A Water-Rich Transition Zone Beneath the Eastern United States and Gulf of Mexico from Multiple ScS Reverberations

Anna M. Courtier* and Justin Revenaugh

Department of Geology and Geophysics, University of Minnesota, Minneapolis, Minnesota

We examine mantle discontinuities beneath the United States and Gulf of Mexico using multiple ScS reverberations from earthquakes in Central and South America captured by 65 broadband and long-period seismometers across the United States. The depths of discontinuities and the impedance contrasts across them were estimated using a hierarchical waveform inversion and stacking method. The path-averaged depth of the 410-km discontinuity varies moderately across the study area and is particularly shallow (~395 km) beneath the eastern United States. Topography on the 660-km discontinuity is more subdued and is close to the global mean depth. The 520-km discontinuity is seen consistently across the study area, though both the depth and the impedance contrast of the discontinuity vary significantly. Corridors in the eastern United States and Gulf of Mexico have extremely strong 520-km discontinuities relative to the corresponding 410-km and 660-km discontinuities. We attribute the shallow 410-km and strong 520-km discontinuities beneath the eastern United States and Gulf of Mexico to a locally water-rich transition zone.

INTRODUCTION

Water must be present to some extent in the Earth’s mantle [e.g. Bell and Rossman, 1992], though the total mass of water and its distribution in the interior are largely unknown [e.g. Drake and Righter, 2002; Hirschmann et al., 2005]. The water content of the mantle is largely governed by the amount of hydrogen that partitioned into the accreting core and mantle during Earth formation, the amount of water that subsequently degassed to form the hydrosphere, and the amount that is recycled into the interior through subduction [e.g. Ahrens, 1989; Williams and Hemley, 2001]. Each of the major mantle minerals has the ability to incorporate at least trace amounts of water in its structure [e.g. Bell and Rossman, 1992; Kohlstedt et al., 1996; Bolfan-Casanova et al., 2000; Murakami et al., 2002; Bolfan-Casanova et al., 2003; Litaso et al., 2003; Bolfan-Casanova, 2005], allowing for the equivalent of half of the world’s oceans in a “dry” mantle [Hirschmann et al., 2005] and potentially much more [Smyth, 1987; Bell and Rossman, 1992]. Hydrogen may be present as point defects, as structurally bound hydroxyl [(OH)\(^-\)], or as molecular water (H\(_2\)O), each of which is colloquially described as “water” in the literature.

The distribution of water in the mantle can be constrained by considering seismic observations in conjunction with experimental constraints on water in mantle minerals. A number of authors have examined the solubility of water in mantle minerals [e.g. Bell and Rossman, 1992; Kohlstedt et al., 1996; Ingrin and Skogby, 2000; Bolfan-Casanova et al., 2000; Bolfan-Casanova et al., 2003; Bolfan-Casanova, 2005] and the pressure-temperature stability fields of minerals containing water [e.g. Irifune et al., 1998; Higo et al., 2001; Ohtani et al., 2001; Smyth and Frost, 2002]. Although all of the most abundant mantle minerals are capable of storing water in their structures, for some the concentration of water may reach only the parts per million level.
Water solubility measurements indicate that transition zone minerals can incorporate considerably more water than both the remaining primary upper mantle minerals and primary lower mantle minerals [e.g. Smyth, 1987; Kohlstedt et al., 1996; Bolfan-Casanova et al., 2000]. While it is highly unlikely that the entire mantle is saturated with water, the transition zone may be a significant reservoir for water due to its enhanced water solubility.

The transition zone seismic discontinuities, herein referred to as the 410-km, 520-km, and 660-km discontinuities, correspond to the phase transition from olivine to wadsleyite ($\beta$-$\text{Mg}_2\text{SiO}_4$), the transition from wadsleyite to ringwoodite ($\gamma$-$\text{Mg}_2\text{SiO}_4$), and the dissociation of ringwoodite to perovskite-type ($\text{Mg,Fe}_2\text{SiO}_4$), and magnesiowüstite-(Mg,Fe)O, respectively (see Helffrich [2000] for a review). These three reactions have pressure-temperature stability fields that vary between hydrous and anhydrous conditions. The 410-km discontinuity migrates to shallower depths in the presence of water [Smyth and Frost, 2002; Chen et al., 2002; Komabayashi et al., 2005; Komabayashi et al., this volume], whereas the 660-km discontinuity may deepen [Higo et al., 2001]. The presence of water also can broaden or sharpen mantle velocity transitions [e.g. Akaoji et al., 1989; Wood, 1995]. Transition thicknesses under anhydrous conditions are 9-18 km for the 410-km discontinuity [Akaoji et al., 1989], 30 km for the 520-km discontinuity [Akaoji et al., 1989], and 4 km or less for the 660-km discontinuity [Ito and Takahashi, 1989]. For under-saturated hydrous conditions, the 410-km and 660-km discontinuities broaden to thicknesses of up to 40 km [Wood, 1995; Helffrich and Wood, 1996; Smyth and Frost, 2002] and 13 km [Higo et al., 2001], respectively. The 520-km discontinuity sharpens in the presence of water, occurring over a thickness of less than 15 km [Inoue et al., 1998].

Seismic methods yield the most direct observations of the transition zone discontinuities. A global study by Flanagan and Shearer [1998] reported mean depths of 418 km, 515 km, and 660 km for the three discontinuities. The 520-km discontinuity is seen more frequently in areas where data density is high, and it typically has a much smaller apparent impedance contrast than either the 410-km or 660-km discontinuities. The poor signal-to-noise ratio of data sampling the low-contrast 520-km discontinuity almost certainly adds to the apparent variability in its depth. It is also occasionally observed as a two-part discontinuity, further adding to apparent depth variability [e.g. Deuss and Woodhouse, 2001]. Transition thicknesses influence the impedance contrasts measured by seismic methods. As a transition broadens, its effective reflection and conversion coefficients decrease [e.g. Richards, 1972], making the discontinuity more difficult to detect and, often, biasing downward the estimated impedance and/or velocity contrast. The contrasts at the 410-km and 660-km discontinuities are large enough that even if water broadens them, they should still be easily detectable with multiple $ScS$ reverberations. The 520-km discontinuity, which narrows in the presence of water, may appear stronger and be more easily detected in hydrous environments. If the transition zone is a significant reservoir for water in the mantle, the effect of that water should be apparent in seismic observations.

Several seismic studies have called upon water to explain anomalous observations of the transition zone and the overlying mantle. A layer of silicate partial melt was proposed by Revenaugh and Sipkin [1994] to explain an impedance decrease detected at an average of 80 km above the 410-km discontinuity beneath easternmost China and the Sea of Japan. The inferred partial-melt layer was detected with multiple $ScS$ reverberations and was seen consistently across the study area, though the thickness of the layer varied from ~50 to 100 km. Volatiles, principally water, were suggested as a catalyst for producing the melt.

Song et al. [2004] reported a very similar feature beneath the northwestern United States. The low velocity zone in this study was detected in triplicated shear-wave arrivals and extended from ~20 to 90 km above the 410-km discontinuity. It too is interpreted as a partial melt layer and linked to hydration from past subduction in the region. Van der Meijde et al. [2004] used receiver functions to infer the presence of up to 1000 ppm by weight of water near depths of 400 km beneath the Mediterranean, another region of subduction, which again may be the source of the water. A second receiver function study [Vinnik et al., 2003] reported a ~60-km thick low velocity layer on top of the 410-km discontinuity beneath the Arabian plate. This is interpreted as a water-rich layer underlying the dry continental root of the plate. The transition zone water filter model of Bercovici and Karato [2003] predicts occasional melt above the 410-km discontinuity as a byproduct of the difference in water solubilities of transition zone and overlying upper mantle minerals. The model predicts that as upwelling mantle material reaches the water-rich transition zone, it is able to incorporate water on the order of 1 – 2 wt% into its mineral structures. When this material passes through the 410-km transition, it becomes at least saturated, if not super-saturated, with water due to the lower solubility of water in olivine than in wadsleyite [e.g. Kohlstedt et al., 1996; Hirschmann et al., 2005]. Water saturation greatly depresses the melting temperature of the olivine and can cause localized partial melting above the transition zone [Inoue, 1994]. The resulting melt may be intermediate in density to olivine and wadsleyite [e.g. Stolper et al., 1981; Ohtani et al., 1995] and could pond above the 410-km discontinuity.

Broadening of the 410-km phase transition to thicknesses of 20 – 35 km is observed in the Mediterranean region [Van...
CoURTIER And REVENAUGH    3

der Meijde et al., 2003]. This can be explained by 0.07 wt % 
H2O (700 ppm) in the surrounding mantle. The 660-km tran-

tion also may broaden in the region, although the receiver 
function method used may not be able to detect the amount 
of broadening across the transition interval expected at that 
depth for the estimated water content. Van der Meijde et 

al. [2005] observed increases in transition zone thickness 
beneath the same region, which is consistent with either 
water or low temperature persisting throughout the depth 
rang of the transition zone. They further note that the veloc-

ity transition at the 520-km discontinuity may be as sharp 
as 20 km in the region. Several thousand kilometers to the 
north-northwest, weak conversions from the 660-km dis-

continuity beneath the North Sea could be the result of a 
broadened discontinuity [Helffrich et al., 2003]. A transition 
occuring over a 13-km interval could cause the observed 
behavior, consistent with the expected influence of water at 
660 km depth [Higo et al., 2001].

Low seismic velocities due to water in the mantle have 
been reported for compressional and shear waves in regions 
of both current and past subduction [e.g. Zhao, 2001; Nolet 
and Zielhuis, 1994]. In the mantle wedge, at depths to 400 
km, a low velocity zone often lies just above the subducting 
slab. Here the low velocities are attributed to water released 
from the slab as hydrous minerals dewater [Zhao, 2001]. 
Deeper in the mantle, Nolet and Zielhuis [1994] find evi-
dence of water in the upper mantle and transition zone in 
shear wave tomography of the mantle beneath the Russian 
platform. There, a low shear wave velocity anomaly extends 
from 300 to 500 km depth and follows the trend of the ancient 
Tornquist-Teisseyre subduction zone, which initiated with 
closure of the Tornquist Sea a maximum of ~450 million 
years ago [Bergstrom, 1990; Scotese and McKerrow, 1990]. 
The authors conclude that 85 million years of subduction 
have injected a significant amount of water into the transition 
zone, substantially lowering shear wave velocities.

In this context, we conducted an ScS reverberation study 
examining variability in discontinuity depths and impedance 
contrasts beneath the United States and Gulf of Mexico. 
Multiple ScS reverberations are useful tools for detecting 
regional variability that occurs due to changes in temperature 
or chemistry along mantle discontinuities. Past and present 
subduction of the Farallon and Juan de Fuca plates lead to 
slab-dewatering and the possibility for areas of water storage 
in the mantle beneath the study area.

DATA

We compiled a dataset of 130 long-period and broadband 
seismograms from fourteen intermediate depth events in 
Central and South America. Only events with magnitude m_b 
≥ 5.9 and depth z ≥ 95 km were considered. Data with low 
signal to noise ratios or with excessive apparent source com-

plexity were discarded. The events occurred between 1974 
and 2001 and were recorded at 65 stations from a variety of 
networks across the United States; the Digital World-Wide 
Standardized Seismograph Network (DWWSSN), United 
States National Seismic Network (USNSN), High-Gain 
Long-Period Network (HGLP), Lamont-Doherty Cooperative 
Seismographic Network (LCSN), IRIS Global Seismograph 
Network (GSN), Seismic Research Observatory (SRO), 
Pacific Northwest Regional Seismic Network (PNSN), Leo 
Brady Network (LB), TERRAscope (Southern California 
Seismic Network, SCSN), GEOSCOPE, Berkeley Digital 
Seismograph Network (BDSN), California Transect Network 
(CT), and ANZA Regional Network (ANZA). See Table 1 for 
source parameters of the events.

Seismograms were rotated, deconvolved to ground veloc-
ity, low-pass filtered, and decimated to a three-second 
sampling interval following Revenaugh and Jordan [1989]. 
Transverse component data were separated into eight source-
receiver paths based on geographic sampling (Figure 1) and 
considerations of data density. Six paths connect Central 
America and the United States. The remaining two connect 
South America and the United States. An example trace is 
shown in Figure 2.

Figure 1. Map of the study area showing earthquakes (circles) and 
seismic stations (diamonds). Black bars indicate particular geo-

graphic paths and are numbered from east to west according to the 
source region; Central America (C) or South America (S). The bars 
are schematic; actual sampling is much broader geographically.
METHOD

Zeroth- and first-order ScS reverberations (i.e. multiple ScS and sScS phases and similar arrivals once-reflected from discontinuities within the mantle; Figure 3) were modeled using the hierarchical waveform inversion method of Revenaugh and Jordan [1991a]. The only change to their method was the addition of a finite source duration, taken as the half-duration of the event as given by the Harvard Centroid Moment Tensor (CMT) catalog which scales directly with seismic moment. Source parameters for the two events prior to 1977 were taken from nearby events in the CMT catalog. Specifically, the July 18, 1983 (12.67° N, 87.18° W, 86 km depth, mb 6.0) earthquake was used for the March 6, 1974 event, and the June 25, 1980 (4.44° N, 75.78° W, 162 km depth, mb 6.0) earthquake was used for the May 19, 1976 event. Combinations of between two and four ScSn−sScSn phase pairs were modeled on each seismogram, with the number depending on the level of ambient noise and interference from major arc arrivals and phases from other earthquakes. Model parameters include the whole-mantle quality factor (QScS), crustal thickness, and whole-mantle travel time. QScS and the crustal thickness are regarded as “nuisance” parameters due to the geographic length of the corridors and complex variations in crustal structure expected along these primarily continental paths [e.g. Sipkin and Revenaugh, 1994].

Modeled zeroth-order reverberations were stripped from the data, leaving a residual signal consisting of first- and higher-order reverberations from discontinuities throughout the mantle [Revenaugh and Jordan, 1989], noise, and residual multiple ScS energy. An estimate of the mantle radial shear-wave reflection coefficient is obtained by 1D migration of the first-order reverberations as per Revenaugh and

<table>
<thead>
<tr>
<th>Date</th>
<th>Origin Time (UTC)</th>
<th>Latitude (Deg. N)</th>
<th>Longitude (Deg. W)</th>
<th>Depth (km)</th>
<th>mb</th>
<th>Paths</th>
</tr>
</thead>
<tbody>
<tr>
<td>March 6, 1974</td>
<td>01:40:26</td>
<td>12.29</td>
<td>86.39</td>
<td>110</td>
<td>6.1</td>
<td>C1 and C5</td>
</tr>
<tr>
<td>May 19, 1976</td>
<td>04:07:15</td>
<td>4.46</td>
<td>75.78</td>
<td>157</td>
<td>6.4</td>
<td>S1</td>
</tr>
<tr>
<td>June 22, 1979</td>
<td>06:30:54</td>
<td>17.00</td>
<td>94.61</td>
<td>107</td>
<td>6.3</td>
<td>C5</td>
</tr>
<tr>
<td>March 1, 1991</td>
<td>17:30:26</td>
<td>10.94</td>
<td>84.64</td>
<td>196</td>
<td>6.1</td>
<td>C5</td>
</tr>
<tr>
<td>June 12, 1993</td>
<td>11:15:07</td>
<td>13.25</td>
<td>87.53</td>
<td>217</td>
<td>5.9</td>
<td>C3, C5, and C6</td>
</tr>
<tr>
<td>April 10, 1994</td>
<td>17:36:57</td>
<td>14.72</td>
<td>92.00</td>
<td>100</td>
<td>6.0</td>
<td>C5 and C6</td>
</tr>
<tr>
<td>August 19, 1995</td>
<td>21:43:31</td>
<td>5.14</td>
<td>75.58</td>
<td>119</td>
<td>6.7</td>
<td>S1</td>
</tr>
<tr>
<td>October 21, 1995</td>
<td>02:38:57</td>
<td>16.84</td>
<td>93.47</td>
<td>159</td>
<td>7.2</td>
<td>C2, C3, C4, C5, and C6</td>
</tr>
<tr>
<td>December 31, 1996</td>
<td>12:41:42</td>
<td>15.83</td>
<td>92.97</td>
<td>99</td>
<td>6.4</td>
<td>C4, C5, and C6</td>
</tr>
<tr>
<td>September 2, 1997</td>
<td>12:13:22</td>
<td>3.85</td>
<td>75.75</td>
<td>198</td>
<td>6.8</td>
<td>S1</td>
</tr>
<tr>
<td>November 9, 1997</td>
<td>22:56:42</td>
<td>13.85</td>
<td>88.81</td>
<td>176</td>
<td>6.5</td>
<td>C1, C2, C3, C4, C5 and C6</td>
</tr>
<tr>
<td>December 11, 1997</td>
<td>07:56:28</td>
<td>3.93</td>
<td>75.79</td>
<td>177</td>
<td>6.4</td>
<td>S1 and S2</td>
</tr>
<tr>
<td>December 18, 1997</td>
<td>15:02:00</td>
<td>13.84</td>
<td>88.74</td>
<td>182</td>
<td>6.1</td>
<td>C1, C3, C4, C5, and C6</td>
</tr>
</tbody>
</table>
The general velocity model of Revenaugh and Jordan [1991a] was used for the migration of all source-receiver paths following simple velocity scaling to match the whole-mantle travel time specific to each region. The velocity used in migration must match the vertical ScS travel time estimated during the zeroth-order reverberation stripping, but is otherwise largely immaterial. For the near-vertical raypaths followed by multiple ScS and higher-order reverberations, move-out is minimal and nearly identical for all reasonable mantle models. What differs are the absolute depths, and we include the uncertainty in this mapping in the error bounds on the discontinuity depths. The stacking of all seismograms sampling an individual geographic corridor accomplished by migration greatly increases the signal-to-noise ratio of low-amplitude reflected shear waves. Unfortunately it also introduces artifacts and biases by collapsing 3D structure into a single spatial dimension, a point we address below. The reflection coefficient profiles are modeled using migrated synthetic seismograms in a graduated series of steps, resulting in a preferred reflection coefficient profile that contains the fewest reflectors necessary to accurately replicate the data profile.

There are several effects that can bias our reflection coefficient estimates; chief among these are along-path heterogeneity and anisotropy. Velocity and attenuation heterogeneity and crustal structure variability along-path are not well accounted for in the 1D migration methods used here. The result is discontinuity response functions (synthetic seismograms used as migration matched filters) that differ in waveform detail and arrival time from the data, lowering waveform cross-correlation and biasing downward the estimates of reflection coefficient. For the low-frequency waveforms we examine, these effects are not great, but they are present. Experiments with migration of synthetic data intentionally made to differ from the discontinuity response functions suggest biases on the order of 25% are possible. Along-path variability in the depth of individual discontinuities imposes an additional downward bias that scales with discontinuity topography, but which may also reach 25%. Anisotropy is present beneath much if not all of the study region, splitting the multiple ScS reverberations and increasing the waveform complexity of arrivals on the transverse components records we use. Because of inefficient reflection at the free surface and core-mantle boundary, SV-polarized multiple ScS phases are not well excited along these paths and are seldom observable in our dataset. (Their amplitude is typically only one-fourth of the amplitude of SH). As a result, the primary effect of anisotropy on the transverse component is a downward amplitude bias that increases roughly linearly with the number of passages through anisotropic depth intervals. This influences our estimates of $Q_{\text{ScS}}$ and reduces apparent reflection coefficients of mantle discontinuities. As with along-path heterogeneity, the effect should be near-equal for all discontinuities and, as such, is unlike the effect of discontinuity topography, which is reflector specific.

To minimize the impact of these sources of bias on our interpretations of mantle discontinuity structure, we will focus on reflection coefficient ratios, which cancel out common downward biases. In theory, the only remaining biases are due to variable degrees of discontinuity topography and transition width.

**RESULTS**

SH reflectivity estimates for each of the eight source-receiver paths are shown in Figure 4. Shallow upper mantle discontinuities are seen sporadically across the study area (Table 2). H (see Revenaugh and Jordan, 1991c for discontinuity nomenclature) may be present beneath the westernmost paths (C5 and C6) as well as path S2, but reflectivity peaks at this depth can also arise as artifacts of inadequate modeling of crustal impulse response, and these interpretations of H should be treated with caution. This is especially so in the case of path S2, where the putative H discontinuity has a
particularly high impedance contrast. Several discontinuities are seen in the depth range of L as well, though the impedance contrast is negative below the western United States and positive beneath the eastern United States. The easternmost paths (C1 and S1) are equally well modeled by either a low velocity zone near 115 km depth or an impedance increase in D” above the core-mantle boundary. This seemingly ridiculous ambiguity, a peak in the upper mantle possibly arising from an opposite polarity impedance contrast in the lower mantle, is due to the sensitivity of first-order reverberation travel time to separation of the mantle reflector from the core-mantle boundary and/or the free surface (see Revenaugh and Jordan, 1991b) and not directly on reflector depth. Path C3 also has a discontinuity in D”; modeling an upper-mantle low velocity zone in its place produced a much lower quality of profile fit, but either interpretation is plausible in light of previous work. These paths cross the Caribbean in a region where a D” discontinuity has been detected [e.g. Garnero and Lay, 2003], but the paths with a potential low velocity zone also traverse oceanic lithosphere, where such a discontinuity is often observed [Revenaugh and Jordan, 1991c]. It is possible that it is some combination of the two that create our results. Two other paths which cross the Caribbean region (C2 and S2) do not require a discontinuity in D”. In general, the sporadic appearances of the shallow upper mantle (or D”) discontinuities limit further interpretation.

Table 2. Path-averaged discontinuity depth, z (km), and measured impedance contrasts, \( R(z) \) (%), of upper mantle discontinuities in the preferred models for the eight paths in the study. Bandpass filter parameters provided are low cut, low corner, high corner, and high cut in mHz. See Figure 1 for path locations.

<table>
<thead>
<tr>
<th>Path</th>
<th>Bandpass (mHz)</th>
<th>H z, km R(z)</th>
<th>G/D” z, km R(z)</th>
<th>L z, km R(z)</th>
<th>410 z, km R(z)</th>
<th>520 z, km R(z)</th>
<th>660 z, km R(z)</th>
</tr>
</thead>
<tbody>
<tr>
<td>C1a</td>
<td>8 10 40 60</td>
<td>2673 6.48</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>C1b</td>
<td>8 10 40 60</td>
<td>123 -6.88</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>C2</td>
<td>12 14 40 60</td>
<td>256 3.86</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>C3</td>
<td>10 12 40 60</td>
<td>2803 3.71</td>
<td>269 2.57</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>C4</td>
<td>10 12 40 60</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>C5</td>
<td>16 18 40 60</td>
<td>54 2.63</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>C6</td>
<td>10 12 40 60</td>
<td>105 2.15</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>S1a</td>
<td>8 10 40 60</td>
<td>2691 5.60</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>S1b</td>
<td>8 10 40 60</td>
<td>110 -4.16</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>S2</td>
<td>14 16 40 60</td>
<td>55 7.28</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
The 410-km, 520-km, and 660-km discontinuities are seen in all profiles and have path-averaged depths and impedance contrasts that vary across the study area (Table 2). The small amount of path-to-path average depth variation (Plate 1) that exists extends beyond the error estimate of 5-7 km for path-averaged discontinuity depth used by Revenaugh and Jordan [1989]. However, we increase that error estimate to ±12 km to account for greater lateral heterogeneity along the long source-receiver paths, which places the topography along the 410-km discontinuity (±12 km) at the edge of resolution and depth variability along the 520-km and 660-km discontinuities (±10 km and ±8 km, respectively) below it. On average, the discontinuities are shallower beneath the eastern part of the study area and impedance contrasts for the 410-km and 660-km discontinuities are small throughout. The latter is likely due to heterogeneity along these long and tectonically variable paths. As previously discussed, impedance contrast estimates obtained in this study should be considered in a relative, as opposed to absolute, sense (Plate 2). Of particular note is path C1, which has a very strong 520-km discontinuity, but only weak impedance contrasts for the 410-km and 660-km discontinuities (Table 2). In fact, the 520-km discontinuity is roughly twice the strength of the other two discontinuities. Nearby paths (C2 and S2) also have strong 520-km discontinuities, on the order of that seen at 410 or 660 km. We believe this is evidence of a wet transition zone along the three paths.

**DISCUSSION**

Wadsleyite and ringwoodite are the major minerals of the transition zone and can incorporate considerable amounts of water into their structures, favoring water storage in the transition zone rather than the upper mantle above [Kohlstedt et al., 1996; Hirschmann et al., 2005]. Water solubility is much lower in the lower mantle [Boffin-Casanova et al., 2000; Murakami et al., 2002], and the partitioning of water favors incorporation into the transition zone at the 660-km discontinuity. The transition zone thus may act as a key water reservoir in the mantle. The presence of a large amount of water there would affect discontinuities bracketing and inside the transition zone, influencing their depths and apparent impedance contrasts.

The most striking characteristic of anomalous path C1 is the unusually large impedance contrast at the 520-km discontinuity (both relative to the 410-km and 660-km discontinuity and in reference to other studies sampling the 520-km discontinuity over much of the globe [Shearer, 1990; Deuss and Woodhouse, 2001; Revenaugh and Jordan, 1991b]). The ratios of the apparent impedance contrasts between the 520-km and 410-km or 660-km discontinuities were the characteristics used to define the geographic extent of the anomalous corridor, with ratios of either 520:410 or 520:660 > 1.25 considered highly anomalous. Only paths C1, C2, and S2 fall into this category. Although geographically proximal, no data is shared between these paths, strongly suggesting some regional structure is responsible for the unusually energetic reflections from the 520-km discontinuity.

The impedance ratios of the 520-km discontinuity to the 410-km and 660-km discontinuities are sufficiently high, however, as to cast some doubt on the veracity of this result. To test it, a modified jack-knife was applied to each of the anomalous paths to examine the contribution of individual seismograms to the measured impedance contrast at 520 km. Stacked reflectivity profiles were recalculated for each permutation of the dataset created by dropping one seismogram. The resulting impedance ratios showed variability, both upwards and down, and scaling with the number of seismograms in the stack. In the cases where removing a trace lowered the 520:410 or 520:660 impedance contrast ratios, no stations, events, or event depths were common among removed traces. The highly anomalous impedance ratios are not the product of a noise burst or “rogue” seismogram or event, but rather appear to be reliable, albeit noisy, estimates.

The phase transition from wadsleyite to ringwoodite at 520 km depth is expected to occur over a depth range of 30 km [Akaogi et al., 1989]. The incorporation of water into the structures of transition zone minerals narrows the range over which the phase transition occurs to as little as 15 km [Inoue et al., 1998]. Similar observations have been made using seismic methods. The 520-km discontinuity is generally observed to occur over a range of up to 50 km [e.g. Cummins et al., 1992]. Due in part to this broad transition interval, the 520-km discontinuity typically has a measured impedance contrast that is only roughly half that at the 410-km discontinuity [Shearer, 1990]. However, in the presence of water, the apparent impedance contrast observed at 520 km depth is increased [e.g. Van der Meijde et al., 2005], as is the ratio between the impedance contrasts for the 520-km and 410-km discontinuities.

Under hydrous conditions, the phase change from olivine to wadsleyite at the top of the transition zone occurs at lower pressure conditions and over a broader range of pressures than under anhydrous or saturated conditions [Wood, 1995; Chen et al., 2002; Smyth and Frost, 2002; Hirschmann et al., 2005]. If water is present, the 410-km discontinuity should be observed at shallower depths [Wood, 1995], and the transition interval will depend on the partition coefficient of water between wadsleyite and olivine at those conditions [Hirschmann et al., 2005]. Similarly, the transition across the 660-km discontinuity is also expected to broaden if the
Plate 1. Topography of transition zone discontinuities superimposed over the NA04 (Van der Lee and Frederiksen, 2005) tomography model sliced across 28° N. Horizontal lines are global mean depths for the three discontinuities (Flanagan and Shearer, 1998).
partitioning of water between the transition zone and lower mantle is of the same order of magnitude as the partitioning between the transition zone and the overlying mantle [Smyth and Frost, 2002], though the broadening expected at 660 km depth is less than that at 410 km [e.g. Higo et al., 2001; Smyth and Frost, 2002].

Broadening of a transition that is centered on the same depth as the original discontinuity will not directly affect the multiple ScS phases, unless that broadening is unusually large (>25 to 30 km for most of our paths). However, it can affect apparent discontinuity depth and estimated impedance contrast if the broadening is not symmetric above and below the original transition. Asymmetric broadening of a discontinuity changes its centroid depth; variable broadening induces additional topography along the path and leads to a dampened or lowered impedance contrast obtained from the ScS reverberation method. Therefore, while the multiple ScS reverberations may not be directly sensitive to the thickness of the phase boundary at a mantle discontinuity, their interactions with that broadened region across a source-receiver path still result in a lowered impedance contrast.

The response of the 410-km and 660-km discontinuities to water results in a broadened transition zone in hydrous regions. Transition zone thicknesses range from 243 to 267 km (± 17 km) in our study, and all are thicker than the global average of 241 km [Flanagan and Shearer, 1998]. Paths that show hydrous signatures in impedance contrasts have the thickest transition zones. The broadened transition zones are primarily the result of shallow 410-km discontinuities. The 660-km discontinuity exhibits less depth variation and no topographic correlation, negative or otherwise, with the 410-km discontinuity.

We see no evidence of a partial melt layer on top of the 410-km discontinuity [e.g., Song and Helmberger, this volume]. The absence of a melt layer does not necessarily contradict the implication of water in the transition zone. There could be enough water in the transition zone to amplify the measured impedance contrast across the 520-km discontinuity without exceeding the saturation limit (~0.4 wt% H2O) for the mantle above 410 km depth [Hirschmann et al., 2005]. Alternatively, the water content of the transition zone may exceed the saturation limit of the mantle above but the hydrous material may be localized within the transition zone and not upwelling across the 410-km discontinuity. Lastly, a thin melt layer may be present but not detected. For the long-period shear waves we examine, any melt layer would need to exceed 20 to 30 km to appear separate from the underlying 410-km discontinuity. A thinner layer would act to lower the apparent impedance contrast across the 410-km reflector, augmenting the effect of transition broadening and topography.

Before concluding that water is present in large quantities in the transition zone along paths C1, C2, and S2, we must consider other mechanisms for increasing the measured impedance contrast across the 520-km discontinuity relative to the 410-km and 660-km discontinuities. And in fact, some additional upward driver is needed to achieve the unusually large ratios we observe. Since path-averaged topography along the three transition zone discontinuities is at or below the error estimates, other mechanisms for producing topography are not discussed. The shallow 410-km discontinuity that exists along two paths sampling the eastern United States is interpreted in conjunction with the elevated impedance contrast at 520 km rather than as a separate line of evidence for a wet transition zone beneath the anomalous region.

Significant along-path topography on the 410-km and 660-km discontinuities would lower the measured impedance contrasts across those boundaries. A more-nearly “flat” 520-km discontinuity would then appear larger in a relative sense. Path-averaged depths of the three discontinuities do not exhibit greater variability for the 410-km and 660-km discontinuities. Nor is this behavior seen globally, as seismic studies show the opposite: a more undulatory 520-km discontinuity [e.g. Deuss and Woodhouse, 2001]. The inferred reflection coefficient of the 520-km discontinuity along path C1 also is large in an absolute sense, not just in relative terms, implying it is being driven up by some process and is not simply less damped than the 410-km and 660-km discontinuities. As a result, we consider relative topography variability as a possible, perhaps likely, addition to water-induced impedance ratio amplification, but not a substitute. A second possibility is that the multiple ScS reverberations are imaging a fast velocity structure confined to the lowermost transition zone and geographically situated beneath the Gulf of Mexico. Under this scenario, what we image as the 520-km discontinuity is a mix of the globally observed phase transition and a seismically fast local thermal and/or chemical anomaly. The latter, however, is not observed in North American tomographic models [Van der Lee and Frederiksen, 2005]. A third possibility is that there is a velocity-neutral, high-density feature in the region. The associated mass excess should produce a large excursion in the geoid in the Gulf region, an excursion not seen in the measured geoid [Lemoine et al., 1997]. A final possibility is that another compositional defect has the same effect as hydrogen on discontinuity depths and impedance contrasts across the transition zone discontinuities and also lowers the shear wave velocities of the minerals in the deep upper mantle. With future research, another feasible option may be presented, however a mechanism for introducing that compositional defect into a localized region of the mantle is also required to explain our results. In the absence of a viable
Plate 2. Path-averaged ratios of impedance contrast for the 520-km discontinuity to the 410-km (left) and 660-km (right) discontinuities. Symbols plotted are surface bounce points of multiple ScS phases for individual traces. Gray rectangles highlight a swath of elevated 520-km impedance contrasts.
alternative, we conclude that the transition zone is unusually wet along paths C1, C2, and S2. The expected behavior of water on transition zone discontinuity depths and impedance contrasts matches most of our observations, and a subducted slab beneath the region provides a mechanism for introducing water to the region.

Previous seismic studies made observations or predictions consistent with a hydrated transition zone beneath the eastern United States. Nominally anhydrous upper mantle and transition zone minerals containing water in some form are expected to have lower seismic velocities [Karato, 1995; Wang et al., 2003; Jacobsen et al., 2004; Jacobsen and Smyth, this volume]. Low compressional wave velocities measured in rocks beneath the Appalachians in the eastern United States [Taylor and Toksoz, 1982] could be an indication of serpentinitization and hydration of the mantle rocks [Christensen, 1966; Karato, 1995]. A low velocity corridor trending along the eastern coast of the United States is found by Van der Lee and Nolet [1997] and Van der Lee et al. [2005]. The low velocity anomaly is strongest beneath the Gulf of Mexico and extends from 200 km to transition zone depths. The authors interpret it as a water-rich region of the upper mantle. Injection of water into overlying mantle by subduction has been proposed as a mechanism for creating the hydrous region [Van der Lee and Nolet, 1997]. Song et al. [2004] and Song and Helmberger [this volume] observe a low velocity layer atop the transition zone beneath the western United States. This layer is not apparent in our reflection profiles for the same region. However, Song and Helmberger [this volume] report a thickness of ~25 km for the layer, which is pushing the detection limit of multiple ScS reverberations.

Turning the hypothesis of subduction-related water injection into the upper mantle and transition zone around, one could ask why other areas of active or recent subduction do not show clear signs of hydration. The lack of a seismic signature of water in areas where it may be expected does not necessarily exclude the presence of water. The seismic signatures of water are generally subtle and easily masked by stronger heterogeneities that result from large lateral thermal gradients, melt, and anisotropy. Williams and Hemley [2001] estimate that seismic methods generally have a minimum detection limit of more than 1 wt% H2O in the transition zone and 2 wt% H2O above it. Water in quantities greater than these limits may be rare, even in areas where there exists an abundant source of water.

CONCLUSIONS

Discontinuity depths and apparent impedance contrasts obtained from multiple ScS reverberations sampling the mantle beneath the eastern United States and Gulf of Mexico are consistent with a water-rich transition zone. A similar hydrous signature is not seen beneath the central or western United States. Laboratory studies find that the major mantle minerals are capable of storing large amounts of water, leading to as much as 0.4 wt% above 410 km depth and 1.5 wt% in the transition zone [Hirschmann et al., 2005]. The distribution of water in the mantle has implications for planetary accretion, mantle composition, rheology, and convection, but its seismic signature can be subtle. We believe that multiple ScS reverberations are important tools in the search for mantle water.

Acknowledgements. Helpful reviews were received from Emile Okal, an anonymous reviewer, and the book editors. GMT software [Wessel and Smith, 1998] was used to prepare some of the figures. This research was partially funded by the National Science Foundation (EAR-0437424). AMC received support from a University of Minnesota Graduate School Fellowship and Department of Geology and Geophysics Mooney and Dennis Graduate Fellowships.

REFERENCES


Inoue, T., Effect of water on melting phase relations and melt composition in the system Mg$_2$SiO$_4$-Mg$_2$SiO$_3$-H$_2$O up to 15 GPa, *Phys. Earth Planet. Inter.*, 85, 237–263, 1994.


