Crustal-scale shortening structures beneath the Blue Ridge Mountains, North Carolina, USA

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ABSTRACT

We present results from a new seismic data set that show evidence for crustal-scale shortening structures beneath the Blue Ridge Mountains in the Southern Appalachians. The data come from six broadband seismic stations deployed on a transect across the Piedmont and Blue Ridge of western North Carolina. The observed structures appear as both a Moho hole and doubled Moho in receiver function CCP (Common Conversion Point) stacks oriented roughly perpendicular to the trend of the Appalachian orogen. We interpret these features as evidence for tectonic wedging and associated delamination and underthrusting of Laurentian lithosphere beneath a crustal indenter. The Moho hole and underlyng deeper Moho correspond closely to a significant regional Bouguer gravity anomaly low, which we interpret as being due to overthickened, normal-density crustal material. Beneath the indenter, we observe a double Moho, which may correspond to the partial eclogitization of the underthrust material. This would be consistent with the sharp increase in the observed gravity above this feature. In addition to these crustal structures, we see evidence for a mantle lithospheric discontinuity at 90–100 km depth. This increase in velocity with depth is spatially limited and may dip slightly to the west, though more data are needed to verify this result. We interpret this anomaly to be a fossil slab accreted onto Laurentian lithosphere. If the westward dip is accurate, this slab may be a remnant of a west-vergent subduction zone that was active during the accretion of Carolina.

INTRODUCTION

The Southern Appalachian Mountains are composed of rocks and structures from repeated tectonic events beginning with the assembly of Rodinia during the Proterozoic Grenville orogeny and ending with the Mesozoic rifting of Pangea following the Alleghanian orogeny (e.g., Hibbard, 2000; Hibbard et al., 2002; Hatcher et al., 2007a; Hatcher, 2010). What is known about these tectonic events is inferred from geological and geochemical observations, gravity and magnetic anomaly maps, and from a limited number of seismic-reflection and seismic-refraction studies (e.g., Hack, 1982; Hutchinson et al., 1983; Iverson and Smithson, 1983; Prodehl et al., 1984; Nelson et al., 1985; Hubbard et al., 1991; Rankin et al., 1991; Aleinikoff et al., 1995; Loewy et al., 2003; Cook and Vasudevan, 2006; Miller et al., 2006; Hawman, 2008; Anderson and Moecher, 2009; Fisher et al., 2010). Geologic structures observed at the surface in the Southern Appalachians are difficult to link with structures at depth because the upper ~10 km of crust was transported at least tens of kilometers and possibly hundreds of kilometers westward during the Alleghanian orogeny (Iverson and Smithson, 1983; Rankin et al., 1991). Information about structures located beneath this Blue Ridge–Piedmont allochthon comes from a limited number of reflection and refraction studies, most of which do not cross the rugged terrain of the Appalachian escarpment and Blue Ridge Mountains (Nelson et al., 1985; McBride and Nelson, 1988; Hubbard et al., 1991; Cook and Vasudevan, 2006). While there are a handful of permanent broadband stations in the southeastern United States, they are widely spaced and generally located at lower elevations.

This study uses a newly collected data set composed of recordings from six broadband seismic stations deployed across the Piedmont and the Blue Ridge Mountains in North Carolina (Fig. 1). This transect fills an important hole in the much-analyzed Consortium for Continental Reflection Profiling (COCORP) reflection profile across Georgia and Tennessee (Prodehl et al., 1984; Nelson et al., 1985; McBride and Nelson, 1988; Cook and Vasudevan, 2006). We use these data to calculate receiver functions, which can identify seismic discontinuities in the crust and upper mantle. Unlike previous active-source reflection studies, we image the Moho clearly beneath the highest elevations in the Southern Appalachians. Furthermore, this is the first study to image sub-Moho structures down to ~120 km depth in this area. We find evidence for crustal-scale tectonic wedging as well as mantle lithospheric discontinuities that may represent fossil accreted slabs at 100 km depth. These results help to explain the observed gravity anomalies in the area, and they provide data that can be used to test tectonic models for Southern Appalachian evolution.

GEOLOGIC SETTING

Along the east coast of North America, the tectonic events spanning the formation of Rodinia during the Grenville orogeny (1.2–1.0 Ga) to the formation of Pangea during the Alleghanian orogeny (330–280 Ma) represent one complete Wilson cycle (Hatcher et al., 2007a). In the southeastern United States, the Appalachian Mountains are the product of four orogenic events during this cycle: the Grenville orogeny, the Ordovician Taconic orogeny, the Late Devonian–Mississippian Neocadian orogeny, and the Pennsylvanian Alleghanian orogeny (Hibbard et al., 2002; Hatcher et al., 2007a; Hatcher, 2010). Subsequent to these orogenic events, Triassic rifting resulted in the opening of the Atlantic Ocean and the formation of the modern passive margin (Schettino and Turco, 2009).

The Proterozoic Grenville orogeny in the Southern Appalachians was likely due to a collision between Laurentia and South America.
The allochthonous nature of the upper crust makes it difficult to link seismically imaged structures beneath the Blue Ridge–Piedmont allochthon with terrane boundaries observed at the surface.

**PREVIOUS WORK**

Constraints on the geometry of the Blue Ridge–Piedmont allochthon and the structure of the underlying basement come from a number of active-source studies. COCORP acquired seismic-reflection data along several NW-SE-trending profiles in northern Georgia. These data were collected between 1978 and 1985 and were reprocessed by Cook and Vasudevan (2006). Their results show a highly reflective lower crust and Moho beneath the coastal plains and Carolina, but little to no reflectivity in the mid- or lower crust beneath the Inner Piedmont or Blue Ridge. In contrast, the 1985 Appalachian ultradepth core hole (ADCOH) profiles, located slightly farther to the north, do show significant reflectivity in the mid- and lower crust beneath the Inner Piedmont (Hubbard et al., 1991). While the Moho is not resolvable on most of the ADCOH lines, a number of mid-crustal arrivals are visible, indicating generally high reflectivity throughout the region.

Crustal and subcrustal velocities are constrained by a 1965 U.S. Geological Survey (USGS) refraction survey in eastern Tennessee that was reexamined by Prodehl et al. (1984). The survey consisted of two 400-km-long refraction lines with shot points on the ends and two intermediate points per line (Borchert and Roller, 1966). Prodehl et al. (1984) found overall fast crustal velocities in three distinct layers: a shallow 10–20-km-thick layer with Vp = 6.1 km/s, a midcrust extending to ~40 km depth with velocities of ~6.7 km/s, and a fast lower crust (7.3 km/s) underlain by a somewhat slow upper mantle (7.9 km/s) at depths of ~50 km beneath the Appalachians. These crustal thicknesses are consistent with recent work by Hawman (2008), who used wide-angle reflections.
from quarry blasts to constrain crustal thickness and average velocities across much of the Southern Appalachians. He found average crustal velocities of ~6.6 km/s, and crustal thicknesses in our study area of between 47 and 53 km.

In addition to seismic studies, several gravity anomaly studies have been performed to help constrain the structures beneath the Southern Appalachians. The dominant feature in this area is the coupled Bouguer gravity low and high, which extends along the full length of the Southern Appalachians. The gravity low is located close to, but just east of, the Blue Ridge Escarpment, an abrupt rise in elevation between the Piedmont and Blue Ridge geomorphic provinces. The escarpment locally coincides with the Brevard fault, but over most of its length, there are rocks of similar composition and, presumably, erodibility on both sides. This fact has led some to speculate that the escarpment coincides with a postorogenic normal fault (White, 1950; Ackerman and Knapp, 2002), although others have proposed that it represents an erosionally retreating rift-flank uplift (Spotila et al., 2004). Karner and Watts (1983) noted that a similar low/high-gravity pair is observed in the Alps and Himalayas. They modeled their anomalies as the obduction of low-density crustal material and the underthrusting of high-density lower crust beneath this. However, as Hutchinson et al. (1983) noted, all gravity models are inherently nonunique, and they showed that the Appalachian gravity low/high pair can be modeled with variations in crustal thickness or with a high-density dipping suture zone.

It is important to note that despite the multiple geophysical studies across the Southern Appalachians, none has been able to image structures beneath the dominant Moho at ~45 km depth. As such, most interpretations of gravity data have been limited to crustal density variations only. We present evidence that deeper structures likely play a central role in the formation of these gravity anomalies, and they may help to answer some questions about the persistence of the topographic relief in this ancient orogen.

**DATA AND METHODS**

Data from this study come from our Appalachian Seismic Transect (AST) (Fig. 1). Between April 2009 and June 2010, we recorded continuous broadband seismic data along a six-station, 75-km-long transect across the Blue Ridge Mountains of North Carolina. The stations were Nanometrics Trillium-120 sensors paired with Nanometrics Taurus digitizers owned by the University of North Carolina–Chapel Hill.

Receiver functions are useful for studying lithospheric structures because they highlight S-waves that are produced when near-vertically incident P-waves encounter a sharp velocity discontinuity. This “scattered” S-wave will arrive at the recording station slightly later than the direct through-going P-wave, and if the P- and S-wave velocities are known, the delay time can be used to calculate the depth to that discontinuity. We used a time-domain pulse-stripping method, which minimizes the difference between the radial component $C_r(t)$ and the convolution between an iteratively improved time series of Gaussian pulses (the receiver function) with the vertical component $C_v(t)$ (Ligorria and Ammon, 1999). Iterations are stopped when improvements to fit become too small, or some goodness of fit is achieved. One benefit of this method is that it allows us to quantify the percent variance reduction of the receiver function at each iteration, giving us some quantitative measure of error.

For this study, we calculated receiver functions for 91 events with magnitude greater than 6.0 that occurred at distances greater than 30° from the array (Fig. 2; Table 1). We band-pass filtered the records from 0.1 to 5 Hz, and then calculated the receiver functions with Gaussians of $a = 2.5$ and $a = 5$ (corresponding to central frequencies of 1.2 and 2.4 Hz, respectively). We considered only receiver functions with better than 70% variance reduction during the iterative deconvolution, and, subsequently, all of the qualifying receiver functions were inspected for consistency. Requiring a higher percent variance reduction reduces our spatial coverage, but it does not appear to change the locations or amplitudes of observed arrivals. In total, this left us with 247 receiver functions (see Table 2 for individual station numbers). Plots of the raw receiver functions for each station sorted by back azimuth can be seen in Figure 3.

**Constraining Vp/Vs**

One of the inherent limitations of receiver functions is the trade-off between velocity and depth (Ammon et al., 1990). In particular, depth trades off strongly with the Vp/Vs ratio. However, constraints on the average Vp/Vs ratio of the layer between the discontinuity and the surface can be obtained by looking at multiples. Multiples are waves that bounce off the free surface and the discontinuity in question before being recorded at the station as an SV wave. The differential arrival times are given by:

$$\Delta T = T_{Ppp} - T_{Ps} = 2h\sqrt{V_p^2 - p^2},$$

$$\Delta T = T_{Ppms} - T_p = 2h\sqrt{V_s^2 - p^2}. \quad (2)$$

However, because multiples are often small arrivals on individual receiver functions, it is generally easier to identify them by noting how $\Delta T$ changes with changing ray parameter (incidence angle) for direct arrivals (i.e., P-s converted phases) and multiples. For direct arrivals, increasing the ray parameter (and incidence angle) results in an increase in the differential time between the through-going P-wave and the converted wave. For multiples, the differential time decreases with increasing ray parameter. By plotting receiver functions as a function of ray parameter, we are able to identify which arrivals are direct arrivals and which are multiples by their respective moveouts (Zandt et al., 1995). Moveout plots for our stations are shown in Figure 4.
### TABLE 1. EARTHQUAKES USED TO CALCULATE RECEIVER FUNCTIONS

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<th>Minute</th>
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(continued)
These observations can be used to predict the range of thicknesses ($H$) and Vp/Vs ratios ($K$) that best match the observed moveouts of direct and multiple arrivals from the Moho (Zhu and Kanamori, 2000). This can be done by calculating predicted arrival times for the Moho direct arrivals ($t_{Ps}$) and multiples ($t_{PsPms}$, $t_{PpPms}$, $t_{PsPms}$) for a given crustal thickness and Vp/Vs ratio and summing the amplitudes for all receiver functions:

$$S(H, K) = W_{Ps} r(t_{Ps}) + W_{PsPms} r(t_{PsPms}) - W_{PpPms} r(t_{PpPms}),$$

where $W$ values are weights applied to the direct arrivals and multiples (we use 1/3 for each), and $r(t)$ is the amplitude of the receiver function at that time. Note that the amplitudes of the PpSms and PsPms arrivals are subtracted due to their predicted negative amplitudes (a consequence of the free-surface–S-wave interaction). This sum is calculated over a range of $H$ and $K$ values, and the amplitudes are plotted and contoured. The highest amplitude corresponds to the best-fitting thickness and Vp/Vs ratio for a given fixed P-wave velocity. We assume a crustal average P-wave velocity of 6.6 km/s based on the results of Prodehl et al. (1984) and Hawman (2008). Results of our $H$-$K$ stacks can be seen in Figure 5.

**Common Conversion Point Stacks**

While we assume near-vertical incidence for our incoming P-waves, in reality, the incidence angle $i$ is always greater than zero. As such, the point at which the incoming ray pierces a given horizon is not directly below the station and may in fact be closer to or directly below an adjacent station. The deeper the horizon in question, the farther the pierce point will be from the recording station. Figure 6 shows pierce points for our receiver functions at 45 and 100 km depth. If lateral heterogeneities exist along a given discontinuity, it is important that we stack, not by station, but by pierce point at that depth. This is the goal of common conversion point (CCP) stacking (Dueker and Sheehan, 1997; Gilbert, 2003, 2004). This methodology back-projects the incoming ray paths and stacks rays that pass within a given distance of a stacking point. We

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**TABLE 2. NUMBER OF RECEIVER FUNCTIONS USED FOR EACH STATION IN THIS STUDY AND AVERAGE CRUSTAL VALUES DETERMINED FROM H-K STACKING**

<table>
<thead>
<tr>
<th>Station name</th>
<th>Number of receiver functions</th>
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<th>Average crustal Vp/Vs</th>
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<td>46</td>
<td>1.78</td>
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<tr>
<td>BSC</td>
<td>40</td>
<td>46</td>
<td>1.73</td>
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</tbody>
</table>

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![Figure 3. Plot of raw receiver functions calculated with a Gaussian width of $a = 2.5$, sorted by back azimuth at each station.](https://pubs.geoscienceworld.org/gsa/lithosphere/article-pdf/4/3/242/3045114/242.pdf)
create a two-dimensional (2-D) grid of stacking points with 5-km spacing that spans the area sampled by our pierce points at 100 km depth (Fig. 6). Smoothing is achieved by including pierce points from distances equal to twice the stacking node spacing. Stacked bins at a given depth are only plotted if they include data from a minimum of five receiver functions. We subsequently apply the phase-weighting algorithm of Frassetto et al. (2010) in order to emphasize coherent arrivals (Schimmel and Paulssen, 1997). This methodology sums the instantaneous phase of the receiver functions in a given bin and uses this value to weight the stacked receiver functions. The result is that coherent arrivals such as the Moho are emphasized, whereas incoherent arrivals from small structures and ambient noise are reduced. Stacks are calculated for both the higher- and lower-frequency receiver functions calculated earlier.

**RESULTS**

**Crustal Structure**

Average crustal thicknesses and Vp/Vs ratios beneath the AST stations can be seen in the H-K stacks in Figure 5. The contours show 80% and 90% of the maximum stacked amplitude for a given station. Vp/Vs is much more poorly constrained than crustal thickness, but within error, all stations show a crustal thickness of ~46 km and an average Vp/Vs ratio of 1.74. Constraints on both depend on the ability to identify direct arrivals and multiples on the individual station moveout plots (Fig. 4). All stations show a very clear direct arrival (positive moveout with respect to p) for a significant increase in velocity with depth (positive receiver function arrival) at 5–6 s after the direct P-wave arrival. This corresponds well with the predicted times for a discontinuity at 46 km depth, assuming a crustal P-wave velocity of 6.6 km/s and a Vp/Vs ratio of 1.74. The negative moveout, positive-amplitude arrival associated with the PpPms multiple is also clear at all stations, with the possible exception of RGR. The negative moveout, negative-amplitude receiver function arrival associated with...
with the PsPms and PpSms arrivals can be seen at BSC, AFH, and WWH, but it is less clear at MFH, BEE, and RGR.

Also noticeable on the moveout plot for BSC is a second arrival just after the Moho arrival, corresponding to an increase in velocity at 60 km depth. The arrival fades at higher ray parameters, perhaps because these rays sample a larger area around the station, resulting in incoherent stacking if the structures vary laterally. A faint arrival at the same time can be seen in the moveout plot for MFH.

CCP stacking results for crustal-scale structures are shown in Figure 7. These stacks use receiver functions calculated with our higher-frequency Gaussian of $\alpha = 5$. These cross sections all show a clear Moho arrival at ~45 km depth, possibly deepening slightly at the NNW edge. The Moho arrival is not continuous, however. All profiles show either a marked weakening or hole, usually just SSE of the Blue Ridge Escarpment. These Moho holes are most pronounced on those transects that cross the escarpment at near-normal angles (profiles 7–9), but they are seen as reductions on profiles west of our stations where the strike of the transects trends subparallel to the escarpment. In addition to the Moho hole, we see a second “Moho” arrival on the central and eastern profiles at ~60 km depth. This “double Moho” extends from beneath the Moho hole (or weakening) to the SSE edge of our resolved area. The locations of the Moho hole and double Moho are shown as ovals in the CCP cross sections in Figure 7, and as stars and circles, respectively, in map view in Figure 8. Figure 8 also shows the correlation between these features and regional Bouguer gravity anomalies. In general, where the Moho is weakened or missing, the Bouguer anomalies are markedly low, but where the Moho is doubled (i.e., clear arrival at both 45 and 60 km depth), the Bouguer gravity anomaly values increase, becoming much less negative.

In order to determine if these observations could be a local anomaly or artifact caused by one station, we restacked the CCP profiles six times, each time omitting the receiver functions from one of our stations. Notably, removing the receiver functions recorded at BSC did not eliminate any of the features discussed (the Moho hole or double Moho). Figure 9 shows crustal profiles 7 and 8 both with and without data from BSC. The absence of evidence for the double Moho in the moveout plots must therefore be due to stacking across heterogeneous structures at adjacent stations.

**Mantle Structure**

In addition to crustal direct arrivals and multiples, the moveout plot for station BSC (and to a lesser extent for station MFH) in Figure 4 shows a prominent direct arrival at ~10 s, which would correspond to a significant increase in velocity with depth at ~90 km depth. Figure 9B shows CCP profile 8 using receiver functions calculated with a Gaussian of $\alpha = 2.5$. This profile also clearly shows the Moho hole and double Moho observed in the higher-frequency receiver function stacks of Figure 7. The deep discontinuity lies at depths between 90 and 100 km, and it may be discontinuous beneath the center of the array. It appears to dip slightly to the west, though more data are needed to evaluate this possibility.

As with the crustal results, we also restacked the CCP profiles omitting one station at a time in order to determine if any one station was controlling one of the deeper arrivals. The deep arrival farthest to the NNW does appear to be largely controlled from data at station BEE. However, the rest of the deeper discontinuity is still visible even with data from any one station removed. Results of the stacking with station BSC omitted are shown in Figure 9B. Note that even though BSC is the only station showing a clear arrival at ~90 km depth in the moveout plots (Fig. 4), removing all of the BSC data from the CCP stacking does not remove these anomalies, thereby demonstrating that the absence of these
Figure 6. Map of common conversion point (CCP) nodes (black diamonds) and receiver function pierce points at 45 km (red dots) and 100 km depth (blue dots). Stations are shown as light-blue triangles. Numbers at the ends of the CCP node lines indicate profile numbers referenced in Figures 5, 7, 8, and 9. Dashed line shows the Blue Ridge Escarpment. Solid lines show state boundaries.
Figure 7. Common conversion point (CCP) stacked profiles using receiver functions calculated with a Gaussian width of $a = 5 \, (~2.4 \text{ Hz})$. Red indicates positive receiver function arrivals; blue indicates negative receiver function arrivals. Black triangles show station locations, and gray profiles show exaggerated topography. The vertical blue line on the topography indicates the location of the escarpment on that profile. Ovals show locations of weak Mohos, Moho holes, and the deeper Moho at 60 km depth.
Figure 8. Bouguer gravity anomaly map (in color with contours). Black diamonds show common conversion point (CCP) node locations (same as in Fig. 4). Circles show locations of Moho holes (white circles), weak Mohos (gray circles), and double Mohos (orange circles). Dashed line shows the approximate location of the Blue Ridge Escarpment. Notice how the Moho holes line up with the lowest gravity values, whereas the double Moho coincides spatially with a strong increase in gravity, especially where the CCP profiles are stacked roughly normal to the trend of the escarpment and gravity anomalies. White triangles are station locations (see Fig. 1 for station names).
Figure 9. Common conversion point (CCP) profiles with and without data from station BSC. Note that the Moho holes, the double Moho, and the 100-km-depth discontinuity are present even without using data from station BSC: (A) Profiles 7 and 8 from Figure 5 (Gaussian $a = 5$). Top plots show CCP stacks with all stations included. Bottom plots show CCP stacks calculated without BSC data. (B) Profile 8 from Figure 8 (Gaussian $a = 2.5$). Left plot uses all data; right plot omits data from BSC.
features in the moveout plots of other stations is due to stacking across heterogeneous structures.

**DISCUSSION**

**Crustal Structure**

Contrary to the results of studies using COCORP reflection data, we find the base of the crust beneath the Piedmont and Blue Ridge to be highly reflective. The average crustal thickness of ~46 km is consistent with the results of Hawman (2008) and Prodehl et al. (1984), but it is somewhat deeper than that suggested by Cook and Vasudevan (2006). This is likely due to the slow crustal velocity used by Cook and Vasudevan (2006) to migrate time to depth (6 km/s).

Locally, the topographic relief along our transect reaches 1200 m, though in other parts of the Blue Ridge Province, it is greater than 1500 m. The persistence of high topographic relief in ancient mountain ranges is a long-standing problem in geology because many models predict that most of the topographic relief in a mountain range should have eroded away after ~100 m.y. (Ahnert, 1970; Tucker and Slingerland, 1994; Beaumont et al., 2000). The Southern Appalachians presumably attained their greatest elevation during the Alleghanian orogeny ca. 300 Ma, and, according to models like Ahnert (1970), almost all of the topographic relief should have been eliminated by ca. 200 Ma. Hack (1982) felt that postorogenic uplift was responsible for much of the present-day topographic relief in the Southern Appalachians. Matmon et al. (2003), however, calculated long-term erosion rates in the Southern Appalachians using cosmogenic isotopes and concluded that the present-day topography is the result of slow erosion of crust that had been thickened during the Alleghanian orogeny. The thickness of our observed crust suggests that the elevations of the Southern Appalachians are isostatically compensated, which is consistent with Matmon’s interpretation. Using a reference crust of 35 km thickness with a density of 2.7 g/cm³ and a mantle density of 3.3 g/cm³, and assuming the crustal thickness is 46 km (applicable beneath the highest elevations in this study), the average crustal density of the Blue Ridge Mountains would have to be somewhat higher (~2.77 g/cm³) to explain the observed roots and elevations.

To our knowledge, we are the first to report discontinuities in the Moho as evidence of a lithospheric downwelling or “drift.” Double Mohos have been reported in the Himalayas, where they are interpreted as evidence of the partial eclogitization of the lower crust (Schulte-Pelkum et al., 2005). A similar explanation was presented by Gilbert et al. (2006) to explain the presence of a double Moho beneath central Chile. Hansen and Dueker (2009) found evidence for a double Moho using both P-s receiver functions and S-p receiver functions recorded across the edge of the Wyoming craton. They suggested that this double Moho represents tectonic wedging of the Wyoming craton into younger Proterozoic crust.

Wedge tectonics have become recognized as a common feature in accretionary margins within continental interiors (Price, 1986; Martinez-Torres et al., 1994; Snyder, 2002; Moore and Wiltshcko, 2004). Price (1986, p. 239) defined wedges as “bounded above and below by shear zones of opposite vergence,” similar to the conjugate surfaces observed in experimentally compressed marbles and limestones. Snyder (2002) showed evidence for whole-crustal tectonic wedging at the margins of the Archean cratons in Canada and Fennoscandia. Moore and Wiltshcko (2004) defined different wedging geometries and attempted to ascribe these to all major orogenic belts.

Figure 10 shows two possible interpretations of the tectonic structures that may be responsible for our observed Moho hole discontinuity and double Moho. In both cases, part of the lower crust of the Laurentian basement is underthrust beneath an indenter. This is consistent with the crustal-delamination model proposed by Moore and Wiltshcko (2004) for the Central Piedmont shear zone. The Moho hole is the result of the downdragging portion of the Laurentian lower crust (similar to the Moho hole in the Sierras being caused by sinking of the lower crust). The double Moho to the SSE suggests that this lower crust is of higher seismic velocity than the indenter above it, perhaps due to the partial eclogitization of the underthrust material. The close correlation of the location of the Moho hole with the pronounced gravity low in the area (Fig. 8) may be due to the presence of overthickened, but not eclogitized, continental crust. The increase in the gravity to the SSE of this low would be consistent with the presence of dense, eclogitic, lower crust beneath the indenter.

The difference between the two models in Figure 10 lies in the affinity of the indenter and the timing of the deformation. In the first model, the observed wedge structures do not necessarily correspond to a terrane boundary, but may instead be whole-crustal shortening structures due to a major collisional event. In this model, we propose that the deformation would have preceded the formation of the Blue Ridge–Piedmont allochthon during the Alleghanian, so that the surface expression of this wedging was subsequently deformed and displaced to the northwest along with all of the other adjacent terrane boundaries during the Alleghanian collision. The second model is similar to models presented in Keller and Hatcher (1999) and Hubbard et al. (2002). In this model, the indenter is the part of Carolina that has both under- and overthrust the Laurentian basement. This wedging could either have occurred during the initial accretion of Carolina or, more likely, subsequently, during the Alleghanian orogeny and formation of the Blue Ridge–Piedmont allochthon. If the wedging did occur during the initial accretion of Carolina, this would imply that the accretion was the result of east-dipping subduction, because no structures continue past the indenter dipping west into the mantle as would be expected with west-dipping subduction. If the thrusting occurred later during the Alleghanian, however, subduction during the accretion of Carolina could have proceeded via either east- or west-dipping subduction.

Given that the high elevations NNW of the Blue Ridge Escarpment appear to be isostatically compensated by the 46-km-thick crust observed beneath them, it is somewhat surprising that the low-lying areas SSE of the escarpment are underlain by a 60-km-thick crust. However, if the underthrust portion of the crust beneath the indenter is eclogitized, this low-lying area may be isostatically compensated. Assuming the same reference crust as before, and assuming the top 45 km section of crust beneath these low-lying areas has the same density we calculated earlier for beneath the Blue Ridge Mountains (2.77 g/cm³), the ~385 m elevation SSE of the escarpment would be isostatically compensated if the bottom 15 km of crust (the underthrust portion beneath the Moho at 46 km depth) had an average density of 3.4 g/cm³. This is well within the range of densities expected for an eclogitized lower crust (3.1–3.6 g/cm³) (Austrheim, 1991). Given that there is a low-gravity anomaly collocated with the Moho hole immediately SSE of the escarpment, we do not propose that this section is eclogitized.

However, the double Moho begins only 20 km farther SSE, coinciding with a marked increase in gravity. Detailed gravity modeling will be required to determine if the precise anomalies observed can be explained with an underthrust eclogitic root. We do note that the location of the Blue Ridge Escarpment is aligned very closely with the NNW edge of the observed Moho hole (Fig. 10). This may be coincidental, given that the high-density lower crust is believed to begin 20 km farther SSE. However, the observed
elastic thickness beneath the Southern Appalachians is particularly high (40–70 km) (Stewart and Watts, 1997), so it may be possible that the high-density lower crust in the area plays a role in perpetuating this long-lived topographic feature.

**Mantle Structures**

The cause of the sharp increase in velocity with depth at ~90–100 km beneath the Blue Ridge Mountains is uncertain. Seismic discontinuities at these depths are often referred to as Hales discontinuities (Hales, 1969), which are believed to be due to the phase transition from spinel to garnet in the upper mantle—a transition for which depths are highly variable and are strongly controlled by the concentration of chromium (Klemme, 2004). In our case, the observed discontinuity appears to be spatially limited, appearing and disappearing over a few tens of kilometers, making it unlikely to be due to such a phase transition. If, instead, this increase in velocity with depth is due to some persistent tectonic feature, then this tectonic feature must be part of the mantle lithosphere, and the mantle lithosphere must in turn be over 100 km thick in this area. While we do not observe the decrease in velocity that would be associated with the lithosphere-asthenosphere boundary, this thickness constraint is consistent with a number of studies.

Figure 10. Tectonic interpretation of profile 7 from Figure 7. Heavy red line above the topography shows the Bouguer gravity anomaly along this profile from Figure 8. CPS—Central Piedmont Suture.
indicating a deep lithosphere-asthenosphere boundary beneath the eastern United States (Rychert et al., 2005, 2007; Abt et al., 2010).

One possible tectonic feature that might fit the observed discontinuity is a discontinuous accreted fossil slab. Similar small regions of seismic discontinuities at these depths have been observed beneath southern Finland and have been interpreted as relics of ancient subduction zones (Yliniemi et al., 2004). The age and affinity of such a fossil slab beneath the Blue Ridge are hard to constrain without a more spatially extensive seismic data set. If the apparent westward dip is verified, this might suggest that the relict slab is lapetan, with subduction ending with the accretion of Carolina either in the Ordovician or mid-Devonian. Hibbard (2000) proposed that the convergence and subsequent accretion of Carolina were accommodated by a westward-dipping subduction zone that was active during the Late Ordovician and Early Silurian. Anderson and Moecher (2009) proposed that the Piedmont terrane was accreted during westward subduction during the Ordovician. Hatcher (2010) proposed that Carolina accreted in the Devonian and showed convergence being accommodated on an eastward-dipping subduction zone. If the slab was associated with accretion of the Carolina or Piedmont terranes, its present-day geometry is consistent with the westward-dipping subduction proposed by Hibbard (2000) and Anderson and Moecher (2009).

CONCLUSIONS

(1) We present receiver function images from the first broadband seismic transect across the Southern Appalachians. These cross sections provide the first look at subcrustal structures beneath the ancient orogen and show evidence for whole-crustal-scale compressional structures beneath the Blue Ridge Mountains of North Carolina.

(2) These compressional structures are spatially closely correlated with regional gravity anomalies, suggesting that the marked Bouguer gravity anomaly lows are due to the observed overthickened crust that resulted from this shortening, and that the increase in gravity to the SSE is due to the partial eclogitization of the underthrust Laurentian basement beneath an indenter.

(3) The high elevations NNW of the Blue Ridge Escarpment are isostatically compensated, assuming an average crustal density of 2.77 g/cm$^3$ for the 46 km thick crust in this region. The low-lying areas SSE of the escarpment are underlain by 60 km of crustal material. This suggests either that these areas are dynamically being held down or that some portion of the underthrust basement has become eclogitized. If the bottom 15 km section of the 60-km-thick crust is eclogitized, then the area would be iso-


costatically compensated. This might help explain the longevity of the Blue Ridge Escarpment.

(4) We observe localized seismic scatterers at depths between 90 and 100 km, suggesting the presence of mantle lithosphere to at least these depths throughout our study area. This seismic scatterer appears to dip slightly to the west, which may be consistent with westward-dipping subduction during the accretion of Carolina, though more data are needed to confirm this result.

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