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Seismic images of the Northern Chilean subduction zone at 19°40'S, prior to the 2014 Iquique earthquake

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SUMMARY

The Northern Chilean subduction zone is characterized by long-term subduction erosion with very little sediment input at the trench and the lack of an accretionary prism. Here, multichannel seismic reflection (MCS) data were acquired as part of the CINCA (Crustal Investigations off- and onshore Nazca Plate/Central Andes) project in 1995. These lines cover among others the central part of the M_W 8.1 Iquique earthquake rupture zone before the earthquake occurred on 1 April 2014. We have re-processed one of the lines crossing the updip parts of this earthquake at 19°40'S, close to its hypocentre. After careful data processing and data enhancement, we applied a coherency-based pre-stack depth migration algorithm, yielding a detailed depth image. The resulting depth image shows the subduction interface prior to the Iquique megathrust earthquake down to a depth of approximately 16 km and gives detailed insight into the characteristics of the seismogenic coupling zone. We found significantly varying interplate reflectivity along the plate interface which we interpret to be caused by the comparably strong reflectivity of subducted fluid-rich sediments within the grabens and half-grabens that are predominant in this area due to the subduction-related bending of the oceanic plate. No evidence was found for a subducted seamount associated to the Iquique Ridge along the slab interface at this latitude as interpreted earlier from the same data set. By comparing relocated fore- and aftershock seismicity of the Iquique earthquake with the resulting depth image, we can divide the continental wedge into two domains. First, a frontal unit beneath the lower slope with several eastward dipping back-rotated splay faults but no seismicity in the upper plate as well as along the plate interface. Secondly, a landward unit beneath the middle slope with differing reflectivity that shows significant seismicity in the upper plate as well as along the plate interface. Both units are separated by a large eastward dipping mega splay fault, the root zone of which shows diffuse seismicity, both in the upper plate and at the interface. The identification of a well-defined nearly aseismic frontal unit sheds new light on the interplate locking beneath the lower continental slope and its controls.

Key words: South America; Image processing; Controlled source seismology; Crustal imaging; Seismicity and tectonics; Continental margins: convergent.

1 INTRODUCTION

The continental margin of Chile is a seismically highly active area. Large megathrust earthquakes occur along the plate interface of this subduction zone (Bilek 2010; Contreras-Reyes & Carrizo 2011; Ruiz & Madariaga 2018). A large seismic gap extended from Southern Peru down to the Mejillones peninsula of Northern Chile until 1 April 2014; when the M 8.1 Iquique earthquake ruptured a part of this subduction zone segment. The whole series of fore- and aftershocks of this event were recorded by the IPOC seismic network and analysed in detail by several studies including Schurr *et al.* (2014) and Sippl *et al.* (2018). Our goal is to link earthquake rupture

processes with seismic structures imaged in depth, to get a better understanding of the plate boundary structure during the earthquake cycle. In this context, the Iquique Ridge seafloor topography seems to play an essential role on earthquake generation, as described by Contreras-Reyes & Carrizo (2011) and Geersen *et al.* (2015).

The Northern Chilean subduction zone is a continental convergent margin undergoing long term subduction erosion. Von Huene *et al.* (2004) introduced a generic model of subduction erosion that is the most prevalent type at convergent margins. Subduction erosion depends on the sediment input into the trench and of subduction of high relief features on the incoming oceanic plate. In Northern Chile the sediment input is extremely low due to the hyperarid climate of

the Atacama desert preventing relevant sediment supply. Here, the Iquique ridge, an oceanic feature with distributed volcanic edifices, enters the subduction zone between S19° and S21°30' latitude (Clift & Vannucchi 2004; Ranero *et al.* 2006; Contreras-Reyes *et al.* 2012; Geersen *et al.* 2015; Bello-González *et al.* 2018). Besides Northern Chile (19°–24°S), other examples for subduction erosion are offshore Costa Rica, Tonga, Peru, North-Central Chile (28°–33°S) and northern Japan (Clift & Vannucchi 2004). From all of these margins, multichannel seismic reflection (MCS) data, acquired and evaluated together with other data, such as multibeam bathymetry, earthquake data, wide angle seismic data and some ocean drilling data, have helped to shape our understanding of the plate interface architecture and, in some cases, the relationship to earthquake nucleation and propagation in erosive settings.

In offshore northern Japan, MCS data were acquired between 1996 and 1999. Tsuru *et al.* (2002) processed 2-D lines at the Japan Trench and analysed the interplate sedimentary units. Bangs *et al.* (2004, 2006) constructed a 3-D cube at the Nankai Trough, using both, time and pre-stack depth migration methods. They derived a decollement-amplitude map of the Nankai Trough subduction thrust and identified a small (up to 1 km high) basement ridge that has been subducted 65 km away from the present deformation front.

At the Costa Rica convergent margin, pre-stack depth migration images (Ranero & von Huene 2000) helped to identify two mechanisms of basal erosion: erosion by seamount tunnelling and removal of large rock lenses of an extending upper plate. Data from an Integrated Oceanic Drilling Program (IODP) expedition (Vannucchi *et al.* 2013) were interpreted together with pre-stack depth migrated 3-D reflection seismic images (Vannucchi *et al.* 2016, 2017) that led to the model of a 'depositional forearc', a mode of rapid basin accretion during an erosive event at a subduction margin. They further suggest that the geological structures found offshore Costa Rica were deformed during seismic slip-to-the-trench events instead of aseismic creep. Post-stack time migration of multichannel seismic data, interpreted together with velocity models from wide-angle seismic profiling, bathymetry and earthquake data (Martínez-Loriente *et al.* 2019) have revealed a clear image of subducted seamounts 18 and 40 km away from the trench, and well resolved tectonic structures in the overriding plate. Here, the M_w 6.4 2002 Osa earthquake nucleated at the seamount's leading flank.

While Costa Rica is a tropical country and the reduction of sediment supply is only local due to the collision of the Cocos Ridge, Tonga is an oceanic-oceanic subduction setting far away from continental regions causing the lack of sedimentation to the Tonga trench axis. The subduction collision of high-relief oceanic features along the Louisville Ridge accelerates subduction erosion in Tonga (Ballance *et al.* 1989; Contreras-Reyes *et al.* 2011; Robinson *et al.* 2018). Ballance *et al.* (1989) evaluated multichannel seismic data and state that seamounts in this environment (hot-spot ridge collision with a sediment-starved trench) are subducted and the inner trench wall is tectonically eroded towards the arc.

The subduction or collision of high-relief oceanic features that accelerate subduction erosion have also been observed offshore north-central Chile, where the Juan Fernández Ridge subducts beneath the South American Plate. Contreras-Reyes *et al.* (2014, 2015), von Huene *et al.* (1997) and Becerra *et al.* (2017) interpreted bathymetric data, multichannel seismic and wide-angle seismic data to derive a detailed 2-D velocity depth model and high-resolution seismic time images. Their results show the subduction erosion processes caused by the quasi-stationary collision of the Juan Fernández Ridge with the continental South American Plate. Offshore Peru, the Nazca Ridge subducts beneath the South American Plate. MCS data were

collected and studied by Krabbenhöft *et al.* (2004) and Hampel *et al.* (2004). They used time migration methods alongside forward modelling based on wide angle seismic data, gravimetry and bathymetry to define structure and P -wave velocities of the subducting oceanic plate including the Nazca Ridge, and the overriding South American Plate. Their results provide evidence for subduction erosion processes, that are temporarily enhanced here by the Nazca Ridge subduction. Contreras-Reyes *et al.* (2019) processed wide angle seismic data, gravity modelling and seismological constraints, and explain the role of the uplifted and smooth seafloor topography of the Nazca Ridge on subduction erosion processes off Peru. Southern Peru is particularly interesting, since as Das & Watts (2009) stated earthquakes in the southern Peru region have been followed by one in northern Chile within about a decade, more than once in known history. Both earthquakes, the 2001 southern Peru and the 2014 Iquique earthquake occurred on the same seismic gap on the plate boundary between the Nazca Plate and the South American Plate. Rupture processes of both earthquakes are very likely affected by the topography of the subducting seafloor. Further, the seismic velocity structure (V_p) of the overriding plate found off Peru are similar to the velocity structure found in this study off Iquique.

All these convergent margins host large earthquake rupture areas that spatially correlate with the presence of subducting oceanic topographic features that seems to play an essential role on the earthquake rupture process (Contreras-Reyes & Carriazo, 2011). To analyse this in more detail, it is important to have a high resolution depth image of the subduction interface and the overlying continental margin in the vicinity of the rupture area. Particularly in subduction zones that have large seismic velocity variations, a significant ocean floor topography and steeply dipping faults, pre-stack depth migrated images outmatch the time migrated images.

The Northern Chilean subduction zone has been a target of major scientific interest and many onshore and offshore geophysical investigations were performed within the last decades (Ranero *et al.* 2006; Oncken *et al.* 2006). Among those studies were 23 seismic profiles (4494 km total length) acquired in 1995 within the project CINCA (Crustal Investigations off- and onshore Nazca Plate/Central Andes, Hinz *et al.* 1998) that partly covered the rupture area of the 2014 Iquique earthquake (Fig. 1). Hence, for this area there is a unique seismological data set capturing a megathrust earthquake and its preparation phase with great instrumental resolution (Schurr *et al.* 2014) in an area that has been previously imaged with geophysical methods at equally high resolution.

Sick (2006) and Sick *et al.* (2006) applied a focusing pre-stack depth migration approach to two seismic lines from the CINCA'95 data set that are located further south of the Iquique earthquake rupture area: line 7 (21°S) and line 13 (23.25°S). By applying a Fresnel Volume Migration technique, resolution is enhanced significantly and artefacts are reduced in comparison to Kirchhoff pre-stack depth migration. These depth images yield high resolution images of structures of the upper crust, but reflections from the plate interface are very weak and fade by 10–15 km downdip of the subduction front.

Line 22 of the CINCA multichannel seismic data is well located, at 19.7°S, to investigate the rupture processes of the Iquique earthquake in detail. Therefore we reprocessed this line with a new focusing, coherency-based pre-stack depth migration approach, for a depth image with higher resolution and fewer artefacts (Hlousek *et al.* 2015), to enhance the visibility of interface structures before a large megathrust rupture. It partly covers the main earthquake rupture area (Fig. 1) and passes the hypocentral area. Here, interface reflections are traceable until 45 km downdip of the subduction

Figure 1. Map of the study area. The MCS lines from the CINCA'95 cruise (offshore) and the ANCORP'96 experiment (onshore, Storch *et al.* 2016) are marked. Line 22, processed and interpreted here, is highlighted. Big stars indicate the Iquique main shock on 1 April 2014 (M_W 8.1) and the largest after shock (M_W 7.6). Contour lines around the main shock area indicate the rupture displacement in meters (Schurr *et al.* 2014).

front. Based on the imaging results, we provide a characterization of the subduction interface in the rupture area prior to the Iquique earthquake and compare it with the location of both the main slip zone, as well as fore- and aftershocks (Sippl *et al.* 2018; Schurr *et al.* 2020), and with results from other geophysical studies, such as refraction interfaces from wide-angle analysis of this profile (Patzwahl *et al.* 1999) and density-depth-modelling from Maksymowicz *et al.* (2018). Furthermore, we discuss the tectonic implications of the imaged structures in the frontal unit and the sediment layers, the structures in the continental crust, as well as the large splay faults.

2 TECTONIC SETTING

The western edge of South America has been shaped by a subduction setting since at least Jurassic times (Ramos 2009). Since the Middle Cenozoic, after the break-up of the Farallon Plate at about 23 Ma (Lonsdale 2005), the Andean margin was controlled by the subduction of the almost sediment-free Nazca Plate beneath South America. Convergence was always slightly oblique and the velocity has been decreasing throughout the Neogene (Norabuena *et al.* 1999) to the present value of 68 mm a^{-1} (Angermann *et al.* 1999). The presence of the cold, upwelling Humboldt Current offshore central South America and the stability of South America's latitudinal position throughout the Cenozoic (and probably since the Jurassic) favoured long-term aridity with the formation of the Atacama Desert (Hartley *et al.* 2005). Therefore, very minor amounts of terrigenous sediment have covered the forearc slope and entered the trench at the north Chilean latitude, with sedimentation confined to the Neogene (Bandy & Rodolfo 1964); sediment thickness has remained below 500 m at most, and more generally below 200 m (Ranero *et al.* 2006; Oncken *et al.* 2006).

The Northern Chilean margin is characterized by subduction erosion that is attributed to the high roughness of the subducting seafloor in combination with the low sedimentation rate (Rutland 1971; Lallemand *et al.* 1994; von Huene *et al.* 1999; Ranero *et al.* 2006). This process involves tectonic abrasion of the upper plate forearc base by the rough surface of the nearly sediment-free subducting Nazca Plate thereby continuously removing upper plate material. As a consequence, the magmatic arc front has migrated more than 200 km to the east since the Jurassic, maintaining a roughly constant distance between arc and trench (Rutland 1971; Scheuber *et al.* 1994). The Jurassic arc is now exposed in the volcano-sedimentary sequence of the La Negra Formation along the Coastal Cordillera. Dredging undertaken during the 'CINCA' cruise recovered samples of volcanic rocks from the frontal wedge at water depths of about 4000 m. von Huene & Ranero (2003 and references therein) summarize results suggesting that these are equivalent to the Jurassic La Negra formation that also forms the Coastal Cordillera. Hence, the Jurassic forearc has been removed by subduction erosion and the Jurassic arc-type rocks of the La Negra Formation (an extensive series of mafic lava flows, dykes and intrusives intercalated by marine and terrestrial sediments; e.g. Buchelt & Cancino 1988) form the present offshore crustal framework.

The present 15 to 20 km-wide prism between the trench and the seaward edge of the continental crust is thought to consist of debris resulting from gravitational failure of the foremost lower slope (von Huene & Ranero 2003). Neogene tectonism of the forearc area is suggested to have been dominated by Late Neogene to recent subsidence, induced by subduction erosion and dominantly normal faulting in the offshore forearc (Clift *et al.* 2003; von Huene & Ranero 2003; Geersen *et al.* 2018a). Extensional tectonics at the surface is also observed in the Coastal Cordillera and the Central Valley (Hartley *et al.* 2000; Allmendinger & Gonzalez 2009). However,

space geodetic techniques have shown that the forearc faults may change their kinematics from extensional to contractional during the seismic cycle as evidenced by INSAR and surface observations associated with the 2007 Tocopilla earthquake in Northern Chile (Shirzaei *et al.* 2012). Similarly, Schurr *et al.* (2012, 2014, 2020) have observed forearc reverse faulting both, following the Tocopilla earthquake and leading up as foreshocks to the 2014 Iquique earthquake (see also Gonzalez *et al.* 2015). In both cases, reverse faulting in the forearc coincided with the updip end of the megathrust rupture.

The structure of the subducting oceanic Nazca Plate is dominated by extensional grabens and half-grabens with mostly landward dipping faults through the oceanic crust. As imaged from swath bathymetry data (von Huene *et al.* 1999; Geersen *et al.* 2015), they form at the outer bulge and are transported into the subduction zone. They cut acutely across an earlier, weaker extensional fabric that likely originated as oceanic spreading fabric at the East Pacific Rise. Typically, the spacing of the extensional faults and the associated graben widths forming at the outer bulge are <10 km and 3–7 km, respectively. On the Nazca Plate, the Iquique ridge (marked in Fig. 1) forms a broad elongate structure with subdued topography trending NE–SW that is accentuated by a few inactive volcanic structures. Its average elevation off the background ocean floor is some 1000 m with internal relief varying between about 300 and 1500 m with an irregular spacing of individual volcanic edifices of 15–50 km from each other. Rarely, individual peaks may reach more than 2000 m elevation relative to the surrounding seafloor. Recently, the Iquique Ridge started to subduct in northernmost Chile at 18–21°S and, as yet, has not left a significant topographic or geochemical trace in the upper plate (Rosenbaum *et al.* 2005; Bello-Gonzalez *et al.* 2018). Along the Chilean margin, the role of subducting oceanic ridges in limiting or localizing earthquake rupture segments has been identified as a major control on the spatial distribution of historical earthquakes (Contreras-Reyes & Carrizo 2011; Bello-Gonzalez *et al.* 2018). Building on these findings, Geersen *et al.* (2015) interpret reprocessed CINCA reflection seismic lines and the forearc topography at the site of the 2014 Iquique earthquake hypocentre to show subducted seamounts to possibly have focused rupture initiation and to delimit the ruptured area.

3 THE DATA SET AND PRE-PROCESSING

In 1995, 23 air gun seismic profiles were acquired offshore Northern Chile between 19°S and 26°S with the research vessel SONNE (Fig. 1, Hinz *et al.* 1995, 1998). Most of these profiles are oriented in the E–W direction perpendicular to the trench. With a length of approximately 150 km and longer, they cover the oceanic Nazca Plate and the continental slope. A few profiles extend parallel to the coast in N–S direction. The aim of the CINCA project was to investigate the oceanic and the continental crust, to detect tectonic structures from the continental margin and the adjacent oceanic plate and thus to understand the geology and tectonic mechanisms of the North Chilean subduction zone. Apart from offshore seismic profiles, wide angle data as well as gravimetric, magnetic, parasound and hydrosweep data were acquired.

Five of the CINCA air gun profiles cover at least part of the rupture area of the 2014 Iquique earthquake (Fig. 1). Using a coherency based pre-stack depth migration approach (Hlousek *et al.* 2015), we have reprocessed the northernmost line 22 that is located closest to the main shock and crosses the centre of the rupture area.

Table 1. Acquisition parameters of CINCA line22.

Profile length	180 km
Streamer	2475 m
	99 channels
	25 m spacing
Shots (air gun)	3525
	50 m spacing
Streamer depth	14 m
Air gun depth	9 m
Registration time	14 s
CMP fold	29

This 180-km-long profile, line 22 comprises 3525 shots with a shot point spacing of 50 m and a streamer cable length of 2475 m consisting of 99 channels with 25 m channel spacing (Table 1). The profile spans over the continental margin, the trench and the oceanic plate. For that reason, a significant ocean bottom topography has to be considered during seismic data processing with water depths between 0.8 and 7.3 km, steep slopes at the continental margin and rough topography with significant horst-and-graben structures on top of the oceanic crust.

The processing flow applied to the data set is summarized in Table 2. The water bottom multiple was effectively suppressed by a 2-D Surface-Related Multiple Elimination (SRME) approach, which worked well for water depths less than approximately 2 km. For greater water depths and hence later multiple arrival times and also a steeper continental slope, it was not possible to sufficiently eliminate the multiple energy using this SRME approach. Other multiple suppression techniques such as wave equation multiple elimination (WEMR), Radon transform, deconvolution or FK-filtering after normal moveout correction also did not yield satisfying results for that case. In the end, all multiples with two-way traveltimes greater than 6 s were muted. This resulted in no major loss because no structures of interest are affected in the resulting image for such long traveltimes.

Data enhancement using predictive deconvolution and FK-filtering was performed after SRME to suppress side scattering effects from the ocean bottom topography and also to further eliminate multiple energy, followed by an automatic gain control (AGC) for amplitude equalization and a final bottom mute to eliminate remaining multiple energy.

4 VELOCITY MODEL AND PRE-STACK DEPTH MIGRATION

The application of pre-stack depth migration algorithms requires a reliable interval velocity model. Standard velocity analysis using NMO velocities derived from CDP gathers is not reliable in an environment where few reflection events are clearly visible, which apart from the ocean bottom reflection is the case here for the western part of the profile line. At the continental margin, more reflections are present in the raw data, but they mainly originate from dipping reflectors. For that reason NMO velocities do not provide useful velocity values.

To address this, we have gathered velocity information published by various authors for nearby seismic lines that run parallel to line 22, for example from wide angle measurements analyzed by Hinz *et al.* (1998), Patzwahl *et al.* (1999) and Contreras-Reyes *et al.* (2012). We assume that in this subduction zone setting velocities do not vary significantly in the North-South direction within one

Table 2. Data processing including 2-D Surface-Related Multiple Elimination (SRME).

	Data processing
Bandpass filter	Ormsby bandpass with edge frequencies 3–6–90–120 Hz
True amplitude recovery	Multiply amplitude by $t^{0.5}$
Top mute	Mute above ocean bottom reflection
	Surface Related Multiple Elimination (SRME)
SRME regularization	Regularization of ensembles in preparation for 2D SRME
SRME macro	Model multiple energy by performing 2-D SRME
SRME unregularization	Restore traces to their original offsets; the multiple model
SRME match filter	Match modelled multiple with multiple energy in the data
SRME adaptive subtraction	Adaptively subtract modelled multiple from data
	Data enhancement
Predictive deconvolution (minimum phase)	1500 ms operator length + Bandpass 4–8–70–120 Hz
FK-filter (windowed proc. OB reflection time + 0.5 s)	Time dependent fan filter, with 3 to 60 Hz
Automatic gain control	Mean in window with length of 0.5 s
Bottom mute	Bottom mute of ocean bottom multiple if arrival is between $t = 6$ –14 s
Top mute	Mute above ocean bottom reflection

litho-tectonic unit so that we can adapt the velocities from, for example the parallel profile line 7 to line 22. The ocean bottom topography has been interpolated from the water depth measured at each source position and afterwards the resulting surface has been used as a geometrical constraint for the upper boundary of the corresponding lithological units to define the velocity model for greater depths. The dip of the slab interface and the boundaries of major lithological units in the forearc and in the frontal unit were defined similarly in an iterative way, meaning that different velocity models were calculated, evaluated and adjusted according to their depth migration results. The resulting final migration velocity model is shown in Fig. 2.

Based on this velocity model, traveltimes tables for each source and receiver position were calculated using a finite-difference Eikonal equation solver (Podvin & Lecompte 1991) and were used for the subsequent depth migration approach. For the latter we have chosen a grid spacing of 10 m in the profile direction and in depth. With a profile length of 180 km and a maximum target depth of 20 km this leads to a migration grid of $18\,001 \times 2\,001$ gridpoints. We applied a coherency based pre-stack depth migration approach (Hlousek *et al.* 2015) which is based on Kirchhoff pre-stack depth migration including an additional weighting term. This term is calculated from the coherence of a reflection event at neighbouring receivers and it focuses the back-propagated reflection event to its actual reflection/diffraction point in the subsurface. The migration is performed for each shot gather separately and the single migrated shot gathers are stacked phase-consistently to obtain the final seismic depth image.

5 RESULTS

The final depth migrated images are shown in Figs 3, 4 and 5.

5.1 The oceanic plate

West of the trench (Fig. 3 top, Fig. 4a), bending-related faults forming horst-and-graben structures and half-grabens are visible along the whole profile line on top of the oceanic crust. The grabens and half-grabens are filled with variable, but generally little, amounts of sediment. These small sedimentary basins have a lateral extent of about 2 km in the profile direction (E–W), show a maximum thickness of about 200 m, and appear every 2–10 km. We do not observe any seamounts on the oceanic crust along this part of the profile line.

5.2 The subduction interface

The reflection from the plate interface is clearly visible but variable along the interface (Fig. 3 bottom, 4b). Parts with strong reflectivity and parts showing very weak reflectivity alternate with similar section lengths (2–4 km long).

The topography of the subduction interface beneath the forearc part is slightly concave and rough. To some extent, the dip angle of the subducting plate and the large scale shape is controlled by the migration velocities in the hanging wall. Depending on the defined velocities in the hanging wall and its thickness, the depth of the slab at a certain distance from the trench can vary within a range of 1 or 2 km.

5.3 The continental crust

The continental margin shows no accretionary wedge but is rather characterized by a frontal prism and an eroded frontal unit due to subduction erosion (Figs 4b and d) corroborating the results of von Huene & Ranero (2003) further south off Peninsula de Mejillones (23.5°S). This frontal unit has a complex internal structure with mainly eastward dipping reflectors, some of which are traceable from the ocean floor down to the plate interface. The dip of these features increases with increasing distance from the trench. We interpret these reflectors as back rotated splay faults. A change in the slope of the sea floor can be recognized where the major splay faults reach the ocean floor.

At $x = 122$ km, a scarp in the ocean floor can be observed and beneath it a strong eastward dipping reflection can be seen (Fig. 5d). With increasing depth this reflector becomes less visible but it can still be traced for several kilometers (Fig. 4c). It separates a seaward domain with the above described steeply dipping high frequency reflections from a landward domain characterized by more shallowly dipping high amplitude reflections. Since this reflector has a greater downdip extent, stronger reflectivity and straighter appearance than the thrusts in the seaward domain, we interpret this structure as a megasplay fault dividing the seaward from the landward domain.

A topographic high at the sea floor at $x = 125$ km (Fig. 5d) denotes the transition between a frontal unit beneath the lower slope and the continental crust and is very likely an anticlinal structure caused by slip on the megasplay fault that crops out at $x = 122$ km. This high also marks the western limit of a larger sedimentary basin (or two contiguous basins separated by a small anticlinal structure and a smaller splay fault) on the forearc crust that begins at $x = 127$ km

Figure 2. Velocity model for CINCA'95 line 22.

Figure 3. Pre-stack depth migration result of CINCA line 22. Top panel: western part of profile line, oceanic crust west of the trench. Bottom panel: eastern part of the profile line with trench and continental margin. Black rectangles indicate the location of zoomed-in panels in Figs 4 and 5.

(Figs 5a and d). The basin has a thickness of up to about 400 m. At a topographic high in similar position further south (Antofagasta Ridge), drilling during the CINCA cruise (Hinz *et al.* 1998; von Huene & Ranero 2003) recovered volcanic rocks of the Jurassic La Negra Formation.

The underlying continental crust exhibits several strong reflectors between $x = 125$ and 160 km (Fig. 5b) at a depth of 6–12 km are probably related to the gently westward dipping La Negra Formation. Within this formation, there are some inclined reflections with no obvious offset of the subhorizontal stratigraphic reflections. Above these very strong reflectors, another similarly dipping strong reflection boundary can be identified at a depth of approximately 5 km at $x = 140$ km (Fig. 5b) and at approximately 3 km depth at $x = 162$ km (Fig. 5c).

5.4 Sediment bed

Along the forearc slope between $x = 95$ – 120 km, the sediment layer is approximately 100–200 m thick, and west of the topographic high at $x = 122$ it is even 400 m thick. Between $x = 105$ and 122 km, the sediment layer is mainly undisturbed with a reduced thickness of 100–150 m (Figs 4d and 5d). At the scarp on the western edge of the topographic high at $x = 122$ km, the sediment thickness is less than 50 m (Fig. 5d) while east of this high a flat undisturbed basin

on the forearc slope has up to 400 m of sediment with reflection truncations at the eastern end of the ridge indicating possible growth structures (Fig. 5a). Besides the megasplay fault outcropping at the limit between the middle and lower slopes (MLS, at 122 km) and a potentially reactivated normal fault at $x = 123$ km, all the faults east of the megasplay in the middle slope are covered by a continuous sediment layer of about 150 m equal thickness indicating that only the megasplay and the normal fault are accumulating displacement after the onset of deposition of this diagnostic sedimentary layer (Fig. 5d). The eastern boundary of the sedimentary basin is at $x = 130$ km. According to mapped cross cutting relationships (Fig. 5a), normal fault activity has propagated landwards causing a larger but uniform sediment thickness west of a normal fault crosscutting the sediments up to the surface ($x = 130$ km).

6 DISCUSSION

One of the main goals of this study is to investigate the subduction interface and the overlying continental crust of the Chilean subduction zone at seismogenic depths prior to the Iquique megathrust earthquake of 2014. Our results contribute to the understanding of rupture processes in subduction zones with extremely little sediment input, the absence of an accretionary prism and hence rupture processes in a subduction erosional environment (von Huene *et al.*

Figure 4. Pre-stack depth migration result of CINCA'95 line 22. The location of the corresponding zoom boxes of the four images (a), (b), (c) and (d) are marked in the bottom figure (e). (a) sedimentary basins on oceanic crust related to horst-and-graben structures, (b) frontal unit, subduction interface, (c) megasplay fault at MLS and (d) frontal unit with back rotated splay faults.

Figure 5. Pre-stack depth migration result of CINCA'95 line 22. The location of the corresponding zoom boxes of the four images (a), (b), (c) and (d) are marked in the bottom figure (e). (a) sedimentary basin east of topographic basement high on the middle slope, (b) middle slope crust, (c) shallow crust at SB (shelf break), (d) MLS (limit between middle and lower slope), basement high on the middle slope and crust beneath it showing complex fault structures.

1999). We detect several large scale fault structures and fault systems, such as back rotated splay faults, in the frontal unit and a megasplay fault, features often characterizing the near-trench subduction environment. These are consistent with recently published results regarding a separation of the continental wedge into distinct units (Maksymowicz *et al.* 2018). However, some findings are partly in contrast to recently published studies as discussed in more detail below. We further combined our image with the relocated hypocentre locations of fore- and aftershocks of the Iquique earthquake (Sippl *et al.* 2018; Soto *et al.* 2019), and also with refraction interfaces from wide-angle analysis of this profile (Patzwahl *et al.* 1999), as well as with results from density-depth-modelling from Maksymowicz *et al.* (2018). Finally, we further discuss here the significant splay faults that are suspected to play an essential role during rupture processes.

6.1 The splay faults and a megasplay—separation of the continental wedge into distinct units

From bathymetric data, the continental wedge can be divided into three parts: the lower slope, the middle slope and the upper slope (Maksymowicz *et al.* 2018; Geersen *et al.* 2018a). The limit between the lower slope and the middle slope (MLS) coincides with a scarp on the ocean floor at $x = 122$ km followed by a topographic high to the east (Fig. 5d). In contrast to Maksymowicz *et al.* (2018), we identified an eastward dipping instead of a westward dipping megasplay fault at the MLS that separates the continental wedge along the CINCA profile line 22 into two units: a frontal unit beneath the lower slope and the continental crust beneath the middle slope (Figs 4c and e). Additionally we also find a small scale westward dipping reverse fault approximately 500 m west of the eastward dipping megasplay which is well developed in the uppermost 5 km but which cannot be traced further downdip (Fig. 5d). The frontal unit likely contains slope debris as suggested by von Huene & Ranero (2003) and—from the high P-wave velocities at its deep rear—probably also forearc crustal rocks as found further south along the Chilean margin (Contreras-Reyes *et al.* 2014). At the toe of the wedge we image a frontal prism.

In the depth image of Figs 4 and 5 we observe several splay faults beneath the lower slope (Fig. 4d) that have generally rather strong reflectivity and are clearly imaged. Their eastward dip is moderate ($30\text{--}45^\circ$) but with increasing distance from the deformation front the dip increases significantly up to 60° suggesting they are back rotated splay faults forming an imbricate fan. They cut through almost the entire frontal unit beneath the lower slope and in some cases can be followed down to the subduction interface. However, some do and some do not cut through the thin sedimentary layer on the ocean floor. For that reason we interpret these faults to probably being and not being active, respectively. Maksymowicz *et al.* (2018) defined this part as a frontal unit made of debris and highly fractured rock from density modelling that agrees with our seismic imaging results. Contreras-Reyes *et al.* (2014) describes a comparable scenario at the northern-central Chilean margin (31°S) from wide angle seismic results and multichannel seismic data, but with a listric master fault instead of a landward dipping megasplay fault, that separates the hydrofractured continental crust from the igneous and metamorphic rocks belonging to the Coastal Cordillera.

Landward of the megasplay, beneath the middle slope, reflectivity and structural characteristics differ significantly from structures in the frontal unit (Fig. 5b). Here, complex structures of the continental crust with strong, mostly subhorizontal reflectivity can be seen.

Maksymowicz *et al.* (2018) defined a transition zone beneath the middle slope, where the degree of fracturing gradually decreases landwards. These significant differences suggest that the megasplay fault at the MLS acts as a boundary (Fig. 5c). According to our interpretation in Fig. 5(d), the topographic high at $x = 125$ km is very likely an anticline or fault propagation fold (with large amplitude) formed on the hanging wall of a set of eastward dipping thrust faults. The westernmost of these thrust faults corresponds to the above described megasplay, which is the most active structure because it offsets the sedimentary sequence covering most of the other structures east and west. Additionally it has the largest traceable down dip extent and seems to crosscut the complete wedge down to the interface. The megasplay that we identified from the reprocessed seismic section presented in this study can clearly be traced in the uppermost 2 km of the crust where offsets in the sedimentary cover can be used as diagnostic marker. Downwards, the structure can be traced following significant reflections and most likely roots itself into the interface at $x = 137$ km at a depth of 15.5 km. The overall dip of the megasplay is roughly 45° and does not show a listric geometry. Comparable fault limiting slope breaks in the outer wedge were imaged by Strasser *et al.* (2009) at the Nankai trough subduction zone and have been suggested by von Huene *et al.* (2004) and Wang & Hu (2006) to be diagnostic features forming at the updip end of seismic locking at the plate interface.

Beneath the middle slope, we also find some less prominent eastward dipping structures in the continental crust (Figs 5b and 6 top right). Those are either identified indirectly by limiting shallow dipping reflectors within the La Negra Formation or directly by their reflections. The reason for the decreased prominence could either be the acquisition geometry with too small offsets that are not optimal for imaging dipping reflections, or the reflection coefficients for these structures are simply too small in comparison to the steeply dipping faults in the frontal unit. Moreover, these reflecting features are not associated with obvious offsets and may not be faults (see sections below).

A third zone has been defined beneath the upper slope/continental shelf that consists of dense basement. The limit between the middle slope (transition zone) and the upper slope is a seaward-dipping structure at the shelf break (SB, Fig. 5e) that is supposed to have been activated during the largest intraplate foreshock of the Iquique earthquake sequence (Maksymowicz *et al.* 2018; Schurr *et al.* 2020). In this part of our processed line the water depth is lower than 2 km and the seafloor multiple is imaged at depths of around 5 km. Here, the multiple is well suppressed, but reflections beneath the multiple are very weak. In Fig. 5(c) we were able to identify only a 1 km short and weak seaward dipping fault beneath the ocean bottom at $x = 168$ km that coincides with the shelf break and shows a seaward increase in slope sediment thickness.

6.2 Varying interplate reflectivity, subducted seamounts and subduction erosion

Along the Iquique ridge on the oceanic Nazca Plate, several large seamounts are known. Although it is unclear whether the Iquique ridge has already reached the trench, Geersen *et al.* (2015) interpreted a large subducted seamount on a time-migrated profile of CINCA line 22 beneath the topographic high at $x = 125$ km. The latter is described as being an uplift structure due to the subducted seamount. They linked the seamount to the M_w 8.1 Iquique earthquake rupture zone, together with interpretations of two additional CINCA profiles that cross the rupture area further south. Geersen

Figure 6. Depth section of CINCA'95 line 22 with refraction interfaces (black lines) from Patzwahl *et al.* (1999) derived from wide angle tomography using land seismic stations of the CINCA'95 experiment. Relocated fore- and aftershocks of the Iquique 2014 event (Sippl *et al.* 2018) are plotted with circles and stars. Location uncertainty is approximately 5 km. Black arrows mark steeply dipping structures. One of these is likely associated with the parallel and nearby refraction interface.

et al. (2018b) interpret shallow ocean floor relief as potential scars in the continental forearc and deduce that the Iquique ridge extends beneath the Northern Chilean forearc.

In our depth migration result of line 22 we do not observe a related topographic high along the plate interface at this described location. The structure interpreted from the time domain image of Geersen *et al.* (2015) is very likely a velocity pull-up artefact: an effect of large velocity contrasts between water and crust, which results in the deeper reflectors, such as the subduction interface, following the shape of the ocean bottom topography. After pre-stack depth migration this effect is eliminated. This specific reflection between $x = 120$ and 122 km is now slightly up-bent at its western edge, possibly due to a migration artefact and the superposition with back rotated splay faults in the hanging wall. Further, the topographic high at $x = 125$ km on the ocean bottom may be an uplift structure linked to folding in the hanging wall of an active megasplay fault that we observe as discussed above in Section 6.1.

The reflectivity pattern of subduction interfaces has been the subject of various recent investigations, for example Sage *et al.* (2006) at the central Ecuador convergent margin or Li *et al.* (2015a) in the Alaska subduction zone. Even though these two subduction zones are sediment-rich and comprise large accretionary prisms, in contrast to the Northern Chilean subduction zone, there are some similarities that can be compared with our results, such as the varying reflectivity with increasing depth (Li *et al.* 2015b) and an 'interplate patchiness' (Sage *et al.* 2006).

In Northern Chile, at CINCA line 22 the reflectivity characteristics of the slab interface varies notably along the profile line

(Fig. 4b). There are sections with strong reflections and sections with none. The spacing of these changes of 5–10 km is comparable to the wavelength of the horst-and-graben structures on the incoming oceanic plate, which are filled with sediment with a maximum thickness of about 200 m (Fig. 4a). This sediment is expected to have a higher water content than the fractured oceanic crust, which may lead to overpressured fluids at the subduction interface with enhanced impedance contrasts, once these sediment lenses are subducted. A comparable scenario has been described by Sage *et al.* (2006) at the central Ecuador convergent margin. It is important to point out that in North Chile sedimentation rates are rather slow in contrast to central Ecuador.

We challenge the hypothesis of Geersen *et al.* (2015) and provide an alternative explanation for the varying reflectivity along the slab interface. Due to the similar spacing and length-scale of the graben structures along the oceanic plate, we associate strong reflections within the subduction channel with fluid-rich sediment lenses originating from buried graben sediments. The reflection element interpreted by Geersen *et al.* (2015) as a subducted seamount could be such a fluid-rich sediment lens. In contrast to the same length spacing for the oceanic grabens west of the trench and the associated reflection bands along the plate interface, the Iquique ridge volcanic edifices typically exhibit much larger values with some irregular 15–50 km spacing, 4–15 km width of individual seamounts at their base and elevations ranging from 0.3 to 2.5 km superposed on the broad swell of some 1 km elevation above background. These values are at odds with the regular spacing of the change in character of the reflectors. While we cannot exclude the

role of some local rough seafloor topography from the Iquique ridge to be present at depth, from the above geometric correlation we suggest that it is overridden by extensional processes at the outer ridge and that the weak reflections from the interface are associated with less-overpressured segments due to horst structures. These changes along the subduction interface may also directly imply an albeit geodetically unresolvable change in lower and higher interplate coupling, because variations in the subducting seafloor roughness have been shown to have an influence on the mechanisms of plate coupling and consequently on the seismic rupture processes (Bilek *et al.* 2003). Scherwath *et al.* (2009) and Contreras-Reyes & Carrizo (2011) pointed out that a thick subduction channel favours the earthquake rupture propagation even if the top of the subducting oceanic crust is rough. In contrast, a thin subduction channel above a heterogeneous oceanic basement limit the size (seismic barrier behaviour) of thrust earthquakes (Contreras-Reyes & Carrizo 2011; Wang & Bilek 2011).

6.3 Correlation of depth-migrated images with seismicity, 2-D velocity–depth and coseismic slip models

Fig. 6 shows the seismic depth image together with the relocated hypocentres from fore- and aftershocks of the M 8.1 2014 Iquique earthquake (Sippl *et al.* 2018; Soto *et al.* 2019; Schurr *et al.* 2020). The seismic image and the hypocentre locations complement each other since many of the hypocentres are located beneath 20 km depth where our seismic image terminates. The location uncertainty of the hypocentres is approximately 5 km in both, depth and horizontally. Considering the fact that all seismic stations used for location of the hypocentres are on land, this implies that the hypocentres with large distances from these stations consequently have a large location error especially in depth, particularly for hypocentres located at $x < 150$ km. Furthermore, the velocity model used for location of the hypocentres and the velocity model used for pre-stack depth migration are not identical, so that a direct comparison is not fully justified. The latter can be achieved and is subject of further investigations.

Comparing the depth image with the hypocentre locations (Fig. 6), we observe that west of the megasplay fault the interface beneath the lower slope is aseismic. Most, but not all of the fore- and aftershocks of the Iquique earthquake are concentrated within the overriding plate and the subduction interface east of this splay fault, beneath the middle slope and the shelf. We should keep in mind that the location error of the hypocentres in the X direction and depth increases to about 7.5 km (horizontally) and 12 km (vertically) in this region of interest close to the trench (Sippl *et al.* 2018).

In Fig. 6 we also added wide-angle refraction interfaces from Patzwahl *et al.* (1999). Patzwahl *et al.* (1999), analysed wide-angle seismic data from a mobile onshore array that was recording the airgun pulses from the offshore seismic profile lines 7, 9, 13 and 22 during the first phase of the CINCA'95 experiment. At line 22, two models were derived from two neighbouring land stations. Only one of those is published, the other one that we compare our imaging results with, is not. In general there is a good agreement of the unpublished wide-angle boundaries with our depth image. In particular, the eastward dipping boundary subparallel to the subduction interface correlates well with a short segment of an eastward dipping reflection in the migrated section at $x = 152$ km and $z = 9$ km, that is highlighted by a pair of arrows in the upper right part of Fig. 6. The origin of the refraction boundary is unclear.

This boundary was interpreted by Patzwahl *et al.* (1999) as a fossil extensional structure or to be caused by tectonic underplating of oceanic crustal material beneath the forearc which we here consider unlikely as there is no apparent offset of reflections. Fig. 6, however, shows this boundary—that cuts obliquely across the stratigraphic architecture—to mark the upper bound of concentrated fore- and aftershock seismicity of the Iquique earthquake. Clearly, the density of seismicity, and hence fracturing, is significantly stronger below this inclined boundary with little seismicity above it. Since the MCS and wide angle data acquisition predate the entire Iquique earthquake series, it appears likely that this part of the forearc basement has undergone multiple rupturing and healing stages of similar geometric extent that may add up to the reflection and wide angle image observed today.

In Fig. 7 we illustrate the results with a structural interpretation of our depth image superposed with locations of the fore- and aftershocks (Sippl *et al.* 2018) as well as the focal mechanisms of the main shock and the two largest foreshocks that are located within a lateral distance of 40 km from the profile. Further, we illustrate the distinct units of the slope and the locations of MLS and SB as published by Maksymowicz *et al.* (2018). The SB is described by Maksymowicz *et al.* (2018) as a westward dipping normal structure that separates a transition segment of the continental slope beneath the middle slope from the shelf. They interpret this to be a fault that was activated during the large M_w 6.7 intraplate foreshock. At the MLS location, we introduce an eastward dipping megasplay fault instead of a westward dipping structure. It separates the aseismic from the seismically active zone as discussed in Section 6.1.

Fig. 8 shows two different slip distributions of the 2014 Iquique earthquake (Schurr *et al.* 2014; Duputel *et al.* 2015) represented by their 3 m slip contours, and the location of CINCA line 22 together with the hypocentres of the main shock (big red star), the largest foreshock (smaller yellow star) and all fore- (grey) and aftershocks (black). Close to line 22, the rupture area of the Iquique earthquake appears to be limited updip where it meets the megasplay fault. The updip limitation of fore- and aftershocks is more diffuse. Obviously, the stress state of this updip domain before and after the main event was sufficient to cause minor brittle rupture in the deep landward parts of the frontal imbricate unit, with the updip parts of the interface remaining aseismic. This is also suggested by several studies inferring seismic locking from GNSS data (Métois *et al.* 2013; Schurr *et al.* 2014; Li *et al.* 2015a). However, we note that GNSS data resolution near the trench is limited and that a stress shadow may impact the kinematic and mechanical status of this part of the interface system (e.g. Almeida *et al.* 2018; Schurr *et al.* 2020, and references therein). Only some aftershock seismicity organized mainly in streaks is seen in this near-trench domain by Soto *et al.* (2019). While an assessment of the exact kinematic status of this part of the interface is beyond the scope of this paper, we note that seismicity indicates that the disrupted interface reflectivity style observed for this part of the plate boundary may be related to a propensity for predominantly aseismic slip—much in line with suggestions by Wang & Bilek (2011) that seafloor roughness and relief reduces locking. Moreover, in Southern Chile, Moreno *et al.* (2014) described seismological evidence that pore fluid overpressure mechanically controls the difference in kinematic response of the plate interface. Therefore, in addition to sea floor roughness, small scale (<10 km) heterogeneity in the hydraulic state of the interface zone may show up as a mechanism in affecting seismic versus aseismic slip in contrast to an interface domain with a more homogeneous hydraulic state.

Figure 7. Final structural interpretation for CINCA line 22 including a frontal unit beneath the lower slope, with back rotated splay faults (black steep lines), a megasplay fault at the MLS (blue line), the continental crust beneath the middle slope and subducting oceanic crust (red). Locations of hypocentres are from Sippl *et al.* (2018). Red dashed line is the boundary between the transition zone and the dense basement at the shelf brake (SB) from density modelling by Maksymowicz *et al.* (2018). Interpreted structures from Figs 4, 5 and 6 (top right) are added. Coseismic slip as derived by Schurr *et al.* (2014) and Duputel *et al.* (2015) is marked underneath, with minimum 3 m for both (see also Fig. 8).

Figure 8. Map view showing two slip distributions of the 1 April 2014 Iquique main shock with contour lines at 4 and 3 m slip (orange: Schurr *et al.* 2014 and black: Duputel *et al.* 2015) and fore- (grey) and aftershocks (black) (Sippl *et al.* 2018). MLS and SB surface traces from Maksymowicz *et al.* (2018). The approximate trace of the MLS on the megathrust at line 22 is also marked. Yellow hatched area is the aseismic lower slope around line 22. Big red star: main shock, smaller yellow star: largest foreshock in the upper plate.

6.4 Sedimentary layer along the continental slope

At the scarp on the western edge of the topographic high at $x = 122$ km, sediment thickness is less than 50 m while east of this high a flat undisturbed basin on the forearc slope has up to 400 m of sediment. Hence, we conclude that fault activity of the east-dipping reverse fault at this location (see above) occurred at the same time with sedimentation leading to a vertical displacement of no more than a few hundred meters. The formation of the topographic high in the hanging wall is kinematically linked to the shortening associated with this reverse faulting. From the approximate sediment

age (Neogene) of the growth strata in the hanging wall, slip rates on this fault can be as low as $<0.1 \text{ mm yr}^{-1}$, a value typical of the Northern Chilean forearc faults (Victor *et al.* 2018). Moreover, the position of this fault is not only correlated with the updip end of coseismic slip and aftershocks of the Iquique earthquake, but also with the boundary between continental framework basement and a narrow prism with imbricate architecture similar to the finding of von Huene & Ranero (2003) further south. The presumed seismogenically locked zone from these observations therefore coincides with a sedimentary basin on top of this zone, a feature also found in other convergent margins (e.g. Fuller *et al.* 2006) including the accretionary southern-central Chilean margin (Moscato *et al.* 2011). While the seismogenic behaviour influences deformation in the upper plate, it appears in turn to be chiefly controlled by both the mechanical properties at the interface, and the overlying hanging wall rock types.

7 SUMMARY AND CONCLUSIONS

We reprocessed the northernmost line of the CINCA'95 multichannel reflection seismic data set from offshore Northern Chile using a focusing pre-stack depth migration method. This line is approximately perpendicular to the trench and crosses the oceanic Nazca Plate, the trench and the continental crust, where the M_w 8.1 Iquique megathrust earthquake occurred on 1 April 2014. Analysing the imaged structures and comparing our seismic depth image with the fore- and aftershocks of that earthquake (Sippl *et al.* 2018), refraction boundaries from Patzwahl *et al.* (1999), and results from density-depth-modelling from Maksymowicz *et al.* (2018) we observed the following key features:

(1) From our depth image and the distribution of fore- and aftershocks from Sippl *et al.* (2018) we support a zoning of the Northern Chile active margin as proposed by Maksymowicz *et al.* (2018): A frontal unit with highly faulted rocks that acts as an aseismic barrier for rupture propagation beneath the lower slope, and a zone beneath the middle slope that appears nearly unfaulted. In this latter zone numerous fore- and aftershocks of the Iquique 2014 earthquake are observed within the deep continental crust and on the slab interface.

(2) We found that both zones are separated by an eastward dipping megasplay fault that reaches the seafloor at the MLS. This sharp separation of an aseismic and a seismically active zone sheds new light on the control of seismic locking/slip versus aseismic creep through possibly both, the rough seafloor and hydrologically heterogeneous nature of the plate interface as well as from the different nature of the hanging wall (continental framework rocks versus narrow accretionary wedge).

(3) We identify several eastward dipping, back-rotated splay faults in the frontal unit, which have largely not been (re)activated after the deposition of a sedimentary cover layer. Only the two westernmost thrust faults show potential offset of the sedimentary cover with <50 m offset. The eastward dipping active megasplay fault is marked by a topographic high in the hanging wall as well as by a sedimentary basin east of this high. East of this active megasplay adjoins the transition zone beneath the middle slope. Here, the wealth of aftershock seismicity in the upper plate occurs in the plate interface and the deeper crust without, however, breaking the surface or the base of the sediment layer. The dip angle of this upper limit of seismicity is comparable to the dip of the megasplay and correlates with a wide angle refraction boundary from Patzwahl *et al.* (1999).

(4) Instead of finding support for a subducted seamount as a potential control on the Iquique earthquake, we provide an alternative interpretation of subducting horst and grabens with their sediment basins that cause variable reflectivity along the slab interface due to variable amounts of overpressured fluids within the subduction channel promoting aseismic slip.

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SUPPORTING INFORMATION

Supplementary data are available at [GJI](https://doi.org/10.1002/gji.10000) online.

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