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Earthquake rupture below the brittle-ductile transition in continental lithospheric mantle

Germán A. Prieto,1* Bérénice Froment,2 Chunquan Yu,3,4 Piero Poli,3 Rachel Abercrombie5

Earthquakes deep in the continental lithosphere are rare and hard to interpret in our current understanding of temperature control on brittle failure. The recent lithospheric mantle earthquake with a moment magnitude of 4.8 at a depth of ~75 km in the Wyoming Craton was exceptionally well recorded and thus enabled us to probe the cause of these unusual earthquakes. On the basis of complete earthquake energy balance estimates using broadband waveforms and temperature estimates using surface heat flow and shear wave velocities, we argue that this earthquake occurred in response to ductile deformation at temperatures above 750°C. The high stress drop, low rupture velocity, and low radiation efficiency are all consistent with a dissipative mechanism. Our results imply that earthquake nucleation in the lithospheric mantle is not exclusively limited to the brittle regime; weakening mechanisms in the ductile regime can allow earthquakes to initiate and propagate. This finding has significant implications for understanding deep earthquake rupture mechanics and rheology of the continental lithosphere.

INTRODUCTION

The role of temperature in the rheology of the lower crust and lithospheric mantle is now well understood, especially in oceanic lithosphere. Earthquakes occur throughout the oceancrust and upper mantle, the latter with olivine-dominated lithology; the brittle-ductile transition is associated with a threshold temperature of about 600°C (1–4). Upper-mantle seismicity is less common in the continental lithosphere. Earthquakes tend to occur mostly in the upper crust and, in some cases, in the uppermost mantle (5). For the latter case, olivine-dominated lithology predicts brittle failure at temperatures below ~600°C (3). It has been proposed (5, 6) that the rare continental lithospheric earthquakes occur in cold and anhydrous mantle rocks (T < ~700° ± 100°C). Earthquakes located below the Moho may thus indicate a strong, seismogenic upper mantle that can sustain large stresses, which are later released during brittle rupture (5–7).

An isolated earthquake with a moment magnitude (Mw) of 4.8 occurred deep in the continental lithosphere on 21 September 2013 near the Wind River Range in central Wyoming at a depth of 75 ± 8 km (8). The focal mechanism indicates a dominant strike-slip faulting mechanism with a small reverse-slip component (9). It was followed within 2 hours by a single M 3.0 aftershock at a similar hypocentral depth. The Wyoming earthquake and its aftershock are unusual because they occur deep in the lithospheric mantle, far from any convergent plate boundary (Fig. 1). A few previous earthquakes have occurred in the region at a similar depth (10), including a M 3.8 earthquake in 1979 beneath Randolph, UT (11), but the 2013 Wyoming earthquake is by far the best recorded. The complexity of the velocity structure in this transitional region and the large uncertainties in the temperature in the hypolithosphere regions meant that previous studies (8, 9, 11) could exclude the possibility that these earthquakes represent brittle rupture in cold mantle.

Here, we analyze broadband seismograms at regional distances to determine the static and dynamic source parameters of the 2013 Wyoming earthquake and for clues concerning the rupture conditions. Estimates of the seismic radiation, the source dimension, and the amount of slip of an earthquake can be used to investigate the stress drop and fracture energy (12). We also perform detailed thermal modeling to ascertain whether the Wyoming earthquake occurred below or above the brittle-ductile transition temperature.

RESULTS

Earthquake source analysis

We use the M 3.0 aftershock as our empirical Green’s function (EGF) to remove propagation effects from the recorded seismograms and to obtain the relative source time functions (STFs) and spectral ratios (see Materials and Methods). The STFs at most of the stations show a clear source pulse, with consistent shapes between nearby stations (Fig. 1). Figure 1 (B and C) shows a clear azimuthal dependence of the width of single-station P- and S-STFs, which reflects predominant rupture propagation along the subvertical northwest-oriented plane; stations aligned along the rupture direction have narrower pulses, whereas stations away from the rupture direction have broader pulses. This angular dependence of apparent duration can be quantified using a directivity function D, which depends on the rupture velocity \( V_R \), the wave velocity around the source \( c \), and the angle between the ray takeoff angle and the rupture direction \( \phi \) at station \( i \)

\[
T_{app} = \frac{T_R}{D_i(V_R, c, \phi_i)}
\]

where \( T_{app} \) and \( T_R \) are the apparent and the actual rupture durations, respectively. Because we have both P and S measurements, we can combine them to obtain the length of the rupture without assuming the rupture duration. We stretch the STFs in the time domain according to Eq. 1, varying the rupture velocity around the source \( c \), and the angle between the ray takeoff angle and the rupture direction \( \phi \) at station \( i \)

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those commonly assumed (~0.8V_s to 0.9V_s) in source studies (Fig. 2D). We sum the single-station STFs corrected for directivity effects using our estimated $V_R$ (Fig. 2C) to improve the signal-to-noise ratio (SNR) and to obtain clear and unbiased, average $P$- and $S$-stacked STFs (Fig. 2B). From these STFs, we calculate a model-independent rupture duration $T_R = 0.47 \pm 0.04$ s and hence the length of the rupture $L = V_R \times T_R \approx 610$ m.

Calculation of the earthquake stress drop (proportional to slip/length or seismic moment/area\(^{2/3}\)) is source model–dependent. We use the global centroid moment tensor seismic moment $M_0 = 2.17 \times 10^{16}$ N·m (www.globalcmt.org). If we assume a square fault with a length of 610 m, we obtain $\Delta \sigma \approx 60$ MPa, while a circular fault with a radius of 305 m yields $\Delta \sigma \approx 330$ MPa (see Materials and Methods).

**Fig. 1. Earthquake location and $P$- and $S$-source time functions.** (A) Map of the seismic stations (triangles) used here and location of the 21 September 2013 Wyoming earthquake (star) and its focal mechanism (beach ball). Colored triangles stand for seismic stations used in the EGF procedure. Black triangles show the extended set of stations used in the radiated energy estimate. Inset shows the study area (red box) as well as the epicenter of the earthquake (red star). Single-station STFs obtained from the EGF procedure on $P$ wave (B) and $S$ wave (C). STFs are sorted as a function of station azimuth from the fault strike. Waveform colors correspond to those of the stations in (A). The gray dashed lines are the predicted widths of the $P$- and $S$-STFs (see main text).

**Fig. 2. Rupture directivity and velocity estimate.** (A) Contours of normalized variance reduction as a function of rupture velocity $V_R$ and percent unilateral rupture values $\epsilon$. The variance of the best-fit model in this plot is set to 1. Dark blue contours indicate variability within 10%, and the dashed line shows the best-fit rupture velocity. (B) Stacked $P$- and $S$-STFs resulting from the correction using our best-fit rupture velocity $V_R = 1.3$ km/s. (C) Single-station $S$-STFs corrected for directivity effects using the stretching method with $V_R = 1.3$ km/s. (D) Same as (C), with $V_R = 3.8$ km/s. Red waveform at the bottom of each figure is the stack of all the corrected STFs. This correction involves stretching the STFs at station $i$ with a factor of $1/D_i$. 


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average corner frequency (~2 Hz) would give a stress drop of ~50 to 150 MPa, depending on source model assumptions (see fig. S2). All these values are relatively high compared to those of shallow earthquakes (14) but similar to those of other intermediate-depth earthquakes (12, 15) of similar size.

We calculate the earthquake-radiated energy \( E_S \) by averaging frequency-domain estimates from \( P \) and \( S \) waves at all available stations (see Materials and Methods; Fig. 1A). We obtain a total radiated seismic energy \( E_S = 3.9 \times 10^{12} \) J. The scaled energy \( E_S/M_0 \) is relatively high \((1.7 \times 10^{-4})\), and the apparent stress, defined as \( \sigma_a = \mu E_S/M_0 \), is equal to ~12 MPa, where \( \mu \) is \( 7 \times 10^{10} \) N m\(^{-2}\). The stress drop obtained from the directivity modeling (60 to 330 MPa) is at least five times larger than the apparent stress and the radiation efficiency \( R \) is less than unity. We compare this value to the radiation efficiency expected from our estimated \( V_R \) using crack theory, resulting in our preferred radiation efficiency \( R \approx 0.4 \). This value is smaller than those of most shallow earthquakes, similar to those of intermediate-depth earthquakes in Japan and Mexico (15, 16), and larger than those of other deep earthquakes (12, 17).

In summary, the deep Wyoming earthquake has an anomalously low rupture velocity, a high stress drop and scaled radiated energy, and a low to average radiation efficiency. We interpret this event to represent a dissipative source mechanism that was slow to propagate with a significant portion of the available potential energy used in or around the source region.

**Temperature modeling**

Knowledge of the temperature at the hypocentral depth of the Wyoming earthquake is necessary to determine whether the earthquake nucleated above or below the expected brittle-ductile transition. Surface heat flow measurements in the Wyoming Craton range from 40 to 60 mW m\(^{-2}\). Temperature estimates at the hypocentral depth of the earthquake vary from ~600° (8) to more than 800°C (18), but large uncertainties remain. The earthquake is located along a strong lateral gradient of shear wave velocities, similar to other lithospheric mantle earthquakes (11, 19) as well as in a transition of lithospheric thickness (20, 21). In addition, estimated temperatures at depth based on xenoliths in the Wyoming area suggest a fairly warm lower crust (18, 22), and recent volcanism in Leucite Hills (about 100 km to the south and <3 million years ago) may require high temperatures in the upper mantle (23).

We estimate the depth-dependent temperature profile using the grain-size viscoelastic relaxation model of olivine in the mantle (24) to predict the shear wave velocities at depth and compare to tomographic models [(25, 26); see Materials and Methods]. The best-fitting geotherms (Fig. 3) predict temperatures at the hypocentral depth of the Wyoming earthquake that are substantially higher than the 600°C transition from velocity weakening to velocity strengthening of olivine extrapolated from laboratory experiments (1). In common with other studies (18, 22, 23), we favor the warmer geotherms in Fig. 3 because those that predict \( T < 750°C \) at the earthquake hypocenter also predict low temperatures down to 400 km, in disagreement with lithosphere-asthenosphere boundary depth estimates (20, 21). A similar analysis of the hypocentral region of the 1979 Randolph, UT earthquake predicts even higher temperature ranges (see fig. S1).

The temperature modeling implies that the Wyoming earthquake occurred in the upper range of the brittle regime or in the ductile regime. The source modeling found a slow rupture with a dissipative rupture process, which is more consistent with a ductile regime. Our estimates of rupture velocity are extremely low, similar to that of the Bolivia earthquake (17) or the initial rupture of intermediate-depth earthquakes (27, 28), and are lower than oceanic lithospheric mantle earthquakes (29–31).

**DISCUSSION**

Previous studies of the 2013 Wyoming earthquake focused on its location and orientation (8, 9). The focal mechanism shows similar stress orientation to that of the shallow seismicity, suggesting that they are responding to the regional stress regimes (8, 9). The location of the earthquake within a high-velocity region (32), combined with the large uncertainties in the temperature profile, meant that previous studies were unable to exclude the possibility that it was caused by brittle failure in cold, anhydrous mantle. More detailed modeling of the thermal

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**Fig. 3. Temperature modeling of the Wyoming earthquake.** (A) The \( V_R \) model (25) in the western United States at a depth of 75 km and east-west cross sections showing the location of the Wyoming earthquake (star) in the strong velocity gradient. (B) Geothermal models (left; inset shows a zoom around the hypocenter) and predicted shear wave velocities using the approach of Faul and Jackson (24) that best fit the \( V_R \) tomography models (right) of Schaeffer and Lebedev (26) (blue area) and Shen et al. (25) (red area). The geothermal models assume a crustal thickness of 50 km and produce surface heat flows between 40 and 60 mW m\(^{-2}\). The range of temperatures at the hypocentral depth is from 750° to 850°C in the upper range of the brittle-ductile transition.

structure of the lithosphere suggests that temperatures in the hypo-
central region are quite high (18), regardless of the presence of the
high-velocity region (32), and our analysis suggests that \( T > 750^\circ C \)
(Fig. 3). Our source analysis found that the earthquake had low radia-
tion efficiency, implying that a large amount of energy was dissipated
during rupture, perhaps as frictional heating (12, 17), in contrast to
brittle-like, fast rupture of other deep earthquakes (33). The additional
lines of evidence presented here raise questions about the brittle nature
of the 2013 Wyoming earthquake and potentially of other continental
lithospheric mantle earthquakes elsewhere.

We hypothesize that an ongoing ductile deformation in the mantle,
for example, along a midlithospheric discontinuity (22), may lead to
shear band formation and slip concentration along subvertical zones.
The earthquake may have nucleated along one of these shear zones
through a thermal shear runaway mechanism or aided by dehydration
reactions, triggering the \( M_w \) 4.8 earthquake and its only aftershock
(12, 27, 28). The 1979 deep Randolph earthquake probably nucleated
in a similar manner (see Fig. S1 for temperature modeling).

Another possible explanation is that compositional or mineralogical
heterogeneity in the mantle [for example, grain-size heterogeneity (34–36)]
can promote brittle behavior or instability in a small volume, wherein
the earthquake nucleates and then propagates into the mostly ductile
mantle rocks. This mechanism has been proposed for the Newport-
Inglewood Fault (37), where deep earthquakes are attributed to a system
of seismic asperities in a ductile fault zone. Nucleation may additionally
be encouraged by the presence of fluids or fluid migration in the mantle,
either by reducing the effective normal stress or promoting strain local-
ization along the shear zones (36, 37).

Source analysis of oceanic lithospheric earthquakes off the shore of
Sumatra show deep slip in the hotter, velocity-strengthening region
(30, 38), but this deep slip is only observed in the greatest earthquakes
(39). Aseismic creep is also observed at shallow depths, and the small
repeating earthquakes at Parkfield, CA are thought to involve slip in
both velocity-weakening and velocity-strengthening regions (40). In
contrast to both of these cases, the nature of rupture in the Wyoming
earthquake implies that the rupture was dominated by slip in the more
ductile, aseismic region.

The presence and mechanisms of earthquakes in the continental
lithospheric mantle have been a highly controversial topic in geophysics
(3, 5–9, 11, 19). Our results on the 2013 Wyoming earthquake shed new
light on the rupture mechanism of these rare earthquakes (12, 17, 28)
and provide important constraints on rheological properties of the
lithospheric mantle (5, 6, 19, 41).

**MATERIALS AND METHODS**

**EGF method**

We studied the source parameters using the EGF method (42). This
approach used a smaller event that is collocated with the earthquake
of interest to remove the path, site, and instrumental effects from the
earthquake source signal in the seismic recording (43). Fortuitously,
the single aftershock was an ideal EGF candidate. Although its location
is less well constrained than that of the main shock, their waveform si-
milarity suggests a similar location, orientation, and depth (1), making
these events ideal candidates for a good EGF earthquake pair (44).

We used raw seismograms recorded in a 10° × 10° area around the
epicenter zone of the two earthquakes, shown as colored triangles in Fig.
1A. We performed the analysis on vertical (P waves) and radial and
transverse (S waves) seismograms. After the EGF procedure (43), we
computed the earthquake STF using the multitaper deconvolution
algorithm (45) at stations for which the cross-correlation coefficient
between the two event waveforms is \( \geq 0.6 \) for P waves and \( \geq 0.5 \)
for S waves.

**Directivity from STFs: Cross-correlating stretched waveforms**

Because some earthquakes exhibit bilateral rupture, we used the direc-
tivity function (13)

\[
D_i = \frac{V_R T_R \sqrt{(1 + c^2)(1 + \zeta_i^2) + 4c \zeta_i}}{(1 - \zeta_i^2)}^{1/2}
\]

where \( 0 \leq c \leq 1 \) is the “percent unilateral rupture” that reflects the
degree of unilateral rupture, and the notation \( \zeta_i = V_R/c \cos(\phi_i) \) is
introduced to simplify the expression.

Static source parameters are usually estimated by modeling the
source signal with simple source models (46). However, earthquake
ruptures may be much more complex than models suggest, and signif-
ificant azimuthal variation of the spectral characteristics of the recorded
signals is expected (47, 48). We proposed here to use the directivity
observations to estimate the rupture velocity and duration without
assuming any particular source model. Our technique was based on
the idea that directivity effects appear as stretching or compression of
the waveforms. In view of the different velocities of P and S waves, di-
rectivity effects were expected to be stronger on S wave recordings.
As a result, the S-STF at a given station \( i \) turns out to be a stretched (or
compressed) version of the P-STF. Following Eq. 1, we could write the
ratio between single-station S- and P-STFs apparent durations \( T_{SP} \) as

\[
\frac{T_{SP}}{T_{PP}} = \frac{D_i (V_R, V_P, \phi_i)}{D_i (V_R, V_S, \phi_i)}
\]

This ratio removed any dependence on \( R \) and could be used to
estimate the rupture velocity \( V_R \).

In practice, we measured the ratio between S- and P-STFs durations
\( T_{SP} \) at each station as the factor \( c \), where the time axis of S-STF has to be
stretched or compressed to obtain the best correlation with P-STF. We
then performed a grid search on \( V_R \) to look for the rupture velocity that
best explains our measured \( c \) over the set of stations, with the model in
Eqs. 1 and 2. Our results showed that a slow unilateral rupture is re-
quired to explain the data (Fig. 2A). The best fit between the predicted
and the measured \( T_{SP} \) was obtained for \( e \approx 1 \) and \( V_R = 1.3 \text{ km/s} \)
\((-0.29 V_S)\).

This measurement of \( V_R \) can then be used to correct the directivity
effect on the individual STFs, as expressed by the actual source pulse
observed at each station. Following our directivity model in Eq. 1, this
correction came down to stretching the STFs at station \( i \) with a factor of
\( 1/D_i \) using the values that best fit the data (that is, \( e = 1 \) and \( V_R = 1.3 \text{ km/s} \)).
Figure 2C shows the resulting single-station S-STFs corrected for di-
rectivity effects. In comparison, Fig. 2D shows the corrected single-station
S-STFs, assuming the widely used value of \( V_R = 0.9V_S \) (12, 14, 15, 33).
In the latter case, we clearly saw that STFs at stations aligned in the direc-
tion of the rupture are overcorrected, implying that the actual rupture
velocity was much lower than \( 0.9V_S \). This test definitively confirmed
that a slow rupture velocity is required to explain our data.

We determined that rupture duration is 0.44 and 0.51 s on the
stacked P- and S-STFs, respectively. Estimation of the width of the resulting
stacked STFs was straightforward and did not rely on a particular source
model, as is usually the case when interpreting source spectra. The rupture
duration was 0.44 and 0.51 s on the stacked P- and S-STFs, respect-
ively, which led to an estimate of $T_p = 0.47 \pm 0.04$ s. Note that we also
investigated the time resolution limit by computing the delta function
(that is, the deconvolution of the earthquake by itself) at each station. All
the resulting delta functions were narrower than 0.1 s. The pulse dura-
tion of single-station STFs are always $>0.2$ s, confirming that our STFs
are all well above the time resolution limit.

**Radiated seismic energy**

To improve the stability of our energy estimate, we extended the num-
ber of stations used in the estimate, including colored and black trian-
gles in Fig. 1. We corrected the instrumental response at each station and
the propagation effect using a frequency-independent quality factor
($Q = 1000$). We estimated the quality factor $Q$ that best explains the EGF
spectral ratio in the previous step, with values ranging from 500 to 1500.
Finally, we performed the radiated energy analysis on almost 70 stations
compared to 30 single-station EGF estimates (see Fig. 1A).

We calculated the radiated seismic energy at each station in the fre-
quency domain by integrating the velocity spectrum from both P and S
wave ground motions for the three components, which are summed to
obtain the total seismic energy ($\mathcal{E}$). We took radiation pattern correc-
tions into account but only used stations with radiation pattern coeffi-
cients of 0.2 or larger in our energy estimates ($\mathcal{E}$) to limit the influence
of stations near the nodal plane.

Estimates of radiated energy required the integration over a wide
frequency band, but the SNR limited our observable band to about
0.5 to 20 Hz; thus, we extrapolated the velocity spectrum, assuming a
Brune spectral model ($\propto f^{-2}$ falloff) at both low and high frequencies.
When not all of the components were available, we set them equal to
the average energy value of the available ones. Not all stations had P and
S energy estimates because of the SNR threshold limitation (we chose
stations for which SNR > 2 over at least 50 samples, that is, a bandwidth
of $\geq 7$ Hz), resulting in 44 P and 17 S wave station estimates. The total
energy radiated by the earthquake was calculated by summing the con-
tributions of the P and S wave energies $E_P$ and $E_S$ after being logarithmic-
ally averaged over all the stations. We assumed $V_S = 4500$ m/s and obtained a
total radiated seismic energy $E_S = 3.9 \times 10^{12}$ J and an S-to-P energy
ratio of 22.

**Calculation of stress drop**

We used multiple approaches to estimate the rupture area, and hence
the stress drop, because they were model-dependent. If we assumed a
unilateral strike-slip rupture with a length of 610 m (calculated from the
directivity analysis) and that fault width ($W$) was equal to length ($L$), we
would obtain ($\mathcal{E}$, $\mathcal{S}$)

$$\Delta \sigma = \frac{2 M_0}{\pi L^3} \approx 60 \text{ MPa}$$

This should be considered a minimum estimate because $W$ could be
much smaller than $L$. If we assumed instead a circular fault with radius
$R = 610/2$ m, we would obtain ($\mathcal{E}$, $\mathcal{S}$)

$$\Delta \sigma = \frac{7 M_0}{16 R^3} \approx 330 \text{ MPa}$$

Alternatively, we could use the average S wave corner frequency of
2 Hz (fig. S2) to estimate a stress drop ($\mathcal{E}$), obtaining

$$\Delta \sigma = \frac{7 M_0}{16 \frac{f^3}{V_S^3 k^5}}$$

where $V_S = 4500$ m/s from the PREM (Preliminary Reference Earth
Model) model, and $k$ is determined by the source model. In the simple
circular model of Madariaga ($\mathcal{E}$) where $k = 0.21$, Kaneko and Shearer
($\mathcal{E}$) calculated $k = 0.26$ for a circular source and $k = 0.19$ for an
asymmetrical ellipse, both with a lower rupture velocity ($0.7 V_S$). These
values gave stress drop estimates of $\sim 90, 50, \text{ and } 120$ MPa, respectively.
However, note that our rupture velocity estimates were lower, and thus
the stress drop listed here should be considered as minimum.

Although it may be preferable to work with more directly observable
parameters, such as potency and strain drop ($\mathcal{S}$), we followed common
seismological practice and used seismic moment and stress drops. Stress
drop variations as a function of depth may be explained by the larger
shear modulus at greater depth, but the equivalent strain drop ($\sim 0.02$) of
the 2013 Wyoming earthquake was also larger than the strain drop of
shallow crustal earthquakes ($\mathcal{E}$). The stress drop difference reported
here cannot be explained solely by the rigidity differences at depth.

**Calculating geotherms**

We estimated the depth-dependent temperature profile using the grain-
size viscoelastic relaxation model of olivine in the mantle ($\mathcal{S}$) con-
strained by surface heat flow and crustal thickness. Our approach was
that of a forward calculation in which we generated a large number of
geotherms and then, using the viscoelastic relaxation model, calculated
the corresponding shear wave velocity as a function of depth and com-
pared it with published velocity models in the mantle ($\mathcal{S}$, $\mathcal{P}$). We
generated random geothermal models with physical constraints, includ-
ing a surface heat flow between 40 and 60 mW m$^{-2}$, a crustal thickness
of 50 km ($\mathcal{E}$), and thermal conductivity in the crust of 2.5 W m$^{-2}$, and a
temperature-dependent conductivity in the mantle ($\mathcal{S}$). For each geo-
therm, a velocity profile was predicted. The range of these predicted
shear wave models that best fit the data and their corresponding
geotherms are shown in Fig. 3 (and fig. S1) for both velocity models used
[blue ($\mathcal{S}$) and red (26) areas]. Our modeling was not capable of repro-
ducing the strong positive velocity gradient between depths of 50 and
80 km, suggesting that shear wave velocities are affected by factors
other than temperature ($\mathcal{S}$).

**SUPPLEMENTARY MATERIALS**

Supplementary material for this article is available at http://advances.sciencemag.org/cgi/
content/full/3/3/e1602642/DC1

fig. S1. Temperature modeling for the Randolph, UT earthquake.

fig. S2. Comparison of estimated S wave corner frequencies and the best-fitting source model
obtained from the STF stretching approach.

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**Competing interests:** The authors declare that they have no competing interests. Data and materials availability: All data needed to evaluate the conclusions in the paper are present in the paper and /or the Supplementary Materials. Additional data related to this paper may be requested from the authors. The IRIS Data Management Center was used to access waveforms and related metadata used here. The seismic data have been downloaded from the IRIS facility using the obspyDMT (57) software and processed using the multilayer algorithm (45).

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