the theoretical residual spheres represent planes tangent to the slab models from which they were calculated. In the Kurils the local slab strike is 31° and the dip below 500 km is 72°; the Mariana slab dips at 90° below 200 km, and its strike is 327°. Numbers in upper right are ISC depths.
thick at \( z_2 \) than at \( z_1 \). Figure 3 illustrates the latter case for the vertically plunging Mariana slab, assuming \( z_1 = 580 \text{ km} \) and \( \zeta = 150 \text{ km} \). Figure 4 displays the isotherms for the Kuril-Kamchatka slab when \( z_1 = 580 \text{ km} \), \( \zeta = 450 \text{ km} \), and \( \gamma \) is increased from 1 to 2 to 4, and when \( z_1 = 580 \text{ km} \), \( \gamma = 3 \), and \( \zeta \) ranges from 150 to 750 km.

**PARTICLE PATHS**

\[ \gamma = x_2 / x_1 \]

\[ x_2 \]

\[ x_1 \]

\[ v_1 \]

\[ v_2 \]

\[ \zeta \]

\[ u_1 \]

\[ u_2 \]

Fig. 3. Lines of constant \( \gamma \) for the subduction flow model of equation (2). Case illustrated is the vertical Mariana slab, for a thickening factor \( \gamma = 2 \) over a depth range \( \zeta = 150 \text{ km} \). The only streamlines shown are for diverging particle paths inside a nominal initial slab width \( x_1 = 100 \text{ km} \) at \( z_2 = 580 \text{ km} \).

**CONSTRAINTS ON SLAB WIDTH AND PENETRATION DEPTH**

In Figure 4 and our subsequent calculations, \( \gamma = 1 \) corresponds to Creager and Jordan’s [1986] best fitting models (Figure 1). To investigate the slab widths allowed by the travel time data, we have computed the variance ratios \( R \) of events 1–4 for a series of models where \( z_1 \) is fixed at 580 km, \( \zeta \) takes on values of 0, 150, 450, and 750 km, and \( \gamma \) is incremented from 1 to 5. Strike and dip were varied, but we found that the values obtained by Creager and Jordan [1986] for \( \gamma = 1 \) were close to the optimal over the entire range of parameters with the exception of \( \gamma \geq 3 \) for \( \zeta = 750 \text{ km} \) in the Kurils where shallower slab dips produced better fits to the data. The level of penetration was maintained at 1400 km, well below the saturation depths of 900–1100 km for events 1–4. For each thermal model the velocity structure used to trace the rays was determined from the thermal coefficient of velocity \( \partial v / \partial T \) that gave the best fit after hypocenter relocation and data smoothing, regardless of its value. We will return to the question of how slab width affects the estimates of \( \partial v / \partial T \) and saturation depth in our discussion below.

**Kuril-Kamchatka slab.** Three-dimensional ray bending tends to reduce the sensitivity of the theoretical times to increasing \( \gamma \). If the ray paths were not significantly perturbed by slab heterogeneity, we would expect the band of negative residuals associated with ray paths in high-velocity slab material to be wider for a thicker (bigger \( \gamma \)) slab. However, because narrower models have larger velocity gradients normal to the slab, rays tend to travel greater distances along the slab core and are bent more sharply as they leave the slab plane. Thus rays to a station not directly along the slab plane will accumulate nearly as much travel time anomaly for a narrow slab as for a wide one. Figure 5 illustrates this effect for the Kuril-Kamchatka slab by superimposing theoretical travel times calculated with and without ray bending for a thin (\( \gamma = 1 \)) and a thick (\( \gamma = 4 \)) model as a function of azimuth (\( \zeta \) was fixed at 150 km). When the times are calculated with no ray bending (top panel), the thick slab produces negative anomalies that are wider by a factor of 3 than the thin slab. Including three-dimensional ray bending (bottom panel) affects the width of the negative troughs for the wide slab very little but substantially widens those for the narrow slab. Engdahl and Gubbins [1987] make a similar observation in their study of central Aleutian slab structure. They conclude that if ray bending effects are not included in the inversion of travel time data, slab width is overestimated.

Contours of variance reduction as a function of \( \gamma \) and \( \zeta \) are shown for the two Kuril-Kamchatka earthquakes in Figure 6. The undistorted slab model gives the best variance reduction, obtaining \( R = 88\% \) and \( 71\% \) for events 1 and 2, respectively. For constant \( \zeta \) the fit degrades monotonically with increasing \( \gamma \). We have shaded in Figure 5 the values of \( R \) outside the 90% confidence level of acceptability. The critical value was calculated from a chi-square distribution with 10 degrees of freedom, the number of degrees of freedom in the residual sphere smoothing filter. According to this statistical criterion, slab models with thickening factors less than about 3 are acceptable, but those with \( \gamma \geq 4 \) provide significantly poorer fits to the data for \( \zeta = 0, 150 \), and 450 km. The \( \zeta = 750 \text{ km} \) models are consistent with the data up to \( \gamma \) of at least 5. This insensitivity to increasing \( \gamma \) expresses the simple fact that much of the widening in the \( \zeta = 750 \text{ km} \) models occurs below 1000–1200 km, the depth at which the residual spheres from even the deepest events "saturate" with respect to slab length.

Figure 6 also indicates that for fixed \( \gamma \), \( R \) is lowest at \( \zeta = 450 \text{ km} \), marginally improving at smaller and larger values. Once the travel time variations associated with unconstrained hypocenter/origin time perturbations have been projected out of the residual spheres, the data are not sensitive to absolute values of \( v_\rho \), but only to velocity gradients in the ray cone below the source [Creager and Jordan, 1984, 1986]. In the case of the Kuril-Kamchatka events the gradients normal to the ray paths are minimized for \( \zeta = 450 \text{ km} \); in cross section the slab flares like the ray cone (cf. Figure 4). This degrades the match to the residual sphere anomaly patterns.

Figure 7 compares the smoothed travel time residuals for event 1 as a function of azimuth, with similar plots computed by subtracting from the observed residual sphere the theoretical patterns for three models where \( \zeta = 150 \text{ km} \) and \( \gamma = 1, 3, \) and 5. The quality of the fit provided by the undistorted slab is evident, obtaining \( R = 88\% \) and \( 71\% \) for events 1 and 2, respectively. For constant \( \zeta \) the fit degrades monotonically with increasing \( \gamma \). We have shaded in Figure 5 the values of \( R \) outside the 90% confidence level of acceptability. The critical value was calculated from a chi-square distribution with 10 degrees of freedom, the number of degrees of freedom in the residual sphere smoothing filter. According to this statistical criterion, slab models with thickening factors less than about 3 are acceptable, but those with \( \gamma \geq 4 \) provide significantly poorer fits to the data for \( \zeta = 0, 150 \), and 450 km. The \( \zeta = 750 \text{ km} \) models are consistent with the data up to \( \gamma \) of at least 5. This insensitivity to increasing \( \gamma \) expresses the simple fact that much of the widening in the \( \zeta = 750 \text{ km} \) models occurs below 1000–1200 km, the depth at which the residual spheres from even the deepest events "saturate" with respect to slab length.

**PARTICLE PATHS**

\[ \gamma = x_2 / x_1 \]

\[ x_2 \]

\[ x_1 \]

\[ v_1 \]

\[ v_2 \]

\[ \zeta \]

\[ u_1 \]

\[ u_2 \]
Fig. 4. Isotherms for Kuril slab models obtained by the finite difference calculations described in the text. A divergent flow with thickening factor $\gamma$ is imposed between $z_1 = 580$ km and $z_2 = z_1 + \zeta$; streamlines for this flow conform to equation (2). Results for $\gamma = 1, 2, 4$ with $\zeta$ held fixed at 150 km are on left; those for $\zeta = 150, 450, 750$ km with $\gamma = 3$ are on right. Contour interval is 200°C; depth scale is in kilometers with no vertical exaggeration.

about $-30^\circ$ in the fourth frame of Figure 7, which is not seen for $\gamma = 1$ and 3.

Varying slab width produces a relatively subtle change in the pattern of residuals on the focal sphere compared to the simple pattern rotations produced by perturbing strike or dip. Figure 8, in a projection identical to Figure 7, shows the effect of decreasing the dip and strike of the reference slab model by $10^\circ$ for event 1. These perturbations decrease $R$ to 54 and 59%, respectively, well outside the 90% confidence region. If we assume $\gamma = 1$, the strike and dip are resolved to $\pm 5^\circ$ for this earthquake. The orientation parameters are somewhat less resolved for larger value of $\gamma$; but the best estimates of strike and dip remain pretty much the same for $\gamma$ values less than 5 and $\zeta$ values less than 750 km.

To examine the constraints on slab length, we plot in Figure 9 the event 2 variance reduction contours and $\partial v / \partial T$ values for slab models of various penetration depths and thickening factors. The best variance reduction is obtained for small $\gamma$ and long slabs. The data are quite insensitive to increasing slab length below about 1100 km, the saturation depth for these events, and variance reduction gets worse with increasing $\gamma$. When the penetration depth is less than 950 km, variance reduction is unacceptable at the 90% confidence level regardless of the value of $\gamma$. At depths less than about 900 km, $\partial v / \partial T$ exceeds $-1.0$ m/s/°K, twice Creager and Jordan's [1986] preferred value of $-0.5 \pm 0.1$ m/s/°K. The latter is consistent with the estimate of $-0.4$ m/s/°K obtained by combining Hager et al.'s [1985] value for the large-scale $v_p - \rho$ variation in the lower mantle (4 km s$^{-1}$/Mg m$^{-3}$) and Jeanlou and Thompson's [1983] thermal expansion coefficient ($2.5 \times 10^{-5}$ °K$^{-1}$), as well as a number of high-temperature laboratory studies [Sumino, 1979; Sumino et al., 1983; Suzuki and Anderson, 1983; Suzuki et al., 1983]. Given these constraints on $\partial v / \partial T$ from several data sources, $\partial v / \partial T$ values much in excess of $-1.0$ m/s/°K are probably unacceptable.

Mariana slab. As in the case of the Kurils, the undistorted ($\gamma = 1$) Mariana slab model gives the best variance reduction for the two earthquakes considered, and the fit at constant $\zeta$ degrades with increasing $\gamma$. However, relative to the Kuril-Kamchatka data, residual spheres from the Marianas place weaker constraints on acceptable $\gamma$: $R$ values for events 3 and 4 both lie within the 90% confidence region for thickening factors as large as 5 (Figure 10).
This lower sensitivity is due primarily to three geometrical differences: (1) Because the Marianas are farther than the Kurils from the large station concentrations in western Europe and North America, the number of data points available for the individual events is substantially fewer (~180 versus 300). (2) In contrast to the Kurils where the slab dips at 72° and the strong downdip gradients in travel time are well sampled by European stations, the orientation and vertical geometry of the Mariana slab lead to smaller amplitudes and gradients on the residual sphere (Figure 2). (3) As in the case of the Kurils, the differences in theoretical times generated by models of differing thicknesses are small due to the ray bending effects discussed previously; however, because of the geometrical weakness associated with the station distribution and vertical Mariana slab, these can more readily be mapped into variations in depth and origin times and are therefore projected out in our relocation procedure.

In an attempt to obtain better constraints on γ, we calculated the variance reduction for the same suites of slab models except, rather than allowing an arbitrary relocation of the hypocenter, we fixed the event depth at various values, including that reported by the ISC. The resulting differences in residuals were largely compensated by shifts in origin times, and the dependence of variance reduction on γ was not substantially altered.

Despite their insensitivity to slab thickness, the Marianas data provide a robust lower bound on penetration depth (Figure 11). As in the Kurils, the data are relatively insensitive to changes in slab length at depths below about 1000 km, but shortening the slab above this depth produces a rapid decline in the quality of the fit, and a sharp increase in \( \partial \gamma / \partial T \). The increase in variance...
Fig. 8. Plots of corrected, relocated, and smoothed travel time residuals as a function of residual sphere azimuth for Kuril event 1. The upper and lower panels are identical to those of Figure 7; the former corresponds to the best fitting slab model of Figure 1, which has a strike of 31° and a dip of 72°. The middle panels correspond to models where \( \gamma = 1 \) (no advective thickening) and the strike and dip vary from their best fitting values; in the second panel the dip is decreased to 62°, and in the third the strike is decreased to 21°. These variations produce unacceptable fits to the data.

Fig. 9. (Top) Contours of variance reduction calculated according to equation (1) for Kuril event 2. Vertical axis is the thickening factor \( \gamma \); horizontal axis is slab penetration depth. \( \zeta = 150 \) km for all models. Contour interval is 2%. The shaded area corresponds to the models whose variance reductions are significantly less at the 90% confidence level than a model with 10 degrees of freedom, the criterion we use to indicate unacceptable fits to the data. For models extending below 900–1000 km variance reduction is good, but it falls off for models ending much above this depth. (Bottom) Contours of \( \partial v/\partial T \) obtained from the least squares fit of the model residual spheres to the data. \( \partial v/\partial T \) increases rather dramatically as slab length shortens above 90 km; the shaded area corresponds to values of \( \partial v/\partial T \) greater than \(-1.0 \) m/s/kK, considered to be unacceptably large on the basis of other data (see text).

Fig. 10. Contours of variance reduction calculated according to equation (1) for Mariana events 3 (left panel) and 4 (right panel). Vertical axis is the depth interval \( \xi \); the horizontal axis is the thickening factor \( \gamma \). Contour interval is 2%. Comparison with Figure 6 shows that the constraints on \( \gamma \) are much weaker than in the Kurils; no points on this plot have variance reductions that are significantly worse at the 90% confidence level than a model with 10 degrees of freedom. Hence for these events, \( \gamma \) of at least 5 are acceptable.

**DISCUSSION**

The results of this study confirm two conclusions of Creager and Jordan [1986]: (1) the lower bounds established by residual sphere data on the depth extent of the high-velocity structures underlying deep focus zones depend only weakly on the cross-strike width of these zones, and (2) structures considerably wider than simple thermal plate models can satisfy the residual sphere data. In particular, significant advective thickening of the slab in and below the transition zone, suggested by the mechanisms of deep focus earthquakes [e.g., Giardini and Woodhouse, 1984] and models of subduction flow from a low-viscosity upper mantle to a high-viscosity lower mantle [Hager, 1986], is compatible with the travel time data.

The residual sphere images from four deep focus events in the Kuril-Kamchatka and Mariana subduction zones have been used
to place an upper bound on the amount of advective thickening. In terms of our particular parameterization of slab deformation (equation (2)), a thickening factor \( \gamma \) as great as 5 or more can be accommodated in the Marianas. In the Kurils, however, the residual sphere data limit \( \gamma \) to be less than 3 or 4 if the deformation occurs largely above a depth of about 1000 km. We should be quick to point out that we cannot say from our data that such advective broadening is really a feature of subduction flow; in fact, among the models examined here, simple slab penetration (\( \gamma = 1 \)) yields the best fit to the residual spheres of Figure 2. But our calculations show there is no discrepancy between the travel time data and the waveform observations and modeling discussed, for example, by Vidale [1987].

Of course, the lateral advection of cold slab material must become important somewhere above the core-mantle boundary, and constraining the nature of this plate tectonic return flow is an important task for structural seismology. Some insights are already available from global tomographic mapping and high-resolution regional studies of lower mantle heterogeneity. Jordan and Lynn [1974] discovered a high-velocity anomaly for both P and S waves at depths of 800–1400 km in the lower mantle beneath the Caribbean Sea, which they associated with subduction flow beneath the western margin of North America (see also Jordan [1975]). Lay [1983] presented additional evidence for the Caribbean anomaly, and Grand [1987] has obtained a spectacular three-dimensional image of this feature by a tomographic inversion of multiple-S travel times. In his maps the Caribbean anomaly extends northward beneath the central United States and downward to at least 1700 km. It dips eastward and is plausibly explained as the accumulation of cold slab material subducted into the lower mantle during the opening of the Atlantic. The analysis of regional travel times by Lay [1983] and the global inversions of body wave data by Dziewonski and Woodhouse [1987] indicate that this high-velocity region continues around the periphery of the Pacific into Eurasia.

At a depth of 1200 km the peak amplitude of the Caribbean anomaly measured from Grand's [1987] tomographic maps is about 1% and its cross-strike width is a little less than 1000 km. Because of the smoothing Grand used in his inversion, however, the actual anomaly may be narrower with a greater peak amplitude (S. Grand, personal communication, 1988). Unsmoothed plots of travel time anomalies by Jordan and Lynn [1974] suggest a total width of about 500 km, which is consistent with the advective thickening models shown in Figure 4.

The horizontal resolution of the global tomographic inversions of Dziewonski and Woodhouse [1987] is about 2000 km in the lower mantle, so their images have been even more severely low-pass filtered. Nevertheless, as these authors have pointed out, the fact that subduction zone anomalies are more evident in the lower mantle than above the seismicity cutoff is an additional argument for some advective thickening. For example, Dziewonski and Woodhouse's [1987] P wave map at 1200-km depth (their Figure 8) has a broad (~3000 km) positive feature behind the Kuril-Kamchatka arc with peak perturbation of about 0.25%. Integrating along a horizontal line perpendicular to the strike of the arc yields a total travel time anomaly of roughly 0.5 s, a value larger than would be expected from the slab penetration model of Figure 1 whose total anomaly of ~0.3 s is concentrated in a much narrower zone which low wave number models could not completely resolve. At 1200 km depth, however, the \( \gamma = 3 \) models displayed in Figure 4 have significantly larger cross-strike anomalies (~0.8 s) spread out over a region that is nearly half as wide as in the global models. Filtering such a structure to low wave numbers could explain the tomographic results. The large volume of slab material in the lower mantle that these broad anomalies imply is consistent with the tectonic history of the western Pacific margin [Hilde et al., 1976]; subduction of slabs along this margin at a rate of 80 mm/yr for at least the last 45 m.y. (the approximate age of the bend in the Emperor-Hawaiian hotspot chain) thrusts enough lithosphere into the upper half of the lower mantle to fill a volume corresponding to \( \gamma = 3 \).

Advective thickening models such as those shown in Figure 4 require strain rates in the deep slab which are significantly greater than the seismic strain rates implied by earthquake moments observed below 500 km. To show this, we approximate the rate of seismic moment release in a way similar to that of Richter [1979]. For every event in the 1964–1984 ISC catalog between 500 and 700 km depth with 5.0 \( \leq M_w \leq 6.0 \) in a given subduction zone, we calculate its seismic moment assuming a linear relation between \( M_w \) and the logarithm of \( M_o \) (Gutenberg and Richter, 1954, 1956; Hanks and Kanamori, 1979); we then sum the moments over this magnitude range, scale the total to an upper moment cutoff of \( 2.0 \times 10^{22} \) N m (\( 2.0 \times 10^{29} \) dyn cm, corresponding to \( m_b = 8 \)) assuming a \( b \) value of unity, and divide by the time interval to get an average rate of seismic moment release, \( \dot{M}_{\text{seismic}} = (\dot{M} = \partial M_o/\partial t) \). In the Kuril and Mariana arcs the rates of seismic moment release are relatively low: \( \dot{M}_{\text{seismic}} = 3.7 \times 10^{19} \) N m/yr and \( 5.3 \times 10^{19} \) N m/yr, respectively. On the other hand, Tonga, the Earth's most active deep seismic zone, yields \( \dot{M}_{\text{seismic}} = 2.7 \times 10^{20} \) N m/yr, a value consistent with Richter's [1979] estimate. These rates are sensitive to the upper moment cutoff; the largest deep focus earthquake documented by Abe and
Kanamori [1979] has an estimated moment of $2.2 \times 10^{21}$ N m, about an order of magnitude less than our assumed cutoff value. Therefore the values of $M_{\text{seismic}}$ given above are likely to be overestimates, unless the largest events are significantly undersampled in the 70-year interval considered by Abe and Kanamori [1979].

The strain rate function for the flow described in equation (2) can be determined from its spatial derivatives. The integral of this function over slab volume scales to an equivalent rate of moment release, $\dot{M}_{\text{flow}}$. It depends linearly on $x_1$ and the along-strike extent of the slab and, over the parameter ranges $2 \leq \gamma \leq 4$ and $1 \text{ km} \leq x_1 \leq 100 \text{ km}$, roughly linearly on $\gamma$ and $x_1$ (the latter being the initial width of the deforming volume at $z_1$; see Figure 3). The integral asymptotically decreases to a constant value with increasing $\zeta$; we assume $\zeta = 200 \text{ km}$, which is consistent with the deformation occurring in the deep seismic zone. We perform this calculation for $x_1 = 25 \text{ km}$ which corresponds roughly to the width of the deep seismicity (except in parts of Tonga where it underestimates the seismic zone) and $x_1 = 100 \text{ km}$ which represents total undeformed slab thickness. $\dot{M}_{\text{flow}}$ values for $x_1 = 25 \text{ km}$ therefore reflect the strain rate due to advective deformation of the seismically active portion of the deep slab, whereas $\dot{M}_{\text{flow}}$ values for $x_1 = 100 \text{ km}$ include the warmer, aseismic outer regions. In Tonga, $M_{\text{seismic}}$ for $\gamma = 3$ is only 40% of $\dot{M}_{\text{flow}}$ if $x_1 = 25 \text{ km}$ and 10% of $\dot{M}_{\text{flow}}$ if $x_1 = 100 \text{ km}$. That is, even in the most seismically active deep slab, the observed moment release can account for less than half of the strain predicted in the seismic core of the slab and less than 10% of the total slab deformation if the thickening factor is 3 or more. In the two subduction zones constrained by the residual sphere data presented in this paper, the discrepancy is much larger: for $x_1 = 25 \text{ km}$, a thickening factor of 3 implies $M_{\text{seismic}}$ is about 5% of $\dot{M}_{\text{flow}}$ for the Kurils and less than 1% for the Marianas. We conclude, therefore, that if substantial advective thickening does occur, the deformation must be largely aseismic.

CONCLUSIONS

While descending lithospheric slabs appear to penetrate the 650-km discontinuity, they may experience considerable advective thickening perpendicular to the arc at or below that depth. Deep event travel times are best fit by undeformed models, but a twofold increase in slab thickness fits all events nearly as well. Although the Marianas times allow advective thickening factors of 5 or more, factors larger than 3 are not consistent with the Kuril data. For all models examined, the data require that slab material extends to depths of 900–1000 km. The insensitivity of the travel time data to slab width, relative to the tight constraints that they place on slab strike, dip, and depth extent, may be explained by the combined effects of three-dimensional ray bending and the geometrical weakness associated with the station distribution and the vertical Mariana slab.

A subducting lithosphere which advectively thickens upon its entry into the lower mantle is consistent with seismicity and moment tensor data which point to compressive stresses in the deep slab. An increase in viscosity between the upper and lower mantle could be the primary dynamical control on such a flow, and an intriguing question for future research is to delineate what range of viscosity structures and slab rheologies are consistent with the above constraints on deep slab thickening. If such deformation does occur, its mechanisms are undoubtedly much more complicated than the simple two-dimensional kinematical flow models used here. Our models assume that the details of the deformation (whether the slab buckles, imbricates, fragments, etc.) are averaged out along the ray paths, and hence we constrain only the scale and not the mode of slab thickening. Comparison of observed and predicted rates of seismic moment release suggests that any substantial advective thickening is largely aseismic, and residual sphere modeling in the Tonga arc, where deep events are distributed both normal and parallel to the arc, holds promise for further elucidation of how the deformation of deep slabs may occur.

Acknowledgments. We thank V. Cormier, A. Dziewonski, S. Grand, B. Hager, J. Vidale, E. R. Engdahl, J. Woodhouse and E. Robinson for preprints and helpful discussions, and we applaud D. Krowitz for keeping Bertha alive. This research was sponsored by the National Science Foundation under grant EAR 8607340.

REFERENCES


—, Geophysics Program AK-50, University of Washington, Seattle, WA 98195.


(Received September 29, 1987; revised January 10, 1988; accepted November 21, 1987.)