Seismological constraints on the down-dip shape of normal faults

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SUMMARY

We present a seismological technique for determining the down-dip shape of seismogenic normal faults. Synthetic models of non-planar source geometries reveal the important signals in teleseismic $P$ and $SH$ waveforms that are diagnostic of down-dip curvature. In particular, along-strike $SH$ waveforms are the most sensitive to variations in source geometry, and have significantly more complex and larger-amplitude waveforms for curved source geometries than planar ones. We present the results of our forward-modelling technique for 13 earthquakes. Most continental normal-faulting earthquakes that rupture through the full seismogenic layer are planar and have dips of 30–60°. There is evidence for faults with a listric shape from some of the earthquakes occurring in two regions; Tibet and East Africa. These ruptures occurred on antithetic faults, or minor faults within the hanging walls of the rifts affected, which may suggest a reason for the down-dip curvature. For these earthquakes, the change in dip across the seismogenic part of the fault plane is ≤30°.

Key words: Body waves; Earthquake source observations; Dynamics and mechanics of faulting.

1 INTRODUCTION

It is not known whether major earthquake-generating normal faults are dominantly planar, or whether the dip of the fault changes with depth so that they have significant curvature. Seismic reflection profiles and geological exposures commonly reveal both listric and planar normal faults in sedimentary units. There is ongoing debate regarding whether faults with steep surface dips flatten into subhorizontal detachment horizons around the depth of the brittle-ductile transition (e.g. Velasco et al. 2010). High-resolution bathymetric studies have revealed in great detail the corrugated, domed footwalls exposed in oceanic core complexes, but the geometrical evolution of these structures, and whether or not they slip in earthquakes, remains enigmatic (Escartin et al. 2008; MacLeod et al. 2009; Smith et al. 2014). The loss in resolution of geodetic techniques (e.g. GPS and InSAR) at depth, together with the non-uniqueness involved in inverting surface measurements for fault geometry and slip, means that it is difficult, using geodetic techniques, to determine the down-dip shape of seismogenic faults, particularly at the depths at which earthquakes nucleate. In this paper, we present a seismological technique for constraining the down-dip curvature of active normal faults. The advantage of using seismology is that it sheds light on the shape of the seismogenic part of the fault at the time it was active, and the technique is only sensitive to the parts of the fault that are important in generating earthquakes.

The wider implications of this work lie in several areas. First, normal fault curvature controls the geometry of basins and affects mechanisms of crustal thinning. For example, curved faults bounding regions which deform internally can accommodate significantly more extension than a series of planar faults bounding rigid crustal blocks (Jackson & McKenzie 1983; Jackson & White 1989). Second, the processes that initiate, drive, impede and finally stop earthquake ruptures are intrinsically linked to the physical attributes of faults (e.g. Sibson 1985; Scholz 1998), therefore knowledge of the shape of faults is important in investigations of fault rheology and mechanics. Third, the down-dip shape of normal faults has implications for assessing earthquake hazard, from the perspective of the location of energy release and the down-dip width of fault available to rupture.

First, we discuss our approach and model set-up, and how our method differs from that of Braunmiller & Nábělek (1996), on whose work we build. We then present a synthetic example to highlight the features of teleseismic waveforms that are indicative of the presence or absence of down-dip curvature. Finally, we describe the application of our method to a catalogue of continental and oceanic normal-faulting events, and present the results of the modelling along with a discussion of the constraints on the down-dip shape of the fault for each earthquake. Though we focus on normal faults, our technique is equally applicable to thrust earthquakes.

2 DATA AND FAULT MODELS

If an earthquake occurs on a fault plane with down-dip curvature, as the rupture proceeds through time and advances across the fault surface, different portions of the fault, which have different dips, will contribute to the signal recorded at a seismometer. If the rupture is of sufficient duration then the moment contributions from different parts of the fault plane, which contain information about the source geometry, are separate enough in time to be detected in seismic
waves. We exploit this complexity in the waveforms in this study.

We define down-dip fault curvature as the difference in dip between the upper and lower portions of the fault. A listric geometry, that is, decreasing dip with increasing depth, is defined as having negative curvature; a convex-up surface has positive curvature; and a plane has zero curvature. A curved surface can be approximated by a series of connected planes with different dips. For simplicity, and building upon the work of Braunmiller & Nábělek (1996), we model earthquakes using three subsources (Fig. 1). In this model, the rupture surface is approximated by three planes (tri-planar), and the subsources are offset in time and space, to represent the progress of the rupture through different parts of the fault.

Our study is restricted to events with magnitudes $> M_w$ 5.5, below which we commonly do not see clear waveforms in broadband seismograms, and $< M_w$ 6.5, above which the waveform complexity resulting from the rupture of multiple fault segments results in trade-offs with the fault geometry. Unlike field-based, InSAR or seismic reflection studies, our resolution is not limited by the depth extent of the rupture, or by geographic considerations (such as site access, errors due to topographic or meteorological effects, or whether the epicentre is on land or under the oceans).

The seismological technique we present is a forward-modelling method, based on the comparison of recorded waveforms with synthetic waveforms for tri-planar faults with planar and curved geometries. To construct the geometry of the three subsources for the tri-planar models, we estimate the dimensions of the rupture plane using the seismic moment and a displacement-to-length ratio of $5 \times 10^{-5}$ (Scholz 1982). This rupture plane is broken into three rectangular patches of length, $L$, and down-dip width, $\frac{L}{3}$ (Fig. 1).

The centroids of the respective subsources are positioned at the centre of each subplane, and each subsource has a moment of $\frac{M_0}{3}$.

The relative positions and depths of the three subsources are recalculated for each degree of fault curvature that we model, while enforcing that the combined parameters of the three-source models average to the centroid solution, and that the structural contiguity of the fault model is maintained. As we prescribe each subsource with the same moment, the moment-averaged source parameters are also the same as the centroid solution. These steps ensure that our models are of physically plausible fault ruptures that are consistent with the overall centroid parameters of the events.

All subsources have strike and rake fixed to that of the centroid solution. The dip of the middle subsource is fixed to that of the centroid solution. To simulate down-dip curvature, the dips of the upper and lower subsources are varied in tandem, such that the average dip of the tri-planar model is the same as for the single-plane solution. The degree of curvature is the difference in dip between the upper and lower subsources ($-80^\circ$ is very listric, $+80^\circ$ is very convex-up). Dips are altered in $5^\circ$ increments, resulting in models at $10^\circ$ intervals of curvature, up to a maximum degree of curvature where the dip of the upper or lower subsource approaches vertical or horizontal.

Rupture of a curved surface is simulated by setting a time delay between subsources. We assume rupture nucleates at depth (initiating with the deepest subsource, $S_1$) and propagates upwards. We show in Appendix B in the Supporting Information the effects of rupture propagating from the surface downwards. The time delay between subsources is calculated using the model geometry and assuming a rupture velocity, $v_r$, of $2.5 \text{ km s}^{-1}$. Each subsource has the same source time function, defined by a single triangular element, the duration of which is set so that the overall duration of the tri-planar event matches that of the centroid solution. We carried out synthetic tests (Appendix B in the Supporting Information) that show that variations in the timing of moment-release (due to our choice of rupture velocity and source time function parameters) have only a minor effect on the waveforms compared to changes in the down-dip geometry, and that our assumptions therefore do not affect our ability to resolve down-dip curvature.

2.1 Previous work

Braunmiller & Nábělek (1996) also used teleseismic body waveform data to examine the fault geometries in 6 normal-faulting earthquakes in Greece and western Turkey. They used three-subsource models to produce synthetic waveforms for rupture on faults with down-dip curvature, and found that $SH$ wave shapes and polarities were strongly affected by down-dip curvature when observed at teleseismic stations lying approximately along-strike of the event. We build on their work to produce a new set of synthetic waveforms and present a technique for both identifying and constraining down-dip curvature from teleseismic data. We replicated the synthetic study of Braunmiller & Nábělek (1996) using the MT5 programme of Zwick et al. (1994) (which is detailed in Appendix A in the Supporting Information), and our method reproduces the differences in amplitude and polarity in the initial parts of $P$ and $SH$ waveforms between a planar rupture model and a curved fault plane (with $40^\circ$ listric curvature) that they observed. We improve upon the study of Braunmiller & Nábělek (1996) by investigating the effects of the choice of model rupture velocity, source duration
and propagation direction, which we describe in Appendix B in the Supporting Information. Different to Braunmiller and Nábělek, we also include the important and high-amplitude depth-phases (i.e. near-source surface reflections) in our synthetic waveforms, and we prescribe the subsource parameters such that our curved fault models are consistent with the centroid solution obtained through body waveform inversion.

3 SYNTHETICS

We examine the effect on $P$ and $SH$ waveforms of changes in the down-dip geometry of a synthetic $M_w 5.9$ normal-faulting event, by computing synthetic seismograms for a suite of tri-planar models covering a $-80^\circ$ to $+80^\circ$ range of curvatures (Fig. 2), using the MT5 algorithm of McCaffrey et al. (1991) and Zwick et al. (1994). The synthetic stations were chosen to represent the typical epicentral distance ranges of teleseismic data used in body waveform inversions, and the magnitude of the synthetic earthquake was chosen to be similar to the real events discussed later. Figs 3 and 4 compare the synthetic $SH$ and $P$ waveforms for a planar fault model, with those for curved fault models at $\pm 10^\circ$, $20^\circ$, $40^\circ$, $60^\circ$ and $80^\circ$ curvature. We investigate how waveform amplitude and complexity is affected by fault geometry and station location, and discuss the implications of our results for examining real data.

3.1 Synthetic waveforms

The synthetic seismograms at stations within $\sim30^\circ$ of fault strike have significantly more complex $SH$ waveforms for all curved fault models than for the planar case, even those with only $\pm 10^\circ$ curvature (Fig. 3). The $SH$ waveforms at these stations also have much higher amplitudes for curved models, and along-strike nodal stations (stations $S_1$ and $S_{13}$) show waveforms whereas none are present for the planar model. These effects are more pronounced for models with greater curvature. At stations with azimuths at high angle to strike, $SH$ waveforms for curved models are not as complex, and have similar amplitudes to (and are harder to distinguish by eye from) those for the planar fault model. However, for models with high curvature, the dip of the lowermost plane (subsource $S_1$) may be high or low enough that the nodal plane for this subsource lies on the opposite side of where these stations plot on the focal sphere, that is, the station projects to a different quadrant of the $SH$ focal sphere. In this case the $SH$ waveform shows a small amplitude peak with the opposite sign to that of the waveform for the planar model (Fig. 5). At nodal stations perpendicular to strike (stations $S_6$ and $S_{20}$), there is no $SH$ arrival in any model. These features arise because the $SH$ nodal planes cut through the centre of the focal sphere, passing through parts of the focal sphere that are sampled by seismometers at teleseismic distances. Changes in dip within the source rotate the $SH$ focal sphere such that the nodal planes sweep across the region in the lower hemisphere projection covered by the take-off angles of teleseismic waves. This effect makes $SH$ waveforms highly sensitive to changes in down-dip curvature. Small changes in curvature therefore have a large effect on the $SH$ waveforms at stations within $\sim30^\circ$ of fault strike. Therefore, to detect down-dip curvature in real earthquakes it is essential to include $SH$ data from stations in the along-strike directions. This finding is consistent with the observations made by Braunmiller & Nábělek (1996).

The shape of synthetic $P$ waveforms is much less sensitive to station azimuth and degree of curvature (Fig. 4), because the $P$-wave nodal planes of normal-faulting earthquakes cut through parts of the focal sphere that are distant from the take-off angles of seismic waves received at stations at teleseismic distances. There are small changes in the width and location of peaks in the waveform, however the main effect is on the amplitude of the waveform, particularly at stations that lie perpendicular to strike. Considering the true (E-dipping) nodal plane that represents our synthetic normal fault, for positive curvature models (convex), stations that lie in the down-dip (East) direction have smaller amplitude and more complex $P$ waveforms than stations in the up-dip (West) direction. The opposite is true for negative curvature (listric) models. This effect is stronger for models with greater curvature, as the dip of one subsource is steep and the projection of the relevant nodal plane passes close to stations in the focal sphere, resulting in less energy being received and smaller amplitudes being recorded at those stations (Fig. 5).
Figure 3. Synthetic SH waveforms generated for a $M_w$ 5.9 normal-faulting earthquake, with a strike, average dip and rake of 000°/45°/270°. SH waveforms were generated at the synthetic stations shown here, at an epicentral distance of 55°, which is typical of the epicentral distance ranges of teleseismic data used in body waveform inversions. The station position (intersection of the ray path with the lower hemisphere projection of the focal sphere) is shown with a solid black dot on the SH focal sphere. Synthetic waveforms are plotted around the SH focal sphere (which shows the nodal surfaces for an event with 0° down-dip curvature) at the approximate station azimuth, and labelled with the alphanumeric station name (bold) and azimuth. Synthetic SH waveforms, generated from tri-planar models, are shown for events with ±10°, 20°, 40°, 60° and 80° curvature in blue. Solid blue lines are for negative curvature (listric), dotted blue for positive curvature (convex up). Model curvature increases radially outwards, that is, the inner ring shows waveforms for models with 10° curvature, the outer 80°. Synthetics for curved models (blue) are plotted on top of a synthetic single-plane, 1-source centroid model (black) with 0° curvature.
Figure 4. Synthetic $P$ waveforms generated for a $M_w$ 5.9 normal-faulting earthquake, with a strike, average dip and rake of $000^\circ/45^\circ/270^\circ$. Key as for Fig. 3, but curved model synthetics are in red and the $P$ focal sphere is shown, with compressional quadrants shaded in grey. The $P$ stations are at an epicentral distance of $60^\circ$, and the station position (intersection of the ray path with the lower hemisphere projection of the $P$ focal sphere) is shown with a solid black dot.
The down-dip shape of normal faults

3.2 Implications for modelling real events

If seismogenic faults have only small degrees of down-dip curvature, the diagnostic signals (SH waveform complexity and/or initial low-amplitude peaks with the wrong polarity) will only be detectable at stations lying at azimuths at or very close to along-strike. Identifying the shapes of the waveforms correctly requires picking the precise onset times for P and SH arrivals. Stations that lie near an SH nodal surface receive a lower energy signal and show lower amplitude SH arrivals in seismograms than stations lying elsewhere in the SH focal sphere. Along-strike broadband data must be examined carefully for clear arrivals that can be distinguished from the background noise. Picking the wrong SH arrival time would skew any attempt to resolve down-dip curvature. For this reason we pick the waveform arrivals by hand, and when modelling real earthquakes we include data from as many along-strike stations as possible, so that one wrong or uncertain SH arrival pick will not adversely distort the results.

4 APPLICATION TO REAL EARTHQUAKES

We present a seismological, forward-modelling technique for determining the down-dip geometry of earthquake-generating dip-slip faults, based upon the synthetic models described above. We have applied this method to a catalogue of continental and oceanic events and present the results of the modelling in the next section.

4.1 Event selection

We performed body-waveform inversions for a catalogue of normal-faulting events (Table 1), from a wide range of extensional environments. We then selected suitable earthquakes on which to apply our seismological technique for determining down-dip curvature. Events were chosen based on (1) magnitude, (2) source time function, and (3) station signal-to-noise ratio and distribution. Events need to be large enough (\( \geq M_w 5.5 \)) that the seismic contributions from different parts of the rupture plane are distinct and separate in space and time; and small enough (\( \leq M_w 6.5 \)) that the approximation of the earthquakes as a point source (centroid), on which the initial waveform inversion is based, is still valid. Factors such as changes in the rupture propagation direction, uneven moment release, or earthquake doublets, have a large effect on the resultant waveforms recorded at teleseismic distances. This level of complexity leads to multiple trade-offs with down-dip fault geometry. Therefore we select events that are one-segment ruptures, with relatively simple source time functions. In order to perform the modelling effectively there must be a good azimuthal distribution of stations at teleseismic distances, a high density of stations at close to along-strike azimuths, and seismograms with clear arrivals and a high signal-to-noise ratio.

4.2 Procedure

The main stages of the method are as follows:

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**Figure 5.** The effect of a high degree of model curvature on synthetic waveforms. This figure illustrates that stations at azimuths >30° from strike may also be sensitive to down-dip curvature. Left: Model fault geometry and P and SH focal spheres for synthetic single-source planar \( M_w 5.9 \) event, as for Fig. 2. Example synthetic waveforms for stations P7 and S9 are shown to the right of the appropriate focal sphere. For both the P and SH focal spheres, quadrants with positive waveform arrivals are shaded grey, and the appropriate station ray path intersection is marked with a black dot. Station labels are as for Figs 3 and 4. Right: Tri-planar models with 60° down-dip curvature, as for Fig. 2. The synthetic P (red) and SH (blue) waveforms for stations P7 and S9 are shown to the right; solid line for negative curvature, dotted for positive. The synthetic waveforms are the signal resulting from the summed contribution of the radiated energy from all three subsources, however we show the P and SH focal spheres for only the first (deepest) subsource, \( S_1 \), which affects the initial part of the waveform. The quadrants with positive arrivals from \( S_1 \) are shaded and the intersection of the ray path for station P7 and S9 on the focal sphere is shown with a black dot. The dip of \( S_1 \) affects the position of the nodal surfaces relative to the station position, and the resultant polarity of the initial part of the waveform.
(i) Perform a body waveform inversion to obtain point-source centroid focal parameters, moment and source time function for the chosen event.

(ii) Use centroid solution to estimate source dimensions and set up geometry of tri-planar models.

(iii) Generate synthetic waveforms for tri-planar models with a range of down-dip curvatures.

(iv) Compare model synthetics with observed seismograms to constrain down-dip curvature.

4.2.1 Body waveform inversion procedure

Teleseismic data was downloaded using the IRIS (Incorporated Research Institutions for Seismology) Wilber III system and P and SH arrivals were manually picked on broadband seismograms. A full list of stations we used, along with their network codes, is found in Appendix C in the Supporting Information. We used data from all networks with suitable records, and restrict the station epicentral distance range to $30^\circ$–$90^\circ$ and $30^\circ$–$80^\circ$ for $P$ and $SH$ data respectively. This ensures that the $P$ and $S$ signals are not affected by arrivals of waves that have diffracted, reflected or travelled through the core, and that they have not been affected by shallow lithospheric phases (e.g. Moho head waves). The broadband seismograms are first deconvolved from their instrumental response, and then re-convolved with the response of a long-period (15–100 s) World-Wide Standardized Seismograph Network (WWSSN) instrument.

This procedure means that earthquakes of magnitudes $M_w$ 5.0–6.5 can be modelled as a finite-duration rupture at a point source (centroid). We then used the MT5 programme of Zwick et al. (1994), based on the algorithm of McCaffrey & Abers (1988) and McCaffrey et al. (1991), to jointly invert $P$ and $SH$ waveforms for earthquake source parameters (strike, dip, rake, centroid depth, moment and source time function). Given an initial model (taken from the GCMT catalogue) and our arrival time picks, we generate synthetic $P$ and $SH$ waveforms and solve iteratively for the source parameters that give a minimum misfit between the observed waveform and synthetics within the inversion window. The inversion window starts at the picked arrival time and the length is set for $P$ and $SH$ data so that it includes only the body waveforms and their depth phases, and no additional phases. $P$, $pP$ and $sP$ phases are modelled on vertical component seismograms and $S$ and $sS$ phases on transverse component seismograms. Synthetics are computed based on the source parameters and calculated take-off angles (with amplitudes corrected for geometrical spreading and attenuation). We solve for the strike, dip and rake of the source (which dominantly affect the shape of the waveforms), and the centroid depth and source time function (which chiefly control the length and amplitude of the waveforms). The source is constrained to be purely double-couple. A full description of this procedure can be found in Taymaz et al. (1991) and Nábělek (1984), and similar waveform inversion techniques are used extensively in the literature (Bergman & Solomon 1985; Braunmiller & Nábělek 1996; Craig et al. 2011; Sloan et al. 2011). This technique refines the accuracy of source parameter estimates made by routine catalogues such as the GCMT, particularly for the centroid depth. The degree to which each focal parameter can be resolved is determined by fixing a parameter, for example, depth, at a series of values away from the best fit, and re-inverting for all other parameters. The fit of the synthetic to the observed seismograms is then compared at successive values of the fixed parameter to determine the range over which a reasonable fit has been found. Typical errors are ±4 km for depth, ±10° for strike, ±5° for dip and ±10° for rake (Molnar & Lyon-Caen 1989; Taymaz et al. 1991; Craig et al. 2014).

We used a simple crustal velocity model consisting of a half-space of $V_p = 6.5 \text{ km s}^{-1}$, $V_s = 3.7 \text{ km s}^{-1}$, $\rho = 2800 \text{ kg m}^{-3}$. Discrepancies between the model velocity structure and the actual (unknown) velocity profile at the location of each earthquake (e.g. due to shallow, low-velocity sedimentary layers above the source) affect the estimated centroid depth, but not the geometry of the nodal planes (e.g. Bergman & Solomon 1985; Nelson et al. 1987; Taymaz et al. 1990; Middleton & Copley 2013). The lack of sensitivity of the orientation of the focal planes to the velocity model results from the constraint that the source be comprised of double-couple subsources, and the simultaneous inversion of $P$ and $SH$ waves at all stations (Fredrich et al. 1988).

### Table 1. Events used in this study.

<table>
<thead>
<tr>
<th>Name</th>
<th>Region</th>
<th>Lat (°)</th>
<th>Long (°)</th>
<th>yymmdd</th>
<th>hh:mm:ss</th>
<th>$M_w$</th>
<th>$M_o$ (N m)</th>
<th>Strike (°)</th>
<th>Dip (°)</th>
<th>Rake (°)</th>
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<td>134</td>
<td>46</td>
<td>261</td>
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<td>12.781</td>
<td>970926</td>
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<td>5.8</td>
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<td>40</td>
<td>270</td>
<td>5.4</td>
</tr>
<tr>
<td>Aegion</td>
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<td>22.307</td>
<td>950615</td>
<td>00:15:50</td>
<td>6.2</td>
<td>2.34 × 10^{18}</td>
<td>290</td>
<td>27</td>
<td>305</td>
<td>11.6</td>
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<tr>
<td>Simav</td>
<td>Turkey</td>
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<td>29.070</td>
<td>110519</td>
<td>20:15:25</td>
<td>5.8</td>
<td>7.02 × 10^{17}</td>
<td>116</td>
<td>37</td>
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<td>Palu</td>
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<td>119.878</td>
<td>090302</td>
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<td>5.6</td>
<td>7.02 × 10^{17}</td>
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<td>2.95 × 10^{18}</td>
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<td>43</td>
<td>262</td>
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Hypocentre locations and times were retrieved from the ISC; location data for events occurring prior to 2009 is from the EHB bulletin. The centroid parameters were obtained through body waveform inversion in this study. If the true fault plane is not known, parameters for both nodal planes are shown.
Table 2. Results from down-dip curvature modelling. Plain grey bars delimit range of down-dip curvature investigated for each event. Grey boxes indicate the extent to which down-dip geometry can be constrained—they show cross-sections of the model geometries that produce synthetics with an acceptable fit to recorded broadband seismograms. The red line marks the modelled centroid depth with respect to the model fault plane. Red dots mark the centroids for the events that show no resolvable evidence for down-dip curvature; orange dots mark the subsource centroids for our preferred tri-planar fault model for events where there is evidence for down-dip curvature. If the true fault plane (FP) is not known, modelling results for both nodal planes are shown (NP1 and NP2).

<table>
<thead>
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<th>Event name</th>
<th>Down-dip curvature modelling results</th>
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4.2.2 Constraining down-dip geometry

Having obtained a centroid solution for the earthquakes we study, we use the same procedure as that for the synthetic event (described above) to set up tri-planar models with down-dip curvature (Fig. 1). The MT5 programme (Zwick et al. 1994) is used to generate synthetic P and SH waveforms for the curved tri-planar models at all stations used in the initial inversion. As a first indicator of down-dip curvature in the source, we look for complex SH waveform shapes with higher than expected amplitudes at stations within 30° of the along-strike azimuth. We then attempt to put constraints on the down-dip geometry by determining the range of down-dip curvature models that produce an acceptable fit to all the waveform data. For each SH station (including those >30° from strike) and for every tri-planar curvature model, synthetic waveforms are compared to the observed seismogram, and the synthetic from the planar solution, in order to determine whether the synthetics produced with a curved fault plane yield a better fit to the observed data, and over what range of curvature there is an acceptable fit to the waveforms.

We calculate the variance of the residuals between the recorded signals and the synthetic waveforms within the inversion window, and plot graphs of variance against curvature for SH-waveforms recorded at stations within 30° of strike. The important observations are how the misfit varies as a function of curvature for each station. The waveforms are visually inspected to check for features that may skew the plots or produce a misleading variance, such as synthetics that fit most of the waveform but have the wrong first motion (which would give a low variance of the residuals, despite having the wrong first motion), or the presence of noise in some seismograms (which may result in a high variance, even if the synthetic broadly fits the observed waveform). In order to constrain the down-dip curvature of an event, we examine the visual fit of synthetics together with the graphs of variance against curvature. If down-dip curvature is present, we expect to see a consistent agreement across the SH stations, with improved fits at curved models; though we expect the stations >30° from strike to be much less sensitive. Finally, to determine a preferred model, and to avoid biases due to uneven azimuthal distribution of stations or biases in favour of fitting SH-waveforms at the expense of P-waves, we examine the fit of the curved models as a whole, to all P- and SH-waveforms. Looking for common features among multiple stations is diagnostic of source, rather than receiver, effects.

5 RESULTS

We present the results of body-waveform inversions (Table 1) and the constraints on the down-dip geometry (Table 2) of 13 events. We begin by presenting the source parameters and modelling results for a well-studied event, the 2009 L'Aquila earthquake. There are multiple independent geophysical inversion and aftershock studies for this event, against which we compare the results of our
Geologic mapping and analysis of coseismic surface ruptures (EMERGEO Working Group 2009; Falconu et al. 2009; Boncio et al. 2010; Bonini et al. 2014), up-dip surface projections of fault plane solutions and analysis of the ground displacements measured by InSAR and GPS (Atzori et al. 2009; Walters et al. 2009) demonstrate that the main shock ruptured the Paganica Fault. There is some disagreement on whether coseismic slip only occurred up-dip of the 8–10 km hypocentre (Herrmann et al. 2011; D’Agostino et al. 2012) or whether there was also a deeper slip patch between 9–14 km (Cirella et al. 2009; Cheloni et al. 2010), however most of the slip was shallower than the hypocentre. Our modelled centroid depth of 5 km, and the depth extent of model fault geometries (1–10 km) are consistent with the estimates of the extent of the fault plane in the source parameter studies mentioned above (1–15 km depth).

Of eight stations within 30° of strike, six have SH-synthetics with a good fit to the observed data for the planar solution (e.g. RAYN, WCI; Fig. 7c). Models with >10° of curvature produce synthetic SH-waveforms for which the overall waveform shape and/or peak amplitudes and polarities are significantly different to the observed waveforms at these stations, that is, the recorded data precludes fault geometries of ≥20° of curvature for this event. Recorded waveforms at the other two stations in this azimuth range (MSEY and DWF) are not fit well by the planar solution synthetics (most likely due to station noise of unknown origin); varying the down-dip geometry does not yield a better fit. Synthetics generated from some curved models yield fits that are equally good or better than the planar case for individual stations (e.g. WCI +30°; Fig. 7c). However, as all the stations record the same event, it is important to find a model that is consistent with the recorded P and SH waveforms at the majority of stations, rather than a model that yields excellent fits at only some of the along-strike SH stations. There is no consistent trend among the stations as a whole favouring either positive or negative curvature, and minor improvements in fit at individual stations are likely due to fitting noise. This feature is demonstrated by the differing minimum-variance model for each of the SH-stations within 30° of strike (Fig. 7b). The average variance of the residuals for these stations is lowest for a model with 0° curvature (i.e. planar). SH-stations at azimuths 30°–30° from strike have wider ranges of down-dip curvature over which the model produces synthetics with an acceptable fit to the recorded data (e.g. TIXI; Fig. 7c); those that lie roughly perpendicular to strike are insensitive to changes in model geometry (e.g. HIA). Increasing the down-dip curvature produces only small amplitude changes in synthetic P-waveforms at most stations (e.g. TSUM), but large changes in stations with high take-off angles and that lie near a P nodal plane (e.g. KBS, ARU). For these stations, models with >20° curvature produce synthetic P-waveforms with larger amplitudes compared to the other stations (and compared to the planar model). We do not observe this azimuthal variation in the recorded P-waveform data. Taken as a whole, all these observations combined mean that we infer that the L’Aquila earthquake ruptured a planar fault. If there is a change in dip in the down-dip direction, it is not more than ±10°.

Chiaraluce et al. (2011) and Valoroso et al. (2013) relocated large catalogues of foreshocks and aftershocks associated with the L’Aquila main shock. All three studies showed tight clustering of the relocated events along a plane dipping ~50° to the southwest. Their close spatial alignment argues for the Paganica Fault having a planar down-dip geometry, consistent with the results of our analysis of teleseismic body waveforms and down-dip curvature modelling. The corroboration of our result by other geophysical methods demonstrates the robustness of the technique.
Figure 7. Down-dip curvature modelling results for the 2009 April 6 L’Aquila earthquake. Figure overleaf. (a) Event location map, focal mechanism and distribution of stations. Centroid parameters obtained through body waveform inversion (planar solution) are shown above the map; the focal mechanism is plotted at the location of the epicentre. Stations used in the inversion are named. (b) Plots of variance against curvature for SH-stations within 30° of strike. Variance was calculated for the residuals between synthetic and recorded SH-waveforms, within the inversion window used to obtain the initial centroid solution, for each station and tri-planar model. (c) Down-dip curvature modelling results and synthetics for select stations. Cross-sections through model fault geometry are shown above. The range of curvature constrained by the data is highlighted in black, the best-fit model geometry in blue. Synthetic waveforms calculated for three-subsource models for select stations are plotted below (red for P-, blue for SH-waves), on top of the recorded waveform (black). For the 0°-curvature model, the synthetic from the point-source planar solution (grey) is plotted on top of the recorded waveform. The station positions are shown on the relevant focal sphere.
5.2 Planar ruptures

Our analysis shows that the majority of normal-faulting earthquakes that are large enough to rupture through the seismogenic layer occur on planar faults. These events showed no strong evidence for down-dip curvature, had dips between 30°–60°, and could be constrained to be planar to within ±20°. We describe in detail below the 1997 September 26 Colfiorito earthquake, Italy, and the 1995 May 13 Grevena earthquake, northern Greece. A full description of our results for the additional planar ruptures (1995 June 15 Aegion earthquake, Greece; 2005 March 6 earthquake, Gakkel Ridge, Arctic Sea) is presented in Appendix D in the Supporting Information.

5.2.1 The 1997 September 26 Colfiorito earthquake, Italy

The 1997 September 26 Colfiorito earthquake was largest of six $5 < M_w < 6$ events that struck the Abruzzo region (Fig. 8) of central Italy during September–October 1997, known as the Umbria–Marche sequence (Chiaraluce 2003). Our modelled centroid dip (40°) agrees well with that obtained by other authors through waveform inversion (Ekström et al. 1998; Morelli et al. 2000), the alignment of relocated aftershocks (Amato et al. 1998; Deschamps et al. 2000; Chiaraluce 2003) and various geodetic analyses (Hunstad et al. 1999; Stramondo et al. 1999; Salvi et al. 2000; De Martini et al. 2003). We obtain a centroid depth of 5 km, consistent with the ~4.5 km depth of maximum slip obtained by Lundgren & Stramondo (2002) through joint inversion of GPS and InSAR data, and infer the depth extent of faulting to extend between >1 km to <8 km. This depth distribution is supported by observations that very little slip reached the surface (Cinti et al. 1999; Cello et al. 2000; Basili & Meghraoui 2001), aftershock hypocentres delineate a cut-off in seismicity beneath ~8–9 km depth (Deschamps et al. 2000; Chiaraluce 2003), and forward-modelling of coseismic displacements on a fault plane with a similar depth extent adequately reproduces measured GPS and InSAR surface displacements (Hunstad et al. 1999; Salvi et al. 2000).

Our planar solution yields synthetics with moderately good fits to the shape and amplitude of $SH$-waveforms recorded at all stations within 30° of strike (Fig. 9c). Models with $\geq 20°$ of curvature produce complex synthetic $SH$-waveforms that do not fit the observed data at these 6 stations, particularly for stations lying SE of the event (e.g. MSEY; Figs 9a and c). The average variance of $SH$-waves at stations within 30° of strike is lowest for models with 0° to $+10°$ curvature (Fig. 9b). For stations $>30°$ from strike, the $SH$-waveforms are less sensitive to changes in the down-dip geometry (e.g. ALE; Fig. 9c), and individual stations have a wider range of curvatures over which the model generates a synthetic that fits the recorded waveform equally well. In general, increasing down-dip curvature has only a small effect on the amplitude of $P$-waveforms (e.g. FURI, FFC; Fig. 9c). Stations with high take-off angles and that lie near a $P$-nodal plane (e.g. KBS, ARU; Fig. 9c) exclude geometries with $\geq 30°$ of curvature; for these geometries the initial dip is high or low enough that the $P$-synthetic produced at these stations may have the wrong initial polarity and/or a substantially different amplitude compared to $P$-synthetics for other stations and the observed data. There is no evidence to suspect down-dip curvature for this event. The data allow for a maximum of $\pm 10°$ curvature; outside of this range the fit of synthetics at key stations deteriorates. We therefore infer that the Colfiorito event ruptured a planar fault, to within $-10°/+10°$ degrees.

Some workers have described the faults in the Colfiorito area as having listric geometries (Boncio & Lavecchia 2000; Barchi & Mirabella 2009), because the coseismic faults do not project to the surface while maintaining their 35–45° dips, and exposed fault planes have measured dips of ~60–70° (Cello et al. 2000; Chiaraluce et al. 2005). Additionally, it has been suggested that the normal faults bottom-out below ~8 km, either onto pre-existing, low-angle structures (e.g. the Alto-Tiberino Fault; Boncio & Lavecchia 2000) or into horizons of Triassic evaporites (Mirabella et al. 2008; Barchi & Mirabella 2009), though there is a dearth of good quality seismic reflection data to test these suggestions (Chiaraluce et al. 2005; Mirabella et al. 2008). Although co-seismic displacements do seem to be taken up on high-angle fault strands at the surface—evidenced by diffuse aftershock seismicity in the upper 1 km (Chiaraluce et al. 2004), complex patterns of surface deformation (e.g. Stramondo et al. 1999; Cello et al. 2000; Salvi et al. 2000; Vittori et al. 2000) and reconstruction of levelling-line displacements (Basili & Meghraoui 2001)—modelling of surface displacements indicates that the high-angle structures accommodating this slip only extend to ~500 m beneath the surface (Basili & Meghraoui 2001). Therefore the faults activated in the...
Figure 9. Down-dip curvature modelling results for 1997 September 26 Colfiorito earthquake. Caption as for Fig. 7.
Umbria-Marche sequence are not listric in the sense of a smooth decrease in dip with depth. Instead, the analysis presented here, and the close alignment of aftershock hypocentres, indicates that the faults are planar at the depths where they rupture in earthquakes (Amato et al. 1998; Chiaraluce 2003), that is, there is less than ±10° change in dip with depth between ~2–8 km. The hypocentres of the larger events in the sequence, including the 1997 September 26 event modelled here, are located at depths close to the base of these aftershock clusters (Stramondo et al. 1999; Barba & Basili 2000), at the same depth as foreshock activity (Cattaneo et al. 2000), and close to the inferred intersection with pre-existing low-angle faults (Chiaraluce et al. 2005; Mirabella et al. 2008). A lack of seismicity at depths >8 km suggest this decollement, if it exists, is not seismically active. Both the modelling in this study and detailed studies by other workers establish that extension in the seismogenic layer is accommodated by rupture on planar faults in the Colfiorito region.

5.3 The 1995 May 13 Grevena earthquake, Greece

On the 1995 May 13, a destructive $M_{w}$ 6.4 normal-faulting earthquake hit northern Greece, approximately equidistant from the towns of Grevena and Kozani (Fig. 10). We obtained centroid source parameters consistent with other estimates from body waveform seismology (Hatzfeld et al. 1998), inversion of horizontal GPS displacements (Clarke et al. 1997) and InSAR studies (Resor et al. 2005) to within 10°, and the strike is similar to that measured for the only mapped fault with conspicuous surface rupture (the 250°-striking Paleochori Fault; Meyer et al. 1996). We obtained the same centroid depth (11 km) as Hatzfeld et al. (1998), and our estimated depth extent of the rupture (6–15 km) is consistent with their 14.2 km main-shock hypocentre and the distribution of aftershocks between 5–15 km.

The planar solution from the body-waveform inversion generates synthetics that fit the shape and amplitude of SH-waveforms very well at 11 of 12 SH-stations within 30° of strike. There is therefore no indication that there may be down-dip curvature. Models with positive curvature produce complex SH-synthetics at these stations which have very different waveform shapes and/or the wrong initial polarity when compared with the observed waveforms (e.g. AAK; Fig. 11c)—this precludes any positive down-dip curvature. Models with negative curvature produce synthetic SH-waveforms with similar shapes to the recorded waveforms. Increasing the degree of negative curvature mainly affects the amplitudes, not the shape, of the synthetic waveforms, and yields a better fit of the resultant synthetic to the data for some stations (e.g. KURK –30°). However, for SH stations at azimuths 30°–65° from strike, models with more than −20° curvature generate synthetics with amplitudes much larger than the recorded data (e.g. SJG –40°). Synthetics for SH stations at azimuths >65° from strike have poor fits irrespectively of model fault geometry, as they lie close to an SH nodal surface and so have a poor signal-to-noise ratio. Stations with high take-off angles and lying roughly perpendicular to strike plot near one of the $P$ nodal planes in a lower hemisphere projection. Small increases in negative curvature produce synthetic $P$-waveforms with the wrong initial polarity at these stations (e.g. NAI, ATD), ruling out more than −20° down-dip curvature. Plots of variance against curvature are skewed towards negative-curvature models (Fig. 11b), reflecting the extremely high misfits for positive-curvature models. Overall, both $P$ and $SH$ waveform data support a planar geometry and the down-dip curvature can be constrained to between −20° and 0°.
Figure 11. Down-dip curvature modelling results for 1995 May 31 Grevena earthquake. Caption as for Fig. 7.
several studies (e.g. Meyer et al. 1996; Resor et al. 2005) to build up a detailed picture of the fault network in the Kozani–Grevena region, and given the large moment-release it is not surprising that multiple fault strands, particularly those up-dip and in the hanging wall of the main shock, were activated by the Grevena earthquake. Despite the fine-scale complexity of the fault system, the results of the down-dip curvature modelling in this study suggest that the major coseismic displacements are consistent with rupture on a planar structure.

5.4 Ruptures with down-dip curvature

Based on the analysis of the $SH$ waveforms at stations within 30° of strike, there is a strong indication of down-dip curvature in 4 events. Two events ($M_c$ 5.7 and a $M_c$ 5.8) were from an earthquake sequence that occurred on the Karonga Fault system, Lake Malawi, and which culminated in a $M_c$ 5.9 (included here for completeness, but which does not show any evidence of down-dip curvature). The other two were 2004 and 2005 $M_c$ 6.2 events that occurred in the Lunggar Rift region, Tibet.

5.4.1 The 2004 and 2005 Zhongba events, Lunggar Rift, Tibet

We studied two $M_c$ 6.2 normal-faulting events that occurred in Zhongba county in the Xizang region of southwestern Tibet, China (Fig. 12). Just north of the Indus suture the tectonics of southern Tibet are characterized by extension across N–S-trending grabens (Armijo et al. 1986). The 2004 July 11 and 2005 April 7 earthquakes described here were located in one of these active grabens, the South Lunggar Rift.

The South Lunggar Rift is comprised of a ~50 km wide central horst, the Surla Range, bounded on either side by the gentle west-dipping South Lunggar Detachment and steeper east-dipping Palung Co Fault (Styron et al. 2013). Small fault scarps are observed to cut glacial moraines along the northern portion of the Palung Co Fault (Styron et al. 2013), and the up-dip projections of InSAR-modelled fault planes for a 2008 normal-faulting doublet (Elliott et al. 2010) also coincide with the range front and mapped fault trace in the north, indicating that the fault is currently active. Elliott et al. (2010) performed a study of a series of large normal-faulting events in Tibet, including the two we model here, using InSAR and body waveform seismology. For both events, the surface deformation in InSAR interferograms was best fit by west-dipping faults located east of the Palung Co Fault (Elliott et al. 2010, and Fig. 12). The seismogenic thickness is ~15 km in Tibet (Craig et al. 2012), and the magnitude and depth of the earthquakes suggest that they ruptured through most of the seismogenic crust. These two events seem to represent break-up of the hanging wall of the major basin-bounding Palung Co Fault on antithetic faults.
2004 July 11 event. Synthetics from the planar solution obtained through body-waveform inversion fit SH-waveform data quite well at the nine stations within 30° of strike. However for the majority of these stations, the synthetics are consistently of a lower amplitude than the recorded signal, and do not match the fine detail of the waveform shape (the polarity of the small peaks in the initial part of the waveform) (e.g. OBN, ARU, SFJ; Fig. 13c). These small initial waveform peaks are a strong indication of down-dip curvature. Model geometries with some negative curvature (between −10° to −30°) consistently yield synthetics with better fits to the recorded waveform than for a planar model, particularly with respect to these details of SH-waveform shape and amplitude (e.g. ARU −20°, SFJ −30°). This improvement in fit is also reflected in the variance-curvature graphs for stations within 30° of strike (Fig. 13b); eight have minimum variances for models with negative curvature, and the average model variance is lowest for a model with −20° down-dip curvature.

The shape and amplitude of SH-waveforms at all stations at azimuths >30° and <55° from strike are also better fit by synthetics from models with −10° to −30° curvature (e.g. MAJ −30°, GNI −20°; Fig. 13c). Stations at azimuths >55° from strike have complex, low-amplitude signals that are not well matched by any synthetics, due to the poor signal-to-noise ratio resulting from their being close to an SH nodal plane. From the synthetic modelling presented in Section 3, stations in this azimuth range are not expected to be sensitive to down-dip curvature. Models with any positive down-dip curvature produce synthetic SH-waveforms with complex wave shapes and initial peaks with the opposite polarity to the recorded signal at all stations within 47° of strike, ruling out any convex down-dip geometry for this event. For stations that plot near the E-dipping nodal plane in a lower hemisphere projection, models with −40° down-dip curvature or more produce synthetic P-waveforms with the wrong initial polarity (e.g. SSE, DAV; Fig. 13c). There is a good overall fit of both P and SH waveforms at models with −30° to 0° curvature. The results suggest that this event involves down-dip curvature, and the best-fit model is slightly listric, with −20° curvature.

2005 April 7 event. Synthetic SH-waveforms generated at five of six stations within 30° of strike for a model with planar down-dip geometry have much smaller amplitudes and different shapes compared to the recorded waveforms (e.g. KONO, KEV; Fig. 14c). Similar to the 2004 event, this is a strong indication of down-dip curvature. Synthetics generated from models with −30° to −10° negative down-dip curvature have very good fits to the data at these stations, in terms of the amplitude, overall shape and polarity of small initial peaks in the waveform (e.g. MBWA −30°, ARU −20°). The fit of the synthetics is significantly better for this range of negative-curvature models than for a planar model geometry. This feature is reflected in the shape of the graphs of variance against curvature for these stations, which have residual variances that tend to a minimum in this model range, and are lower than the variance for the planar model for 4 stations (Fig. 14b). In almost all cases the fit of synthetic SH-waveforms to the seismograms is very poor for all models with positive curvature. For strongly listric models (−40° down-dip curvature or more) synthetic SH-waveforms have very high amplitudes compared to the recorded waveforms, and the fit of the synthetic deteriorates with increasing negative curvature. The fit of synthetic P-waveforms at stations with high take-off angles and at azimuths that are roughly perpendicular to strike (e.g. MAJO, TATO) also becomes significantly worse in this range, as the synthetic waveform has the wrong initial polarity. For almost all stations at azimuths >30° and <50° from strike, models with −30° to −10° curvature also yield synthetics with a better fit to the recorded SH-waveform than synthetics from the planar model, in terms of both the detail of waveform shape and the amplitude of the peak. Based on the analysis of the SH waveforms, there is strong evidence for down-dip curvature in the range −10° to −30°. The best-fit model has −20° curvature.

5.4.2 The 2009 Karonga sequence, Lake Malawi

We studied the three largest events (6th December Mw 5.7, 8th December Mw 5.8, and 19th December Mw 5.9) in an earthquake sequence that struck the Karonga region of northern Malawi, on the western shore of Lake Malawi, between November and December 2009 (Fig. 15). The rift basin is strongly asymmetric; the footwall of a steep west-dipping fault (the Livingston Fault) forms a 2 km high escarpment on the eastern bank of the lake, and its hanging wall is broken up by minor west-dipping faults, mapped by offshore seismic reflection profiles (Flannery & Rosendahl 2009; Ebinger et al. 1999). The east-dipping Karonga Fault forms the western-most boundary of the basin and divides the hilly topography and exposures of Precambrian to lower Palaeozoic basement rocks in the west from the flat alluvial plain in the east (Hamiel et al. 2012). Following the earthquake sequence, surface breaks were observed in an area north of Karonga city with no previous evidence of surface-rupturing faults, offsetting Quaternary sediments in the flat plain (Hamiel et al. 2012; Machevsky et al. 2015). Biggs et al. (2010) used seismology and InSAR to demonstrate that the sequence occurred on shallow, west-dipping structures in the hanging wall of the Karonga Fault. It has been suggested that failure initiated on a favourably orientated pre-existing structure, and then migrated to adjacent planes of weakness as a result of static stress transfer (Fagereng 2013).

2009 December 19 event. Unlike the other earthquakes we modelled from this sequence (6th December Mw 5.7 and 8th December Mw 5.8 events), there was no evidence of down-dip curvature for this event. Of nine stations within 30° of strike, the planar solution from the body-waveform inversion generates synthetics with good fits to the shape and amplitude of SH-waveforms at five stations (e.g. MAW, TUE, KIEV; Fig. 16c). At the other four, the synthetic waveform shape has a poorer fit and the amplitude is lower than the recorded data (e.g. TAM, BFO). Down-dip curvature modelling does yield synthetics with better fits than the planar model, for these stations. However, it also results in SH-synthetics with increasingly worse fits to the waveform data for the five stations that are well-matched by the initial planar solution. Positive-curvature models generate very different synthetic SH wave shapes, which in some cases have the wrong initial polarity compared to the recorded data (e.g. BFO +20°). Models with negative curvature produce synthetic SH-waveforms with high amplitudes, which yield individual best-fit models for some stations (e.g. BFO −30°), but very poor fits for others (e.g. MAW −30°, DPC −30°). The minimum-variance model for most of the along-strike SH-stations is between −10° to +10° (Fig. 16b). The average is −10°, however this is most likely skewed by the result for EIL, for which the observed waveform data is not fit well by any model, irrespective of geometry, presumably due to noise in the data.

SH-waveforms at stations at azimuths >30° from strike are less sensitive to down-dip curvature and are overall well-matched by the initial planar solution. Increasing degrees of down-dip curvature generates synthetics with poorer fits (e.g. XMIS, UOSS; Fig. 16c). The complexity, polarities and reduced amplitudes of synthetic
Figure 13. Down-dip curvature modelling results for the 2004 July 11 Zhongba earthquake. Caption as for Fig. 7.
Figure 14. Down-dip curvature modelling results for the 2005 April 07 Zhongba earthquake. Caption as for Fig. 7.
Figure 15. Map of northern Lake Malawi. (a) Regional map of the southern part of the Western Branch of the East African Rift. Country boundaries are in grey, major normal faults mentioned in the text are in red. The traces of the Kanda (KF), Livingstone (LF) and Bilila-Mtakataka (BMF) Faults are after Versfelt & Rosendahl (1989), Jackson & Blenkinsop (1997) and Vittori et al. (1997), respectively. The perimeter of box B is shown in grey and the inset globe shows the location of box A in red. (b) SRTM30 topography of northern Lake Malawi (scale to top left). Focal mechanisms obtained in this study are offset from their EHB locations (red dot) and labelled with a date (ymmd) identifier. The location of Karonga is marked with a black square; KFS is the Karonga Fault system. Faults, delineated in black, are retraced from Biggs et al. (2010); the surface ruptures of Macheyeki et al. (2015) are in white with a black outline (just N of Karonga). The trend of basement foliations are marked in purple, after Versfelt & Rosendahl (1989).

P-waves at stations with high take-off angles and that lie at high angles from strike precludes geometries with >30° down-dip curvature (e.g. KAPI, RCBR). The overall fit of synthetic SH-waveforms is significantly worse outside the range −30° to +10°. We therefore suggest that this event in the Karonga sequence ruptured a planar fault.

2009 December 8 event. The planar solution from the body-waveform inversion generates synthetics with similar waveform shapes, but much smaller amplitudes than the recorded SH-waveform at all five stations within 30° of strike (e.g. TUE, ID1; Fig. 17c). This is a strong indication of down-dip curvature. Synthetics generated from models with negative down-dip curvature have excellent fits to the data at these stations, both for the waveform amplitude and shape, particularly in the −30° to −60° range. The fit of the synthetics is significantly better for this range of negative-curvature models than for a planar model geometry. This is reflected in the shape of the graphs of variance against curvature for these stations (Fig. 17b). For all five stations, models with negative down-dip curvature yield lower variances than that for the single-plane solution and their respective minimum variances lie in the range −50° to −20°.

For stations at azimuths >30° from strike, there is a wide range of model geometries over which the fit of the synthetic (in terms of waveform shape and amplitude) to the observed SH-waveform is adequate. In general the fit is equally good or slightly better for negative-curvature models (e.g. SYO −30°; Fig. 17c), and equally good or slightly worse for positive-curvature models (e.g. SYO +10°). Although the overall fit of SH-synthetics is best for models with strong negative curvature (−50°), the fit of synthetic P-waveforms at stations with high take-off angles and at azimuths that are roughly perpendicular to strike (e.g. CHTO, DGAR and XMIS) becomes significantly worse as the synthetic waveforms have the wrong initial polarity. The overall fit of all P and SH synthetics is good for models with geometries in the −40° to −10° curvature range. Due to the location of this event, much of the region lying to the south and within the teleseismic distance range is covered by ocean (Fig. 17a), therefore there are large (−90°) azimuth gaps in the S-wave data-coverage. It would be preferable to have a more even distribution of station data, particularly for SH-waveform data, however there is strong evidence from the SH-waveform data that does exist (particularly from the along-strike stations to the north, and the one SH-station to the south, SYO).
Figure 16. Down-dip curvature modelling results for the 2009 December 19 Karonga earthquake. Caption as for Fig. 7.
Figure 17. Down-dip curvature modelling results for the 2009 December 8 Karonga earthquake. Caption as for Fig. 7.
that this event ruptured a fault with negative down-dip curvature. From this evidence, and considering the fit to all available $P$ and $SH$ waveforms, our preferred model is $-30^\circ$ curvature.

2009 December 6 event. The planar solution from the body-waveform inversion generates $SH$-synthetics with similar waveform shapes and amplitudes to the observed models for the 9 stations at azimuths within $30^\circ$ of strike (e.g. GNI, KBA; Fig. 18c). There is therefore no strong indication of down-dip curvature. However, the minimum-variance model for all but one station has a geometry with negative curvature (Fig. 18b), and models with some negative curvature ($-30^\circ$ to $-10^\circ$) produce synthetics with similarly good wave shapes and amplitudes as the synthetics for the planar model.

For all stations at azimuths $>30^\circ$ from strike, models with strong negative curvature yield synthetics with a better fit to the recorded $SH$-waveform than synthetics from the planar model, in terms of both the detail of waveform shape and the amplitude of the peak (e.g. UOSS $-50^\circ$, DBIC $-50^\circ$). At stations with high take-off angles and at azimuths that are perpendicular to strike (e.g. UOSS, PALK and SHEL), synthetic $P$-waveforms have small amplitudes and initial peaks with the wrong polarity compared to the recorded waveforms for models with $-40^\circ$ or more down-dip curvature, ruling out an extremely listric fault geometry. The fit synthetic $SH$-waveforms deteriorates for all stations for models with increasing positive curvature.

Due to the small size ($M_w 5.7$) and the location of this event, there is not a good azimuthal coverage of $SH$-stations at teleseismic distances (Fig. 18a), especially south of the event. There is a possibility that this event may have down-dip curvature, especially given its proximity to the 8th December event, however we are unable to constrain the geometry further than to say models with down-dip curvature in the range of $-30^\circ$ to $+10^\circ$ (including a planar geometry), give equally good overall fits to $P$ and $SH$ waveforms.

6 DISCUSSION

As described in the sections above, the down-dip curvature modelling results suggest that continental normal-faulting earthquakes that are large enough to penetrate to the bottom of the seismogenic layer rupture planar faults; most events showed no strong evidence for down-dip curvature, had dips between $30^\circ$ and $60^\circ$, and could be constrained to be planar to within $\pm20^\circ$.

In both the Apennines and the Aegean, recent large normal-faulting earthquakes have generated very little surface rupture in regions with complex mapped faults. The rise of geodetic observational techniques (InSAR and GPS) has resulted in numerous studies that produced competing models of rift-parallel ridged relief forms as a result of simple domino-style extension and rotation of successive planar normal faults, or whether it is formed from the steep surface breakaways of listric detachment faults (Smith et al. 2008). Resolving the down-dip geometry of oceanic normal-faulting earthquakes offers a way to discriminate between models. Based on the observed waveforms, we inferred that both of the oceanic events we studied (Tables 1 and 2, and Appendix D in the Supporting Information) ruptured through the full seismogenic layer on planar faults.

In contrast to the majority of events we studied, there is strong evidence for listric down-dip curvature in some events from Tibet and East Africa. For events showing down-dip curvature, the maximum change in dip is $20^\circ$–$30^\circ$. Although resolving this change in dip is a significant observation, it should be noted that the overall change is much less than that for the strongly listric fault geometries ($>50^\circ$ change in dip) that are often drawn on geological cross-sections. The tightness of the constraints on the geometry is similar to the planar events, that is, within $10^\circ$–$30^\circ$ of the best-fitting model geometry. The ruptures we studied that displayed down-dip curvature occurred on anhritic faults, or minor faults within the hanging wall of a rift.

The location and the inferred depth extent of the rupture for both the 2004 and 2005 Zhongba (Tibet) events indicate that these earthquakes occurred on faults that cut through the hanging wall of a major rift-bounding normal fault. Our best-fit models, of $-20^\circ$ curvature, indicate the lower portion of the active faults may dip at $\sim30^\circ$. This angle is similar to the $\sim28^\circ$ dip of the basal detachment Kapp et al. (2008) inferred to underlie the North Lunggar Rift, $\sim100$ km north-northwest of the Zhongba epicentres. The deeper structure of the South Lunggar Rift (where the Zhongba earthquakes were located) is not known. West of the Zhongba epicentres, the South Lunggar Detachment dips at $20^\circ$ at the surface and exhumes mylonites and foliated rocks with a normal sense of shear (Styron et al. 2013). However, we have found no evidence for seismogenic slip on very low-angle normal faults ($\leq20^\circ$ dip) in the Lunggar Rift, as has been debated for metamorphic core complexes in other parts of the world (Wernicke 1995; Axen 1999; Collettini 2011).

The 2009 Karonga earthquakes ruptured a series of $\sim6$–$8$ km long fault segments in the hanging wall of the Karonga Fault (which is...
Figure 18. Down-dip curvature modelling results for the 2009 December 6 Karonga earthquake. Caption as for Fig. 7.
The down-dip shape of normal faults

7 CONCLUSIONS

We have developed a new seismological technique to assess the down-dip shape of earthquake-generating faults through the analysis of body waveforms, and have successfully applied the technique to earthquakes in a variety of extensional environments. We have found that most normal-faulting earthquakes that rupture through the full seismogenic layer are planar and have dips of 30°–60°. There is evidence for faults with listric-sense down-dip curvature from earthquakes occurring in two regions; Tibet and East Africa. These ruptures occurred on antithetic faults, or minor faults within the hanging wall of a rift, and the change in dip across the fault plane is ∼30°.

ACKNOWLEDGEMENTS

This work forms part of the NERC- and ESRC-funded project ‘Earthquakes without Frontiers’ and was partially supported by the NERC large grant ‘Looking inside the Continents from Space’.

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SUPPORTING INFORMATION

Supplementary data are available at GJI online.

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