Seismic slip on an upper plate normal fault during a large subduction megathrust rupture

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Summary

Quantification of stress accumulation and release during subduction zone seismic cycles requires an understanding of the distribution of fault slip during earthquakes. Reconstructions of slip are typically constrained to a single, known fault plane. Yet, slip has been shown to occur on multiple faults within the subducting plate[1] due to stress triggering[2], resulting in phenomena such as earthquake doublets[3]. However, rapid stress triggering from the plate interface to faults in the overriding plate has not been documented. Here we analyse seismic data from the magnitude 7.1 Araucania earthquake that occurred in the Chilean subduction zone in 2011. We find that the earthquake, which was reported as a single event in global moment tensor solutions[4,5], was instead composed of two ruptures on two separate faults. Within 12 seconds, a thrust earthquake on the plate interface triggered a second large rupture on a normal fault 30 km away, in the overriding plate. This configuration of partitioned rupture is consistent with normal-faulting mechanisms in the ensuing aftershock sequence. We conclude that plate interface rupture can trigger almost instantaneous slip in the overriding plate of a subduction zone. This shallow upper plate rupture may be masked from teleseismic data, posing a challenge for real-time tsunami warning systems.

Main body

A recent succession of large ($M_w > 8$) earthquakes in circum-Pacific subduction zones has focussed attention on the relationship between physical properties and stress distribution along the megathrust plate interface. Seismic ruptures along the megathrust can be viewed as smooth and spatially varying patches of slip on a single fault; in this case, the subducting plate interface[6]. The fault geometry used in early coseismic slip models is underpinned by centroid moment tensor (CMT) solutions often reported by earthquake monitoring agencies. Although more sophisticated slip inversions use curved faults based on regional subduction geometry[7], slip is nearly always assigned to a single fault.

An alternative rupture configuration is slip occurring on separate faults due to static or dynamic triggering processes[2], resulting in phenomena such as doublets[3]. A doublet is the occurrence of two nearby earthquakes with similar magnitude. The time delay between ruptures can range from months[3] to seconds[1]. Many documented cases of subduction zone doublets involve triggering between the subducting plate interface and deep-rooted faults in the downgoing plate[3,8]. Although the implications for tsunami hazard are significant, there are no reported cases of rapid triggering from the plate interface to the upper plate, where there are complex faulting networks[9,10]. To resolve triggered faulting in such cases, dense local seismic observations are needed.

A region with a suitably dense network of seismometers is the central Chile subduction zone following the M_w 8.8 Maule earthquake in 2010. The ensuing aftershock sequence was captured in detail by the International Maule Aftershock Deployment[11]. Here, we focus on the largest interplate aftershock of the Maule sequence: the M_w 7.1 Araucania earthquake that occurred on 2 January 2011 at 20:20:18 UTC. Based on CMT solutions derived from teleseismic waveforms, the Araucania earthquake appears to be a 'straightforward' plate interface thrusting event[4,5]. Its epicentre (Supplementary Note 1, Supplementary Table 1) lies in a region that acted as a barrier during the 1960 M_w 9.5 Valdivia[12] and 2010 M_w 8.8. Maule[6,11] earthquakes (Fig. 1). Moreover, the upper plate in this region is heavily faulted[6,10]. Therefore, the Araucania earthquake is an ideal candidate to examine possible interactions between the plate interface and upper plate faults.

We employed a multiple point-source inversion of regional seismic data[13,14]. Compared to conventional slip inversions along pre-defined fault planes[6] and single point-source CMT inversions[5], we can retrieve centroid times of sub-events and permit multiple faulting styles on a grid of trial point-sources. A detailed understanding of 3-D crustal velocity structure[15,16] ensures robust waveform inversion. Synthetic tests (Supplementary Note 2) show that we can accurately resolve a range of extended source configurations involving offshore rupture using the available station distribution.

Using the observed data, we first investigated whether low-frequency waveforms (0.02–0.04 Hz) in the near-field could represent the earthquake as a single point-source. The optimal regional CMT solution provides a good fit to the observed waveforms at most stations (Supplementary Figure 2). The centroid lies close to our relocated epicentre; its mechanism is consistent with the teleseismic GCMT and USGS solutions (Fig. 1), indicating thrusting along the plate interface. A high double-couple percentage (%DC) indicated by the global (98%) and our regional solutions (85%) suggests a simple faulting mechanism. When we increase the upper frequency limit to >0.06 Hz, waveform variance reduction (VR) sharply decreases and, at the upper limit of 0.08 Hz, we notice two clear arrivals in the observed waveforms (Fig. 1b & Supplementary Figure 3). Therefore, the next step is to consider whether a complex source can be resolved using higher frequency waveforms and a multiple point-source parameterisation.

A two-point-source model is a logical progression; an $M_w \sim 7$ earthquake likely comprises no more than two to three patches of slip[17]. Compared with that of using the first source alone (VR = 0.57), introduction of the second source significantly increases the waveform fit (Fig. 2a) by 30% (VR = 0.73), which is statistically significant to within the 99.5% confidence interval (Supplementary Note 3). Mechanisms at each trial point-source position are very consistent, with sharp correlation maxima (Fig. 2b). Based on our results, we can confidently identify the following sequence of events, which can be regarded as a closely-spaced doublet (CSD), both in time and space. Following nucleation, Event I (M_w 6.8) ruptured the megathrust beneath the coast. No more than twelve seconds later, Event II (M_w 6.7) ruptured to the southwest at a shallower depth and with an oblique normal faulting mechanism (Fig. 2).

Locations and mechanisms of aftershocks (Supplementary Note 4) that followed the Araucania earthquake support this CSD configuration. From our 44 relocated events, it is clear that there are two distinct groups of aftershocks (Fig. 3a). One group is located in the coastal region (hereafter, Group A); the other 30–40 km to the southwest (hereafter, Group B). Group B aftershocks have shallower depths, located within the marine forearc, up to 9 km above the plate interface (Fig. 3b). We obtained 19 robust CMT solutions from this aftershock sequence (Fig. 3), all of which have depths in agreement with their hypocentral location, based on a 3-D velocity model and ocean-bottom observations[15]. Normal faulting mechanisms dominate aftershock Group B. Group A aftershocks comprise mixed mechanisms, but interplate thrust faulting is most common.

A puzzling location discrepancy between Event II and aftershock Group B (Supplementary Figure 4) leads us to assess location bias in the multiple point-source inversion. So far, we have computed synthetic seismograms by calculating Green’s functions in a 1-D velocity model. However, in the shallow regions of subduction zones, there are strong lateral velocity gradients (Figs. 3 & Supplementary Figure 5), particularly in S-wave velocity. Therefore, a more realistic velocity model can improve waveform fits and make source inversions more stable. To account for lateral velocity variations, we simulated waveforms in a 3-D velocity model[15,16] using the spectral element code SPECFEM3D[18]. We used 3-D synthetics based on our two-point-source solution as input to a multiple point-source inversion using 1-D Green's functions. While the position of Event I remains stable, we find that the inversion shifts Event II 12 km to the south (from position 16 to 15; Supplementary Figure 6). Similarly, when we simulate the waveforms from Event II at the location of aftershock Group B (position 17), we find a similar southward shift, as implied from the real data inversion. Therefore, it is likely that Event II occurred ~12 km northward with respect to the formal inversion result of Fig. 2b (Supplementary Table 5). This result demonstrates the importance of 3-D structural models to obtain accurate source parameters of offshore subduction earthquakes.

Based on our aftershock analyses and 3-D waveform simulations it is now clear that Event II ruptured on a normal fault near the base of the overriding crust (Figs. 3 & 4). Group B aftershocks are located close to the prominent Mocha-Villarrica fault zone (Fig. 3). This fault may be related to strong velocity contrasts in the marine forearc beneath Isla Mocha, where Group B aftershocks are located (Figs. 3 & 4). Crustal faulting in the region is pervasive and may extend through the entire crust[10,19]; it is plausible that the geometry of fault networks becomes more complex at the base of the forearc with possible conjugate faulting (Fig. 4). We speculate that these faults are compressional during the interseismic period, but a stress inversion following the Maule earthquake[20] may favour post-seismic extension. Based on approximate fault areas from scaling relations[21], the two fault planes of Events I and II likely do not intersect. There are several possible mechanisms for the triggering of a rupture by a preceding earthquake. Dynamically triggered rupture of the normal fault is likely the dominant failure mechanism given that Event II's centroid time coincides with the passage of highamplitude S-wave arrivals from Event I (Supplementary Figure 7). However, we cannot completely rule out static stress transfer acting as a partial trigger.

To our knowledge, these results provide the first documented case of plate interface thrusting instantaneously activating a large rupture in the overriding plate through dynamic triggering. Past subduction zone doublets have been identified by high non-double components in their CMT solutions[8]. Conversely, in the case of the Araucania earthquake, the low-frequency single point-source solutions of both the global and regional CMT solutions did not yield a low %DC (Fig. 1). This discrepancy is also evidenced by our synthetic tests. It is possible that the short time delay and small distance between Events I and II masks rupture complexity in teleseismic CMT solutions. Therefore, CSDs may be completely hidden from global networks. CSDs may, however, be detected from a greater proportion of high frequency radiation in regional waveforms (Supplementary Note 5, Supplementary Figure 8), although this character may depend on several other source parameters, such as rupture duration. CMT solutions provided by global reporting agencies are accepted by the seismological community and form the basis of slip inversions and examinations of the stress field. CMTs are, therefore, a pillar of earthquake science, yet our results recommend their careful use in the case of slip on multiple fault planes.

The precedent set by this study also presents a new perspective for tsunami hazard assessment in subduction zones. Reverse faults as well as normal faults could theoretically be immediately triggered by megathrust slip, causing large seafloor displacement. A wide variety of upper plate faults are present in many subduction zones. For example, steeply-dipping normal faults have been imaged in the upper plate along the N. Chile and S. Peru margins[9]. Furthermore, a large upper-plate reverse faulting event preceded the 2014 M_w 8.2 Pisagua, N. Chile earthquake[22] and backthrust faults are widespread in the Sumatra subduction zone[23]. An M_w 7.0 rupture in the upper plate could result in substantial vertical seafloor displacement of 1.2 m (Supplementary

Note 6). If this scenario were scaled up to a larger rupture (M_w ~7.5), slip on the forearc fault could cause a localised tsunami on the continental shelf, although the upper limit of rupture size is controlled by the geometry and frictional properties of these faults (Fig. 3). A tsunami may be caused by static vertical displacement or through submarine landslides (Fig. 4), which have occurred locally in the past[24]. Yet without local strong-motion instruments, GPS networks, or close inspection of regional waveforms, near-field triggered ruptures will be difficult to detect. We speculate that the lack of evidence for Event II in single-source CMT solutions may result in part from the short timing between the two sources. Therefore, we recommend that the capability of teleseismic CMT inversions to resolve different doublet configurations is given a full assessment. Furthermore, there is a need to re-evaluate CMT solutions for large earthquakes using local and regional waveforms in subduction zones globally to examine whether CSDs involving the upper plate are ubiquitous.

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Supplementary information

Supplementary information is linked to the online version of the paper at www.nature.com/nature.

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Author contributions

S.P.H. carried out the single and multiple point-source inversions, as well as the moment tensor inversion and aftershock relocations. S.P.H. wrote the manuscript, interpreted the results, and generated all figures. A.R. carried out the 3-D full waveform simulations, wrote the manuscript, and interpreted the results.

Author information

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Figure Captions

Fig. 1: Location and single source solution. a) Location map. Stations used for CMT inversion are labelled with station codes. Other stations are for hypocentre relocation only (Supplementary Note 1). Shading indicates rupture areas of great earthquakes in 1960[12] and 2010[6]. Inset: Regional tectonic setting. b) Double-couple percentage (%DC) and variance reduction (VR) of the single point-source versus frequency. A transition occurs at 0.057 Hz, where VR suddenly decreases because the waveforms cannot be explained by a single source alone. This change is illustrated by representative waveforms at low and high frequencies (see Supplementary Figure 2 & Supplementary Figure 3 for details).

Fig. 2: Two-point-source solution. a) Observed (black) and synthetic (red) waveforms for the optimum high-frequency (0.02–0.08 Hz) solution. Station names are labelled. Numbers alongside each waveform component denote VR. Blue and green shading denotes the contribution from each event. b) Waveform correlation for each event as a function of trial point-source position (numbered). The optimum time shift of Event I and II is shown. Black beach balls are solutions that lie within 90% of the optimum solution's (red beach ball) VR. The red star denotes the earthquake's epicentre. c) Resulting moment-rate function obtained using the NNLS method.

Fig. 3: Aftershock analysis. (a) Map and (b) cross-section showing locations and focal mechanisms of aftershocks (Groups A and B) and mainshock events (labelled EV-I and EV-II). Faulting style is classified on principal stress orientations[28] and minimum rotation angle with respect to plate interface thrust faulting[29], accounting for plate interface geometry (black line)[15,16]. We plot the revised location of Event II, based on 3-D waveform modelling. Mapped faults are shown[10,30]; MVFZ = Mocha-Villarrica fault zone. The cross-section background is from P-wave velocity tomography models[15,16]. The star denotes the hypocentre of the Araucania earthquake; the triangle shows the coastline.

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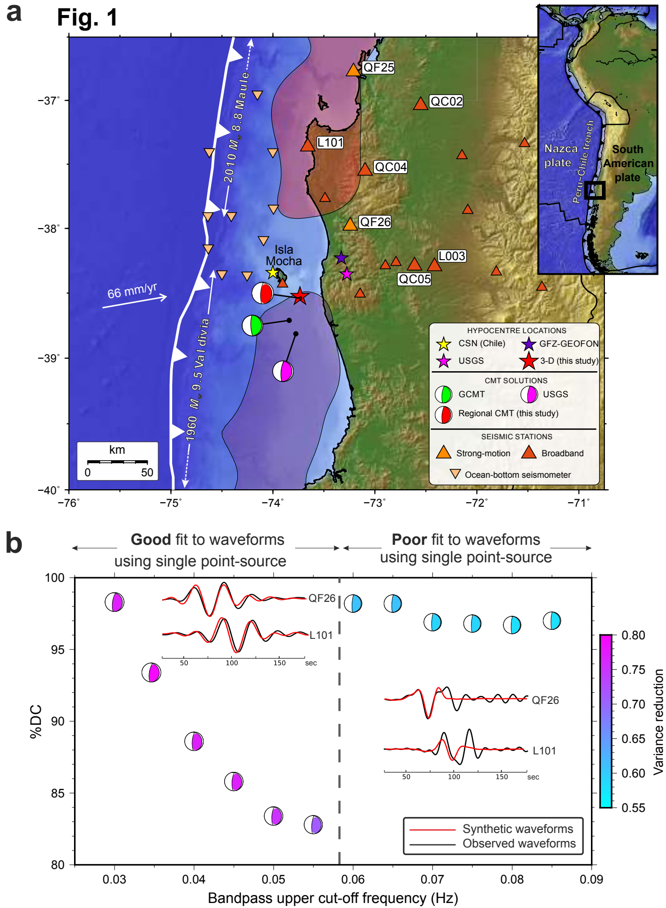
Fig. 4: Schematic interpretation of the Araucania earthquake rupture. Plate interface thrusting (Event I) triggered a rupture along an extensional fault in the overriding plate (Event II). It is likely that two great earthquakes in 1960 and 2010 brought both faults closer to failure. As shown by ancient submarine landslide deposits in the area, a larger-scale rupture in the overriding plate has the potential to act as a tsunamigenic earthquake. Beach balls represent the focal mechanisms of both events from Fig. 3. The inset shows the interpreted structure of conjugate normal faulting with the background colour representing v_p/v_s ratio[15].

Fig. 1



a
QF25
QC02
L101
QC04
QF26
L003
QC05
Isla Mocha
2010 Mw 8.8 Maule
1960 Mw 9.5 Valdivia
66 mm/yr
km
0
50
Nazca plate
South American plate
Peru-Chile trench
HYPOCENTRE LOCATIONS
CSN (Chile)
GFZ-GEOFON
USGS
3-D (this study)
CMT SOLUTIONS
GCMT
USGS
Regional CMT (this study)
SEISMIC STATIONS
Strong-motion
Broadband
Ocean-bottom seismometer
−37°
−38°
−39°
−40°
−76°
−75°
−74°
−73°
−72°
−71°
b
Good fit to waveforms using single point-source
Poor fit to waveforms using single point-source
QF26
L101
50
100
150
sec
%DC
Bandpass upper cut-off frequency (Hz)
Variance reduction
0.80
0.75
0.70
0.65
0.60
0.55
Synthetic waveforms
Observed waveforms
100
95
90
85
80
0.03
0.04
0.05
0.06
0.07
0.08
0.09

Fig. 2

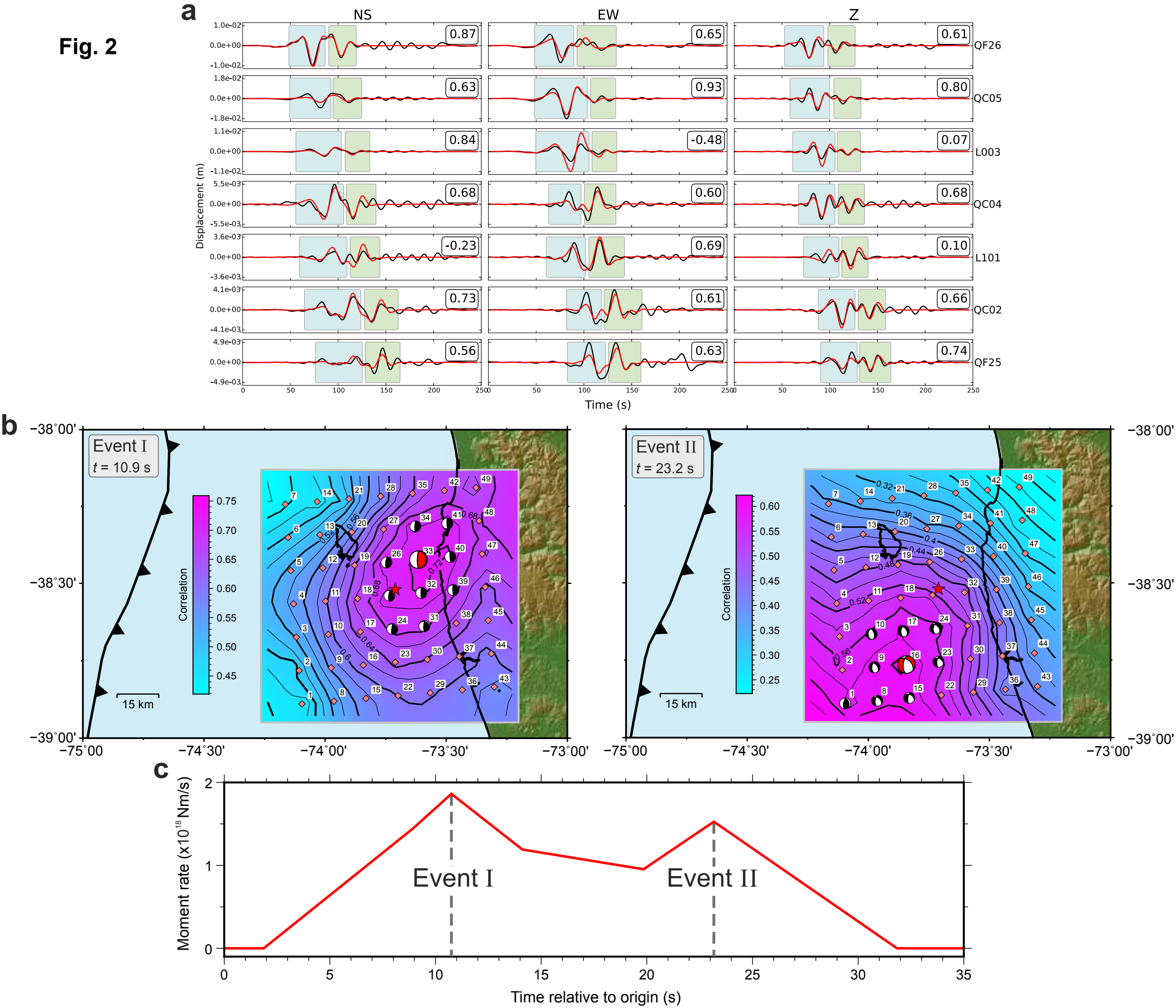
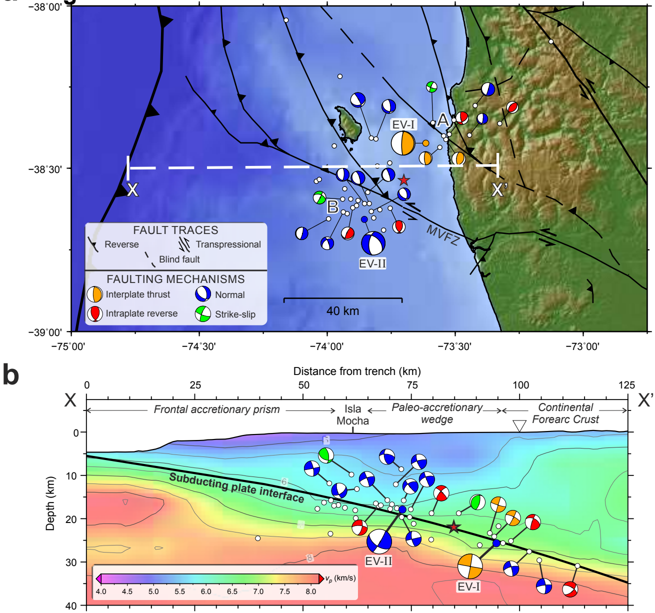

a
NS
EW
Z
QF26
QC05
L003
QC04
L101
QC02
QF25
Displacement (m)
Time (s)
b
Event I
t = 10.9 s
Event II
t = 23.2 s
Correlation
15 km
c
Moment rate (x10^18 Nm/s)
Event I
Event II
Time relative to origin (s)

Fig. 3


a
-38°00'
-38°30'
-39°00'
-75°00'
-74°30'
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-73°30'
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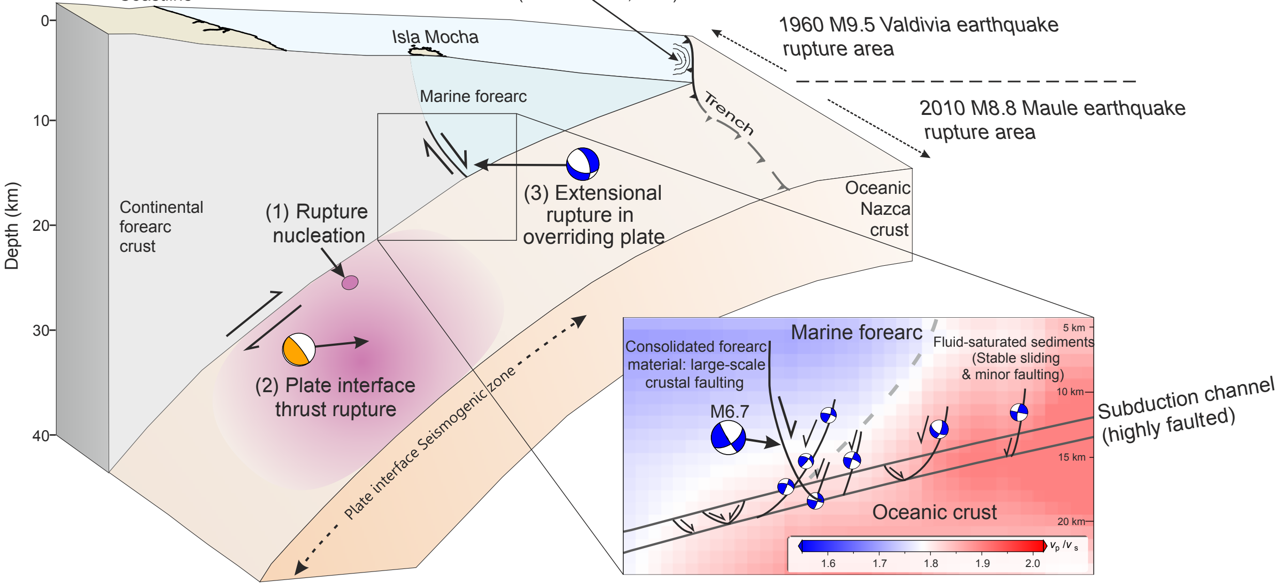
X
X'
A
B
EV-I
EV-II
MVFZ
40 km
FAULT TRACES
Reverse
Transpressional
Blind fault
FAULTING MECHANISMS
Interplate thrust
Normal
Intraplate reverse
Strike-slip
b
Distance from trench (km)
0
25
50
75
100
125
X
X'
Frontal accretionary prism
Isla Mocha
Paleo-accretionary wedge
Continental Forearc Crust
Depth (km)
0
10
20
30
40
Subducting plate interface
EV-II
EV-I
v_p (km/s)
4.0
4.5
5.0
5.5
6.0
6.5
7.0
7.5
8.0

Fig 4.


Vertical exaggeration: 2.5x
Horizontal scale:
40 km
Coastline
Slope failure deposits
(Geersen et al., 2011)
1960 M9.5 Valdivia earthquake
rupture area
2010 M8.8 Maule earthquake
rupture area
Isla Mocha
Marine forearc
Trench
Depth (km)
0
10
20
30
40
Continental
forearc
crust
(1) Rupture
nucleation
(3) Extensional
rupture in
overriding plate
Oceanic
Nazca
crust
(2) Plate interface
thrust rupture
Plate interface Seismogenic zone
Marine forearc
Consolidated forearc
material: large-scale
crustal faulting
Fluid-saturated sediments
(Stable sliding
& minor faulting)
M6.7
5 km
10 km
15 km
20 km
Subduction channel
(highly faulted)
Oceanic crust
v_p / v_s
1.6
1.7
1.8
1.9
2.0

Methods

Data selection and processing

For the waveform inversion of the Araucania earthquake, we used broadband and strong-motion stations that were located onshore within an epicentral distance of 200 km from the Araucania earthquake. We only used waveforms from stations that have a high signal-to-noise ratio (> 10) in the frequency range 0.01–0.10 Hz (Supplementary Figure 9). Due to the close proximity of some stations to the earthquake, we excluded waveform records that were either clipped, had long period disturbances, or instrument tilt effects. These quality-control checks resulted in a set of seven stations (including two strong-motion stations) located north and east of the Araucania earthquake (Fig. 1a).

Source inversion algorithm

Iterative deconvolution (ID)[14] is used for the multiple point-source inversion of deviatoric moment tensors. ID works by inverting for the optimum focal mechanism and timing of sources for a prescribed set of points to minimise the L2 misfit between observed and synthetic waveforms. A grid search is then performed to select the source position that produces the highest correlation between observed and synthetic waveforms. The first inversion explains the full waveforms using a single source, the synthetics of which are then subtracted from the observed waveforms. The remaining waveforms are then used to invert for subsequent sub-events[14]. After the retrieval of each sub-event, VR is calculated and manually assessed to ensure that additional sub-events are required by the data and the waveforms are not just fitting correlated noise. For moment tensor inversion, we use the software package, ISOLA[13], which can be accessed athttp://seismo.geology.upatras.gr/isola/. In the inversion, the moment-rate of the source is prescribed; it is found by manually searching for the source length that produces the maximum VR. If the moment-rate of the source is shorter than the minimum inverted period, then the source can be represented by a delta function. To negate artifacts produced by the ID method, we also test the stability of our multiple point-source solution using a non-negative least squares (hereafter, NNLS) inversion method[17]. In the NNLS approach, the double-couple focal mechanism at each source is prescribed. At each trial point-source position, the moment rate is represented by a set of shifted triangles. The weight of each triangle is then inverted for using NNLS. In this paper, we use one-second triangle shifts. The moment of each source can be constrained, which stabilises the inversion, although the exact value of total moment does not dramatically influence source timings or positions[25].

The inversion is performed on bandpass-filtered displacement waveforms. The effect of different 1-D velocity models was tested; the final solutions were calculated using a velocity model appropriate for the coastline of south-central Chile (Supplementary Figure 10). We analysed the effect of data errors and imperfect Green's functions by systematically removing pieces of data from the inversion (jackknifing). Where subsurface structure is complex, removal of certain stations may have a large effect on the final solution[26]. Based on the analysis of signal to noise ratio (Supplementary Figure 9), we used a lowermost frequency limit of 0.02 Hz throughout this paper. The upper frequency limit was dependent on the source parameterisation used (single or multiple source).

Single point-source inversion strategy

Guided by preliminary inversions and the anticipated fault size[21], we used a trial point-source grid with a spacing of 12 km in the down-dip and along-strike directions (Fig. 2b). At this stage, we wanted to resolve the simplest possible source, so the maximum frequency was kept well below the corner frequency (approximately 0.1 Hz for an $M_w \sim 7$ earthquake). Therefore, we chose an upper frequency limit of 0.04 Hz and assumed a delta moment-rate function. We tested the robustness of the solution by jackknifing stations and their individual components. The source position changes slightly when varying the dataset, but by no more than 17 km (Supplementary Figure 2); the largest shifts occur if the closest stations are removed from the inversion. The zone of maximum correlation is not particularly sharp, corresponding to the possible source locations from the jackknifing analysis (Supplementary Figure 2). These tests show that the source location is reasonably stable and its mechanism is consistent throughout.

We also find that as the upper frequency bandpass cut-off increases, %DC gradually decreases. This trend continues until around 0.057 Hz, above which, the full waveforms can only be explained using Events I and II, and %DC becomes very high (Fig. 1b).

Multiple point-source inversion strategy

We first carried out a multiple point-source inversion using ID, in which the deviatoric moment tensor mechanisms of both sources were allowed to vary. The grid of point-sources was kept the same as for the single point-source inversion. For the source-time function, we found that with increasing length of the triangle, the total moment gradually increases, while VR and %DC of each source reaches a maximum at 18 s (Supplementary Figure 11). We therefore fixed the triangle length of each source to 18 s for the ID multiple point-source inversion, although the point-source mechanisms remain consistent for different triangle lengths, suggesting a stable solution.

We used the NNLS method to test the robustness of the solution obtained by ID. To search for the best-fitting source configurations, we performed two inversions: one in which total moment was constrained by the ID solution; the other in which moment was allowed to vary. We tested a number of source positions and faulting styles for Events I and II using the NNLS method, but we found that the highest VR came from the two-point-source configuration found using the ID method. Using the mechanisms given by the ID solution, we then performed a grid-search over all possible combinations of the two-point-source locations using the NNLS method. As expected, the moment-constrained inversion is most similar to the ID solution (Supplementary Figure 12). Nevertheless, both inversions produce results consistent with the ID solution. Importantly, the resulting source-time function obtained by NNLS shows that both events have a similar time function to the 18 s triangle source used in ID (Fig. 2c). In summary, we find no bias in the results caused by the inversion method.

As a further test of solution stability, we perform jackknifing tests by removing one station at a time from the inversion. The results of these tests are shown in Supplementary Table 6 and demonstrate remarkably consistent centroid positions and focal mechanisms for Events I and II. The jackknifing test therefore indicates that the optimum multiple point-source solution is not dependent on one single waveform. Furthermore, a three-point-source approximation did not meaningfully improve the waveform fit (VR = 0.76; 3% increase in VR compared with two-point-sources).

Since the ID method inverts for the first point-source before subsequently calculating the second source, we carried out a test to determine whether Event II is dependent on the chosen location and mechanism of Event I. Normally, we accept the source position that produces the highest waveform correlation. However, for this test, we fixed the position of Event I and chose the corresponding best-fitting mechanism. We carried out this test at all trial point-sources adjacent to Position 33 (the optimum position of Event I). The results of this test are shown in Supplementary Table 7. For all but one position of Event I, the position, timing, and mechanism of Event II remain consistent with the optimum solution. When Event I is fixed to Position 25, the MT solution of Event II appears less stable. However, Position 25 is directly adjacent to Position 16 (the optimum location of Event II from ID; Fig. 2b), so this discrepancy is expected because the inversion tries to explain both events at this position with a single source. In summary, we find that the Event II solution is stable with respect to the exact position and mechanism of Event I.

Mesh design for the 3-D waveform simulation

For the wave propagation simulations, we constructed a hexahedral unstructured mesh using the GEOCUBIT software package[27]. The lateral resolution at the surface is 5 km, coarsening at a refinement layer (45 km depth, which is an average Moho depth for the region[15,16]). The mesh honours surface relief and bathymetry to ensure that topographic effects on waveform propagation are accurately simulated. Our mesh does not contain dipping geological discontinuities in the subsurface, such as the oceanic Moho, due to the lack of constraints on its geometry. This mesh has been designed for simulations that are accurate up to ~0.3 Hz, well above the maximum frequency of our waveform inversions, ensuring numerically stable simulations. The Mesh used is shown in Supplementary Figure 13.