P-wave Reflectivity of the Crust and Upper Mantle Beneath the Southern Appalachians and Atlantic Coastal Plain using Global Phases

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KEY POINTS

- P-wave reflections generated using PKPdf as a virtual source are observed over depths from 1 km to greater than 200 km.
- Reflection times reach 17.4 s (> 55 km) for the highest elevations, supporting previous evidence that the southern Appalachians are in isostatic equilibrium.
- Reflections at 32-36 s (120-135 km) may indicate drag-induced flow at the base of the lithosphere. These arrivals are observed for single as well as stacked events.

INDEX TERMS

- Seismology
- Tectonophysics

KEYWORDS

- southern Appalachians, Atlantic Coastal Plain, PKPdf, PKIKP, Moho, LAB

PLAIN LANGUAGE SUMMARY

The main goal of this work was to study the nature of the boundary between the tectonic plate beneath the southern Appalachians and the underlying, more fluid mantle. We used echo soundings to determine the depth and physical characteristics of major layers. Our strategy was to use seismic waves generated by earthquakes on the opposite side of the planet as an energy source. We see a major transition at depths of 120-135 km that we interpret as drag-induced flow just beneath the plate. We also see evidence for a thickening of Earth's crust beneath the highest elevations of the southern Appalachians that suggests that these very old mountains are in gravitational equilibrium.

Abstract

Reflection profiles generated using PKPdf as a virtual source show laterally continuous reflections from structures at depths less than 1 km to roughly 200 km beneath the southern Appalachian orogen and Atlantic Coastal Plain. Arrivals interpreted as reflections from the Moho increase in time from ~10 s beneath the Coastal Plain to 17.4 s (~57 km) beneath the Blue Ridge Mountains, providing additional evidence that the southern Appalachians are in rough isostatic equilibrium. Reflections at 32-36 s (120-135 km) are consistent with the depth to the base of the lithosphere found in recent inversions of Ps arrivals and surface waves. Alternatively, these and later reflections at times up to 58 s (~224 km) may be due to layering associated with drag-

induced flow in the asthenosphere, suggesting largely horizontal rather than vertical flow for depths less than 225 km beneath the Georgia coastal plain.

1. Introduction

The passage of USArray across the eastern United States has resulted in fundamentally new insights into the fine-scale structure of the continental lithosphere. Recent analyses of Transportable Array (TA) data for upper mantle structure, including body-wave and surface-wave tomography and analyses of shear-wave splitting, indicate that the lithosphere thins from a maximum of 200-250 km beneath the North American craton [*Yuan et al.,* 2014] to roughly 150 km beneath the Grenville Front and Appalachian Mountains (Valley & Ridge and Blue Ridge), to less than 100 km beneath portions of the SE Atlantic Coastal Plain [*Pollitz and Mooney,* 2016; *Shen and Ritzwoller,* 2016; *Savage et al.,* 2017]. Analyses of shear-wave splitting generally show fast axes parallel to absolute plate motion (APM) beneath the craton [*Yuan et al.,* 2014; *Long et al.,* 2016; *Yang et al.,* 2017] with more complex patterns eastward toward the continental margin.

These patterns have been interpreted using a variety of models. For instance, fast axes parallel to the belt of highest elevations in the Appalachian Orogen are also roughly parallel to APM; they have been interpreted both as evidence for strain associated with Alleghanian collision frozen into the lithosphere [*Long et al.,* 2016] and as simple shear generated just above and below the lithosphere-asthenosphere boundary (LAB) by ongoing plate motion [*Yang et al.,* 2017]. Beneath the southeastern U.S. coastal plain, interpretations of a broad zone of null [*Long et al.,* 2016] or small [*Yang et al.,* 2017] splitting times include vertical flow in the asthenosphere (possibly a consequence of edge-driven convection driven by an abrupt transition in lithospheric thickness) [*Long et al.,* 2016; *Savage et al.,* 2017] combined with weak or spatially incoherent anisotropy in the lithosphere. An alternative model involves cancellation of the effects of APM-parallel flow within the broad lithosphere-asthenosphere transition zone by roughly N-S directed flow in the asthenosphere diverted around the keel of the craton [*Yang et al.,* 2017].

Vertical incidence reflection profiling has the potential to help resolve some of these issues by tracking the lateral continuity of structures within the uppermost mantle. Several large-scale, active-source experiments across Eurasia have been able to detect coherent reflections at normal incidence in the mantle to depths greater than 200 km using large explosions [*Knapp et al.,* 1996; *Steer et al.,* 1998]. In this study, we use the global seismic phase PKIKP (PKPdf) as a virtual source [*Ruigrok and Wapenaar,* 2012] to construct normal-incidence reflection sections that are analogous to those produced by active-source reflection profiling of the crust. The strategy for this work is to take advantage of the relatively dense station spacing of the broadband arrays deployed during the Southeastern Suture of the Appalachian Margin experiment (SESAME) (Figure 1) [*Parker et al.,* 2013] to investigate P-wave reflectivity or "fabric" over a portion of the upper mantle for which the long-wavelength velocity and anisotropy structure are well constrained by recent analyses of TA data.

2. Geologic and Tectonic Setting

The southern Appalachians (Figure 1) are the product of diachronous, largely oblique collision of Laurentia with Gondwana and a number of continental fragments and island arcs [*Hatcher,* 1989, 2002, 2010; *Hatcher et al.,* 2007], beginning ~480 Ma and culminating in the Alleghanian orogeny (330-260 Ma). Detachment faults imaged by the Consortium for Continental Reflection Profiling (COCORP) indicate that late Alleghanian collision drove rocks of the Carolina terrane, Inner Piedmont, and Blue Ridge several hundred km to the northwest [*Cook and Vasudevan,* 2006; *Duff and Kellogg,* 2017].

Rifting of the orogen began in the late Triassic, followed by seafloor spreading and opening of the Atlantic by 180 Ma. Beneath Line E, rift basin sediments are restricted to the north end, where they are less than 2500 m thick, but are more extensive beneath Line W where they reach thicknesses of 1000-6000 m [*McBride,* 1991]. Subsequent variations in sea level are recorded in the Atlantic Coastal Plain of Georgia and Florida as a sequence of poorly consolidated Cretaceous and Cenozoic carbonates and siliciclastics up to 2000 m thick [*Chowns and Williams,* 1983].

3. Using PKPdf/PKiKP to image P-wave reflectivity

The method used in this study is a modification of an approach known as global phase seismic interferometry or "GloPSI" *[Ruigrok & Wapenaar,* 2012*]*. This approach uses PKIKP (referred to in the rest of this discussion as PKPdf) and PKiKP phases as virtual seismic sources for generating P-wave reflections from the crust and upper mantle. In contrast with other methods for seismic imaging such as receiver functions, GloPSI uses only the vertical component of ground motion. Upon reflection from the earth's surface, these phases reverse polarity and propagate downward as plane waves with near-vertical raypaths. Therefore they preferentially image structure with small dips.

3.1. Processing

Most previous analyses of PKPdf/PKiKP (e.g., *Ruigrok and Wapenaar,* 2012) have taken advantage of the equivalence of the P-wave reflection response (the wavefield recorded for a coincident source and receiver) to the positive lags of the autocorrelation of the transmission response (the wavefield recorded at the surface for a source below the depth range of interest) [*Claerbout,* 1968]. Unfortunately, in the absence of large numbers of earthquakes for stacking, the lags of the autocorrelation corresponding to the early part of the output section tend to be dominated by energy associated with the extended source wavelets.

We try a different approach based on deconvolution of traces prior to stacking using an estimate of the source wavelet for each earthquake. We combine traces from the three profiles (D, W, E) into a single gather and then align first (PKPdf) arrivals by cross correlation. This provides a common time base with the time of the deconvolved first arrival serving as the origin time of reflections. This approach is similar to the method employed by Langston and Hammer [2001] and Yang et al. [2012] for the analysis of teleseismic waves. Following those authors, we estimate source wavelets by stacking seismograms for stations deployed on bedrock, north of the Coastal Plain. The assumption is that lateral variations in structure lead to cancellation of reflections, leaving only the common source wavelet of the earthquake.

This approach can be complicated by the arrival of two phases (PKPdf and PKiKP) in the time window of interest. For a source depth of 100 km and distances of 115°-140°, travel times for the two phases differ by ~ 0-3 s and ray parameters differ by 0.01–0.26 s/°. Free-surface reflections (pPKPdf and pPKiKP) in the source region have ray parameters that are nearly identical to those for the corresponding arrivals PKPdf and PKiKP. For the earthquakes used in this study, the effective array apertures for stations deployed north of the Coastal Plain ranged from 1.7°–2.6° and the differential moveouts ranged from 0.1-0.4 s. The stacking procedure described above treats all these phases as a single arrival and therefore yields an effective source wavelet of extended duration, with some loss of resolution at higher frequencies. Differential moveouts for PKPdf and PKiKP between those stations and the southernmost stations range from 0.1-1.1 s; this causes some broadening of deconvolved waveforms that is minimized by stacking (Figures S8-S9).

Over the distance range 115°-145°, interference from PP, which follows PKPdf by 60 – 180 s, is largely avoided, but earthquakes in the distance range 145°-155° are not included in the analysis because of interference with the phases PKPab and PKPbc. Scattering by lateral heterogeneities near the base of the mantle can also contaminate the record [*Hedlin et al.,* 1997]; this situation is flagged by coherent energy arriving shortly before PKPdf. Beyond 160°, PKPdf and PKPab diverge rapidly, but the listening window is restricted by PKP_{diff}, the diffraction along the inner/outer core boundary, which follows PKPdf by 14-34 s.

3.2 Results

Deconvolution was carried out in the frequency domain. We used a water-level value of 0.0001 to stabilize spectral division and a range of Gaussian functions (α = 1.0 – 4.0) to smooth the output waveforms [Langston, 1979]. Deconvolved gathers for single earthquakes were stacked to suppress noise (both random and coherent) and to enhance signal levels. Prior to stacking, each trace was normalized by the amplitude of the direct PKPdf arrival, then divided by the RMS value of noise 0-18 s prior to PKPdf to give greater weight to seismograms with higher signal levels. After stacking, samples were multiplied by a factor equal to the square root of two-way time to smoothly increase amplitudes of later reflections.

The deconvolved direct arrival at time zero is assigned a positive polarity. As noted earlier, upon reflection at the earth's surface, this polarity is reversed. Therefore the direct arrival and reflections from positive impedance contrasts (e.g., crust over mantle) will show opposite polarities. The 16 earthquakes used in this study are summarized in Table S1. The resulting stacks (Figures 2-4 and S3-S6) show coherent reflections with dominant periods ranging from 1-4 s and from interfaces in the near surface to depths of ~200 km.

3.2.1. Coastal Plain Sediments

For stations over the Coastal Plain, vertical stacks and many of the records for individual events show a series of strong, coherent reflections 0.2-4.2 s after the direct arrival (Figures 2 and S4). Travel times for the peak of the first half cycle (0.2-1.4 s; opposite in polarity from PKPdf, indicating a positive impedance contrast) match those predicted for reflection from the base of the sequence of Cretaceous and younger sediments and poorly consolidated sedimentary rocks. This event starts very close to the feather edge of the sequence at the northern boundary of the Coastal Plain and increases in travel time to the south, in agreement with thicknesses derived from well data [Chowns and Williams, 1983]. For Line E (Figure S4), travel times are best fit by an average P-wave velocity of 2400 m/s for stations E31 to E25 and by a slightly greater average velocity (2600 m/s) for the thicker sequences beneath stations E24-E07. For Line W (Figure 2), the best fits are obtained for velocities of 2200 m/s for stations W21-W19, 2400 m/s for stations W18-W10, and 2100 m/s for stations W09–W01. These values are consistent with those reported by Iverson and Smithson [1983] and Barnes and Reston [1992].

Travel times for the later 2-3 half-cycles are consistent with times predicted for free-surface multiples. Travel times for the second multiple bounce overlap those predicted for the base of the underlying Triassic/Jurassic sedimentary sequence; resolution of the latter will require additional deconvolution (work in progress).

3.2.2. Moho

For Line W, events interpreted as Moho reflections are clearest at lower frequencies (α=1.0; Figure 2). Two-way times vary between 9.7 and 10.1 s for stations W02 to W06 and between 10.0 and 10.9 s for stations W07 and W22, all within the Coastal Plain. Moho reflection times reported for coincident COCORP lines [*McBride,* 1991] are somewhat greater (11.2 – 12.0 s), suggesting that the broader PKPdf-generated waveforms may represent a composite of reflections from lower crustal layering and the crust-mantle transition itself (work in progress). A similar result is obtained for Line E (Figure S3). From station W22 the times increase northward, from 12.2 s at station W23 to 12.8 s (crustal thickness: 42 km) at station W31 (Inner Piedmont), with a pronounced increase from 13.8 s (45 km) at W315 (foothills) to 17.4 s (57 km) beneath W35 in the Blue Ridge Mountains. A similar trend is observed for Line D (Figure S2). These are the first observations of normal-incidence P-wave reflections from the Moho beneath the higher elevations of the southern Appalachians. The results for Lines W and D are consistent with other broadband and active-source results suggesting that the southern Appalachians are in rough local isostatic equilibrium [*French et al.,* 2009; *Hawman et al.,* 2012; *Schmandt et al.,* 2015; *Parker et al.,* 2013; 2016; *Hopper et al.,* 2016].

3.2.3. Upper Mantle

Stacks for Line E show reflections at 16-20 s (depths: 50-75 km; Figures 3 and S4). A similar reflection sequence is not seen in the stacks for Line W, but reflections in this depth range do appear in sections derived by pre-stack migration (Figure S10; see below). Models of S-wave velocity for the southern Appalachian highlands, derived by inversion of surface waves [*Savage et al.,* 2017], show a low-velocity zone over a similar depth range. Similar "mid-lithospheric discontinuities" occur throughout the North American craton and are variously interpreted as low-velocity cumulate layers, refertilized depleted mantle, stacks of oceanic plates, and underplating [*Abt et al.,* 2010; *Calo et al.,* 2016]. Alternatively, the reflections could be generated by layers within the uppermost mantle depleted by partial melting during Mesozoic extension [*Pollitz & Mooney,* 2016].

For Lines E and W, multicyclic reflections between 32 s and 36 s are consistent with a layered zone at depths of 120 - 135 km (Figures 3, 4, S4, S5). This is roughly 30 km deeper than the lithosphere/asthenosphere boundary (LAB) inferred from analysis of Sp receiver functions for a nearby broadband station [*Abt et al.,* 2010] but is in better agreement with more recent studies based on inversion of surface waves [*Savage et al.,* 2017] and joint inversion of surface waves and Ps receiver functions [*Calo et al.,* 2016].

The frequency content and multicyclic character of the reflections indicates much smaller wavelength variations in velocity than expected for a purely thermal boundary. At these depths, quarter-wavelength estimates of vertical resolution, corresponding to layer thicknesses required for constructive interference of reflections from multiple layers, range from 3-8 km (Supplement). This scale of layering would be consistent with models incorporating the effects of shearing at the transition and also with intrusions triggered by partial melting of hydrated asthenosphere [*Fischer et al.,* 2010*; Till et al.,* 2010]. Alternatively, these and deeper signals at roughly 44 s (~167 km) and 58 s (~ 224 km) (Figure 4) may be reflections from layering associated with drag-induced flow in the asthenosphere [*Eaton et al.,* 2009*; Fischer et al.,* 2010*; Long et al.,* 2016]. The event at 58 s is close in depth to a transition identified beneath North America as the Lehmann discontinuity [*Calo et al.,* 2016]; this feature was originally interpreted as the base of a low-velocity zone [*Lehmann,* 1960] but could also represent an abrupt decrease in transverse anisotropy marking the base of a zone of strong coupling between asthenosphere and lithosphere [*Gaherty and Jordan,* 1995; *Eaton et al.,* 2009]. These deeper reflections are consistent with largely horizontal rather than vertical flow at depths less than 225 km beneath this portion of the Georgia Coastal Plain.

3.2.4. Individual Events

The above approach is capable of recovering useful signal levels for single earthquakes, even without stacking. Deconvolved records for individual events (Figure 5) show consistent results for earthquakes over a range of magnitudes and focal depths, indicating that major arrivals (particularly those interpreted as reflections from the LAB at 32-36 s) are indeed reflections and not artifacts generated by incomplete deconvolution of source-side scattering and other coherent noise (Figures S6 and S7).

However, the results for individual earthquakes do show some variations in arrival times and relative amplitudes for major arrivals, which tend to degrade the stacks (e.g., the events at 32-34 s in Figure 4), and also show variations in reflection density over different time windows. We attribute these variations to differences in dominant frequencies for different magnitude earthquakes, processing artifacts, particularly near the end of listening windows (Figure S6 and S7), small-scale lateral heterogeneity, and variations in back-azimuth. To define major structural boundaries, we rely on the stacked images. To preserve a composite image of finer-scale structure or "fabric", we migrate the deconvolved sections for individual earthquakes separately and stack the results. Preliminary migration results based on a subset of events are presented in the Supporting Information (Figures S10 and S11).

4. Conclusions

Reflection profiles generated using PKPdf as a virtual seismic source show laterally continuous structures at depths from the near-surface to roughly 200 km beneath the southern Appalachian orogen and adjacent Atlantic Coastal Plain. Waveforms are deconvolved for each earthquake separately, using estimates of the source wavelet derived by stacking over multiple stations. The results for multiple earthquakes then are stacked to form composite images of the crust and uppermost mantle.

The profiles show clear, continuous signals with times in close agreement with those expected for primary and multiple reflections from the base of coastal plain sediments. Arrivals interpreted as reflections from the Moho increase in travel time from ~10 s beneath the Coastal Plain to 17.4 s (~57 km) beneath the Blue Ridge Mountains, in agreement with trends observed for wide-angle reflections and Ps and Sp arrivals, and provide additional evidence that the Appalachian highlands are in rough isostatic equilibrium. Reflections at 16-20 s (50-75 km) beneath the coastal plain are consistent with layering associated with depletion of the uppermost mantle during Triassic rifting. Reflections at 32-36 s (120-135 km) are roughly 30 km deeper than previous estimates of LAB depth based on Sp receiver functions, but are consistent with the depth to the LAB found in more recent inversions of Ps arrivals and surface waves. Alternatively, these and later reflections at roughly 44 s (~167 km) and 58 s (~224 km) may be due to layering associated with drag-induced flow in the asthenosphere, suggesting largely horizontal rather than vertical flow for depths less than 225 km beneath the Georgia coastal plain.

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Data Sources

The waveform data used for this study can be accessed from the IRIS Data Management Center at http://www.iris.edu/SeismiQuery. The network code for the SESAME experiment is "Z9". IRIS Data Services are funded through the Seismological Facilities for the Advancement of Geoscience (SAGE) Award of the National Science Foundation under Cooperative Support Agreement EAR-1851048.

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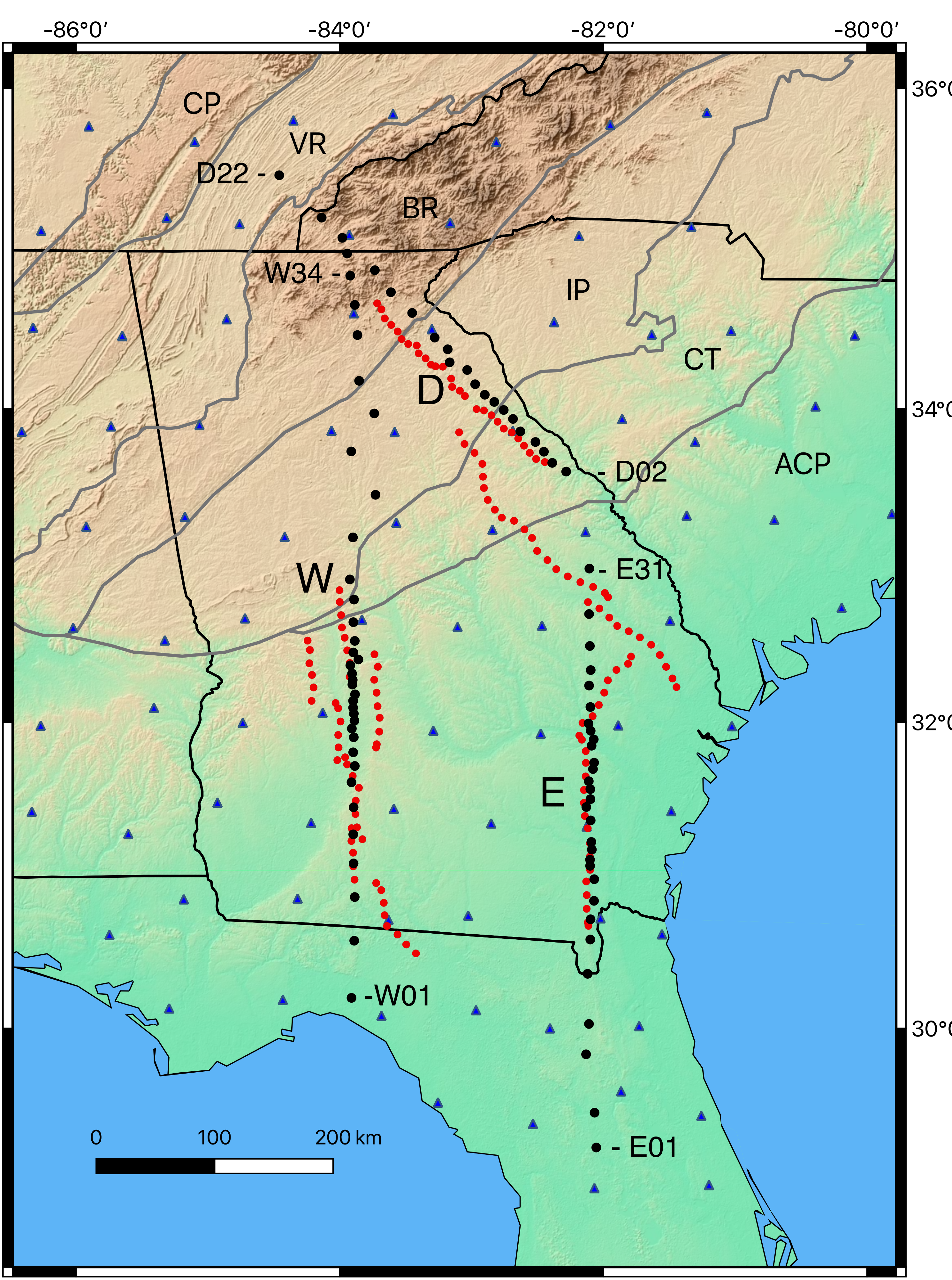

-86°0′
-84°0′
-82°0′
-80°0′
36°0′
34°0′
32°0′
30°0′
CP
VR
BR
D22 -
W34 -
IP
CT
D
- D02
ACP
W
- E31
E
-W01
- E01
0
100
200 km

Figure 1

Figure 1. Map of the study area. Blue dots: SESAME stations along lines W, D, and E. Blue triangles: TA stations. Red dots: subset of stations for overlapping COCORP lines.

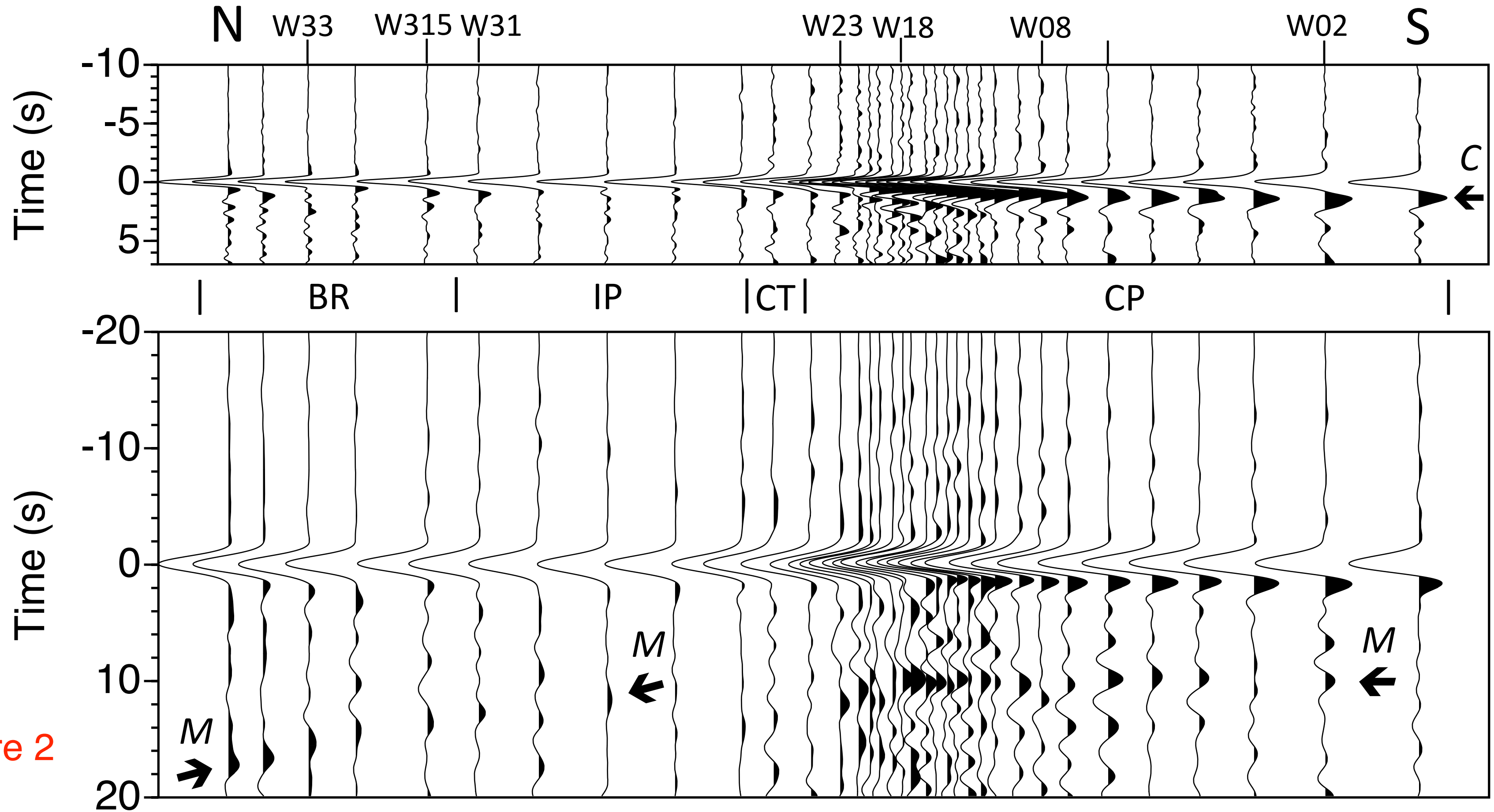
N
W33
W315
W31
W23
W18
W08
W02
S
Time (s)
-10
-5
0
5
C
BR
IP
CT
CP
-20
-10
0
10
20
M
M
M
M

Figure 2

Figure 2. Stacks of deconvolved records for 16 earthquakes (Table S1) showing PKPdf-generated reflections beneath SESAME Line W, using Gaussian smoothing parameters α=3.0 (top) and 1.0 (bottom). Plotted with reverse polarity (negative polarity for the direct PKPdf arrival at 0 s). Top: “C” is interpreted as the reflection from the base of Cretaceous-Tertiary sediments and poorly consolidated sedimentary rocks. Times are consistent with well data. This reflection merges with the direct arrival near the northern edge of the Coastal Plain. Bottom: “M” is interpreted as the reflection from the Moho, increasing in two-way time from roughly 10 s beneath the Coastal Plain to 17.5 s beneath the Blue Ridge Mountains. A similar trend in crustal thickness is seen at higher frequencies for Line D (Figure S2).

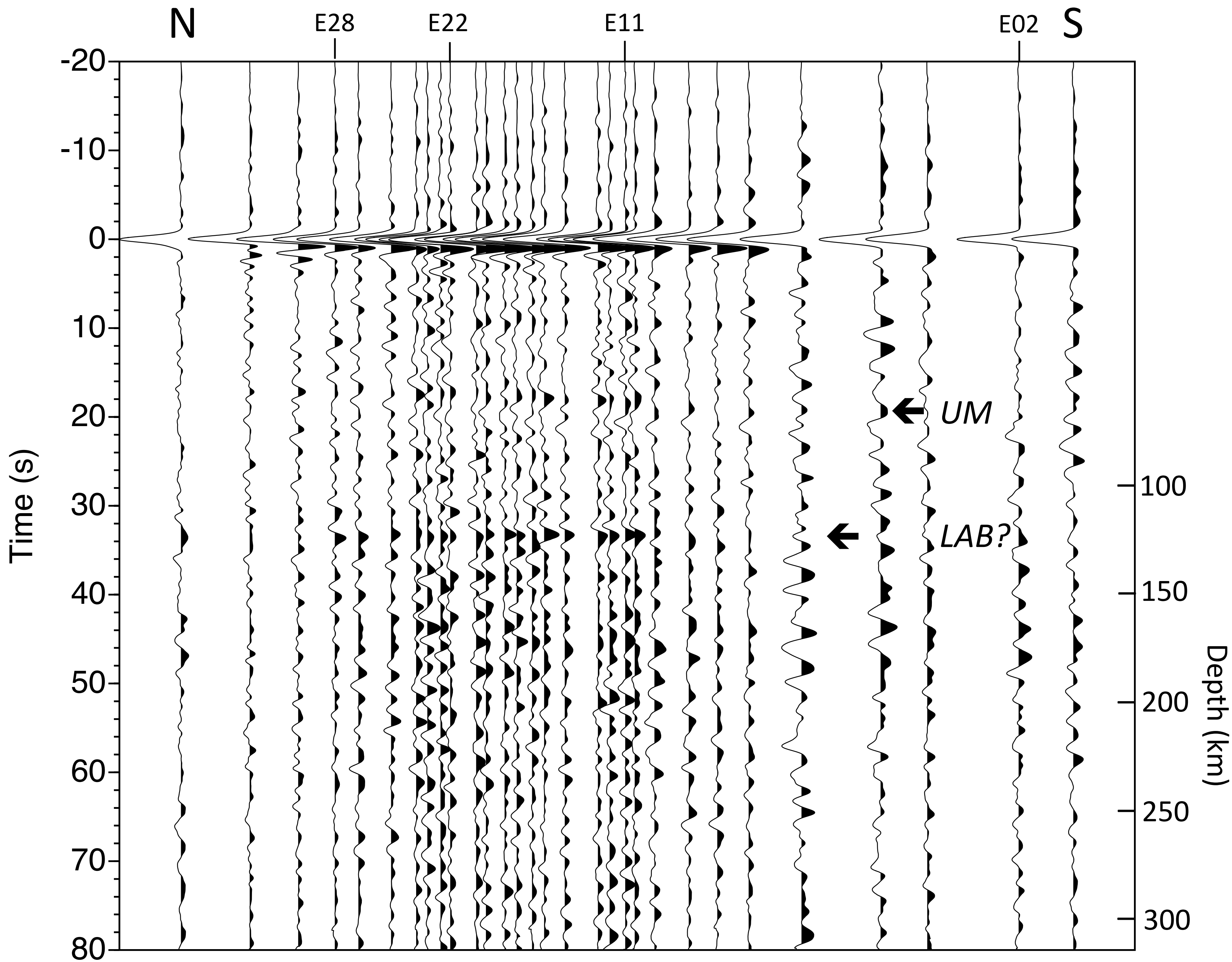
N
E28
E22
E11
E02
S
-20
-10
0
10
20
30
40
50
60
70
80
Time (s)
UM
LAB?
100
150
200
250
300
Depth (km)

Figure 3

Figure 3. Stack of deconvolved records (α=2.0) for 6 earthquakes (Table S1) showing reflections beneath SESAME Line E, plotted with reverse polarity. Depths are approximated using a laterally uniform velocity model (crustal thickness: 55 km; average velocities: 6.5 and 8.1 km/s for the crust and upper mantle, respectively). This approximation contributes 0.5-2.5 km to uncertainties in depth within the mantle. UM: upper mantle reflections. Multicyclic reflections observed at 34–36 s (~127-135 km) may mark the effects of shearing in the vicinity of the LAB and/or intrusions triggered by partial melting of hydrated asthenosphere.

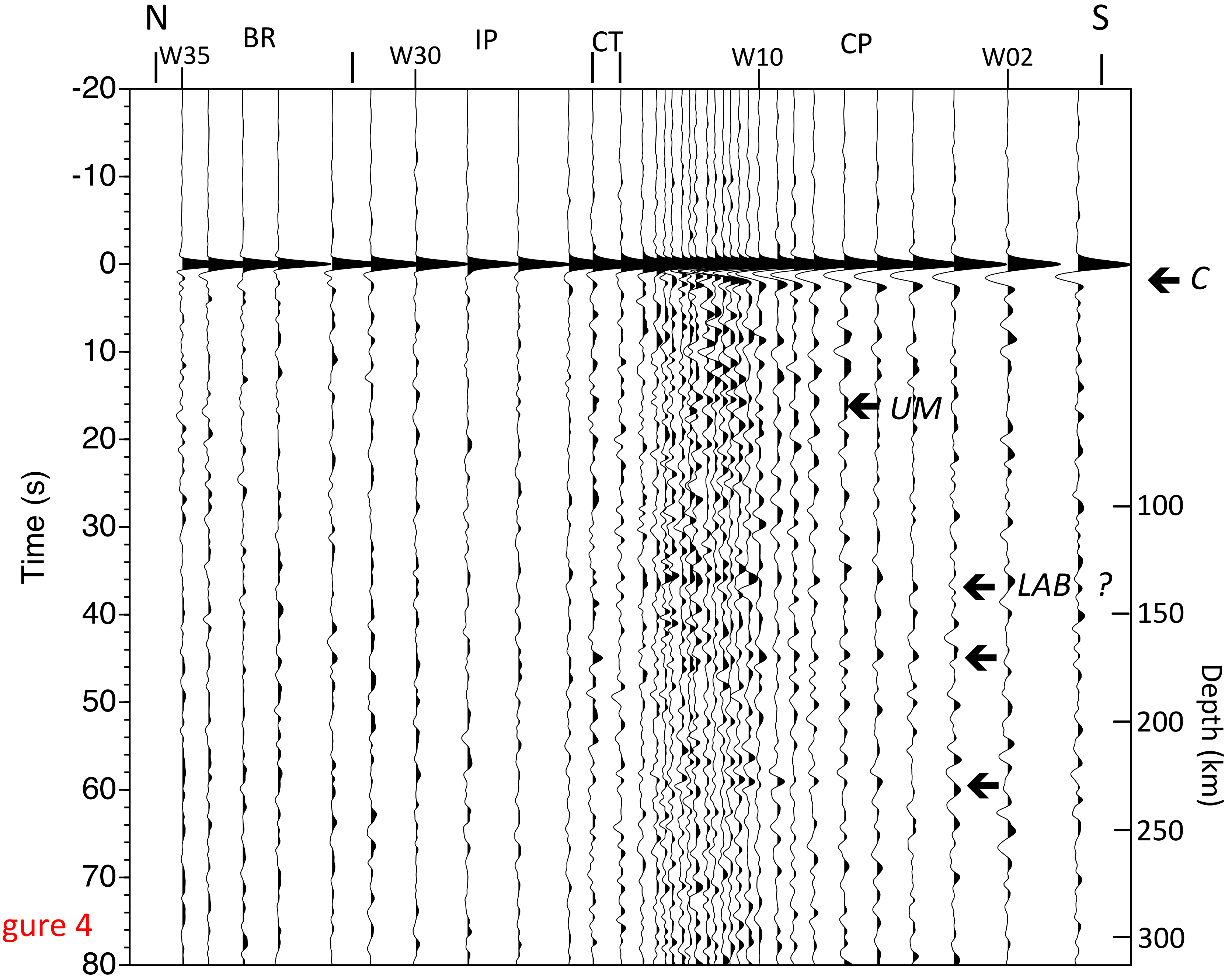
N
S
W35
BR
W30
IP
CT
W10
CP
W02
C
UM
LAB ?
Time (s)
Depth (km)
-20
-10
0
10
20
30
40
50
60
70
80
100
150
200
250
300

Figure 4

Figure 4. Stack of deconvolved records (α=2.0) for 16 earthquakes showing reflections beneath SESAME Line W, plotted with normal polarity (see also Figure S5). C: reflection from the base of Cretaceous and Tertiary sediments and poorly consolidated sedimentary rocks (see also Figure 2). The arrival at 35 s (depth approximately 130 km) is in close agreement with arrivals interpreted as the LAB in Figure 3. Later arrivals at roughly 44 s and 58 s (arrows; ~167 km and 224 km) are interpreted as reflections from layering within the asthenosphere.

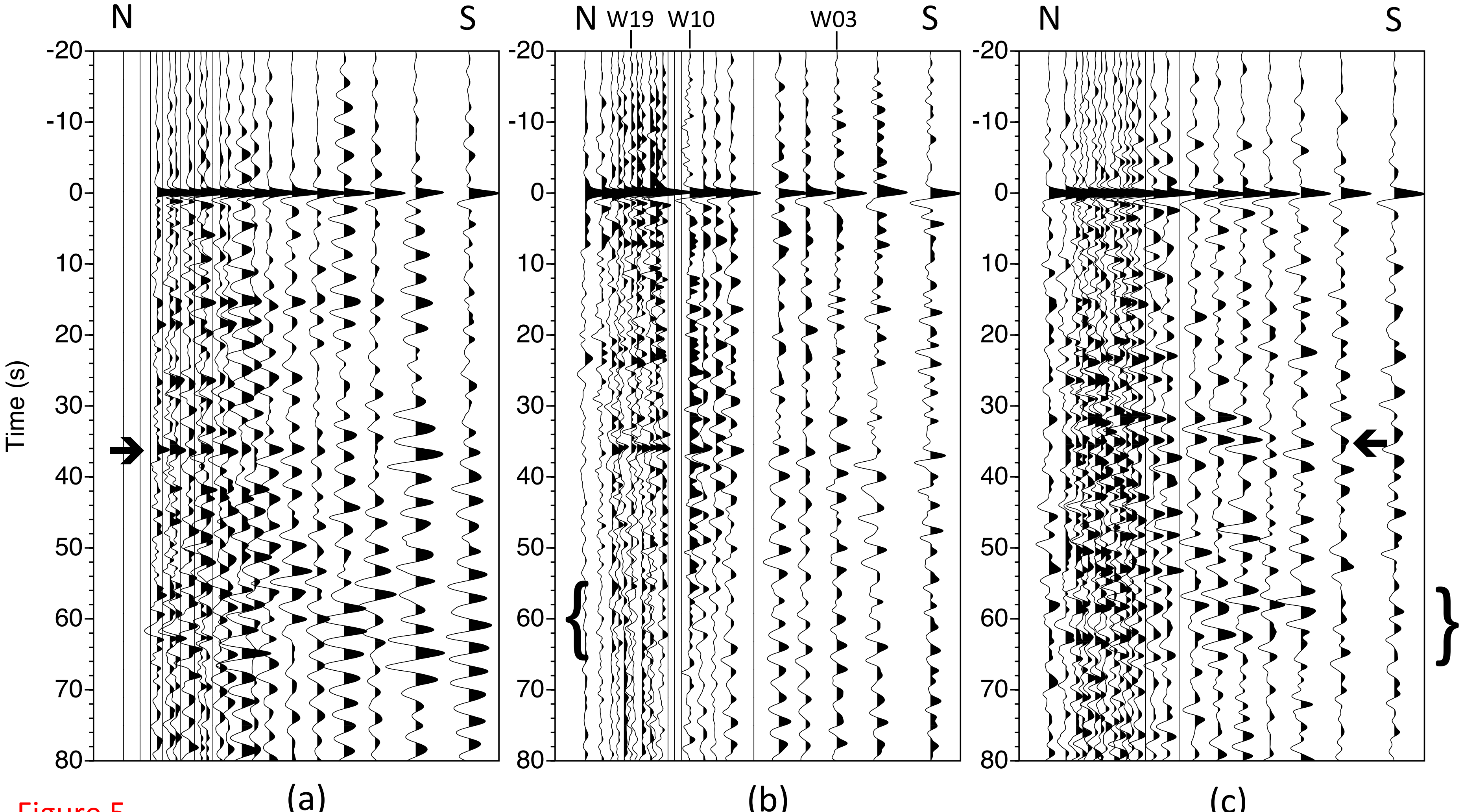

Figure 5

Figure 5. Unstacked, deconvolved sections (α=2.0) for the southern half of SESAME Line W, plotted with normal polarity. Note the agreement in travel time for events at approximately 32-36 s (depth: 120-135 km), interpreted in Figure 4 as reflections from the LAB. See also Figures S6 and S7.

(a) Event 1: Mw=7.3, depth=386 km, Δ = 118°– 120°, back-az = 284°–286°.
(b) Event 5: Mw=6.6, depth=20 km, Δ = 125°– 127°, back-az = 290°– 293°.
(c) Event 9: Mw =6.7, depth=18 km, Δ = 129°– 132°, back-az = 323°- 325°.