Active tectonics of the North Chilean marine forearc and adjacent oceanic Nazca Plate

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Key Points:

- Multibeam bathymetric and seismic reflection data image the structure of the North Chilean marine forearc and the oceanic Nazca plate
- The structural character and tectonic configuration of the offshore forearc and the oceanic plate change significantly along the margin
- The derived pattern of permanent deformation may hold information for studying seismicity or other types of short term deformation

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Abstract

New multibeam bathymetry allows an unprecedented view of the tectonic regime and its along-strike heterogeneity of the North Chilean marine forearc and the oceanic Nazca Plate between 19-22.75°S. Combining bathymetric and backscatter information from the multibeam data with sub-bottom profiler and published and previously unpublished legacy seismic reflection lines, we derive a tectonic map. The new map reveals a middle and upper-slope configuration dominated by pervasive extensional faulting, with some faults outlining a >500 km long ridge that may represent the remnants of a Jurassic or pre-Jurassic magmatic arc. Lower slope deformation is more variable and includes slope-failures, normal faulting, re-entrant embayments, and NW-SE trending anticlines and synclines. This complex pattern likely results from the combination of subducting lower-plate topography, gravitational forearc collapse, and the accumulation of permanent deformation over multiple earthquake cycles. We find little evidence for widespread fluid seepage despite a highly faulted upper-plate. An explanation could be a lack of fluid sources due to the sediment starved nature of the trench and most of the upper-plate in vicinity of the hyper-arid Atacama Desert. Changes in forearc architecture partly correlate to structural variations of the oceanic Nazca Plate, which is dominated by the spreading-related abyssal hill fabric and is regionally overprinted by the Iquique Ridge. The ridge collides with the forearc around 20-21°S. South of the ridge-forearc intersection, bending-related horst-and-grabens result in vertical seafloor offsets of hundreds of meters. To the north, plate-bending is accommodated by reactivation of the palaeo-spreading fabric and new horst-and-grabens do not develop.

1 Introduction

Although marine forearcs are among the geologically most active regions on Earth, they are less accessible and therefore generally less studied than their terrestrial counterparts. Located directly above the boundary between overriding and subducting plate, they are continuously deformed and shaped by a variety of tectonic processes. Ongoing deformation over millions of years ultimately results in distinct forearc morphologies and deformation patterns. Documentation of these patterns can contribute to understanding both the long-term geologic and tectonic evolution of an active margin and the characteristics of permanent deformation that are related to earthquake activity. Furthermore, marine forearcs are inferred to represent regions of important exchange of fluids and volatiles between the lithosphere and the hydrosphere, but the exact parameters that govern this exchange remain poorly understood.

Here we study the little-explored marine section of the convergent margin of Northern Chile. The margin represents an endmember erosional system as characterized by trench sediment thickness. Located adjacent to the hyper-arid Atacama Desert, there is little terrestrial sediment influx, and large parts of the oceanic plate and the marine forearc are starved of sediment (Coulbourn and Moberly, 1977; Moberly et al., 1982; von Huene et al., 1999; Sallares and Ranero, 2005; Ranero et al., 2006; Geersen et al., 2015, 2018). Consequently, the tectonic signature of both the oceanic Nazca Plate and the forearc slope of the overriding South American plate is well preserved by the seafloor morphology and not overprinted by sedimentary basins. The pristine preservation of tectonic structures, which has also been noted onshore (Allmendinger and González, 2010; Armiijo et al., 2015 and references therein), offers an unprecedented opportunity to study deformation accumulated over long (millions of years) timescales.
Subduction erosion has been the long-term dominant tectonic mode in Northern Chile since at least Mesozoic times (Rutland, 1971; von Huene and Scholl, 1991). This is manifested in pervasive extensional faulting in the forearc (Armijo and Tiele, 1990; von Huene et al., 1999; Allmendinger and González, 2010). Subduction erosion has further resulted in ~250 km of margin loss since 150 Ma (e.g., Scheuber and Reutter, 1992) and eastward migration of the volcanic arc. While the Jurassic arc crops out along the coast, the presently active arc is located about 200 km farther to the east (e.g. Rutland, 1971). However, tectonic erosion may have slowed down over the last 20 Ma (Clift and Hartley, 2007). For the last 2 Ma there is evidence for uplift of the coastal regions in southern Peru and northern Chile (Charrier et al., 2007; Clift and Hartley, 2007). The causes, exact timing, and spatial extent of the Quaternary uplift of the terrestrial forearc and its possible relation to underplating, however, are controversial (Armijo and Tiele, 1990; Clift and Hartley, 2007; Encinas and Finger, 2007; Armijo et al., 2015).

Today the large-scale tectonic framework of the margin is controlled by the subduction of the Eocene oceanic Nazca Plate underneath the South American Plate, which occurs at an azimuth of about 80° and a rate of about 7 cm/yr (Angermann et al. 1999; Müller et al., 2008). Recent earthquakes along the margin include the 1995 Antofagasta (Mw 8.0), 2001 Arequipa (Mw 8.4), 2007 Tocopilla (Mw 7.7), and 2014 Iquique (Mw 8.2) events (Fig. 1). A significant part of the margin located to the north and south of the 2014 Iquique earthquake, however, did not rupture during these events and is thought to have last ruptured during a pair of great earthquakes about 141 years ago (M8.5-9.0 Arica earthquake of 1869 and ~Mw 8.6 Iquique earthquake in 1877) (Lomnitz, 2004). These regions are referred to as the Southern Peru and Northern Chile seismic gaps, respectively.

In this work we combine new high-quality multibeam bathymetric data from two recent research cruises of the R/V SONNE (in 2015) and R/V Marcus G. Langseth (in 2016) with older legacy data, several seismic reflection profiles imaging the marine forearc and the oceanic Nazca Plate, and 4 kHz sub-bottom profiler lines. We use the data to map tectonic structures and analyze the seafloor morphology in order to characterize tectonic deformation in the regions of the 2014 earthquake and the seismic gap to the south. The structural and tectonic interpretations are summarized in tectonic maps, which are the foundation for a discussion of forearc deformation over long timescales. In conjunction with seafloor backscatter information derived from the new multibeam data, we also discuss indications for seafloor seepage for the first time in Northern Chile. Knowledge of long-term forearc deformation and inferences on the hydro-geological system may hold valuable information for ongoing and future studies looking at seismicity and other types of short-term deformation.

2 Materials and Methods

2.1 Multibeam bathymetry

Most of the multibeam bathymetric data were acquired during the 2015 R/V SONNE cruise SO244 (Behrmann et al., 2016; Kopp et al., 2016) and the 2016 R/V Marcus G. Langseth cruise MGL1610 (Tréhu et al., 2017). We complement the new high-resolution multibeam bathymetric data with older data collected over the last two decades during multiple seagoing campaigns with German research vessels (R/V SONNE, R/V METEOR) and masked data from the Global Multi-Resolution Topography (GMRT) synthesis (Ryan et al., 2009).
During R/V SONNE cruise SO244 a Kongsberg Maritime EM 122 multibeam echosounder with a 1° by 0.5° configuration was used to collect bathymetric and backscatter data at survey speeds between 8 kts and 10 kts. The beam angle was slightly reduced for higher sounding density and varied between 120° and 140°. The echosounder operated in dual-ping mode and the beam was slightly tilted forward by 2-5° in order to avoid signal penetration at nadir in soft sediments. During MGL1610, bathymetric data were acquired with an EM 122 echosounder at 4.5-5 kts while the ship was also acquiring multichannel seismic reflection and large aperture data for a 3D tomographic imaging survey of the source region of the 2014 earthquake (Tréhu et al., 2017).

Bathymetric data processing mainly consisted of eliminating outliers through a triangulation filter with three iterations as well as manual cleaning of erroneous soundings. The data were subsequently gridded using a near-neighbor algorithm that takes into account four neighboring cells, eliminates cells with less than two soundings per cell, and interpolates for two rows/columns. The recorded data were gridded into 75 m cells. For processing of the backscatter information, all data recorded during turns of the ship were eliminated before gridding to a 20 m cell size throughout the study area.

2.2 Seismic reflection

The five seismic reflection lines used in this study were collected in 1995 during R/V SONNE cruise SO104, led by the German Bundesanstalt fuer Geowissenschaften und Rohstoffe (BGR). The cruise SO104 was conducted under the framework of the crustal investigations off- and onshore Nazca/Central Andes (CINCA) project. Seismic signals were generated with a ~3,124-cubic-inch (51.2 l) well-tuned airgun array which was shot every ~50 m. The reflected signals were recorded using a 3 km long digital streamer with 25 m group spacing. See Geersen et al. (2015, 2018) for extended sections of the seismic lines.

Re-processing with the Claritas software package was conducted at the Barcelona Center for Subsurface Imaging (Barcelona-CSI). The seismic pulse reverberations were eliminated with a pre-stack statistical deconvolution with two overlapping windows for filter design. Subsequently, shot point interpolation to 25 m and offset regularization was conducted for two-dimensional filtering. Stacking velocity was picked every 5 km. The water-layer multiple was attenuated with parabolic radon filtering on super gathers after normal-moveout correction. A pre-stack time migration was done on receiver gathers with time and space varying velocities based on the stacking velocity model. After pre-stack time migration, new velocity analyses were picked for common-mid-point stacking. After stacking, data were post-stack time migrated using a finite difference algorithm, with velocity layers mimicking the large-scale geological structure of the margin. For the shallow geology, seismic velocities were based on velocity analysis, whereas deeper in the section, seismic velocities were based on wide-angle seismic profiles collected across the North Chilean margin (Contreras-Reyes et al., 2012). After post-stack time migration, data were bandpass frequency filtered with time- and space-varying filters, based on the geological structure. For display purposes, the data are presented with an automatic gain control equalization based on the mean amplitude within a sliding window of 1 s.
3 Structural and tectonic observations

We first describe the structural setting of the oceanic Nazca Plate prior to subduction (Fig. 2) before concentrating on the marine forearc. With respect to the latter we distinguish between a northern (19°S – 21°S; Fig. 3) and a southern (21°S – 22.75°S; Fig. 4) study area. This distinction is convenient for organizing the figures and discussing the data and does not reflect a sharp along-strike structural change, although gradual and subtle along-strike changes are observed. For all three regions we show the combined bathymetric data (Figs. 2a, 3a, 4a) in conjunction with the derived tectonic maps (Figs. 2b, 3b, 4b). Figure 5 shows the backscatter information recorded during R/V SONNE cruise SO244 in combination with selected sub-bottom profiler lines, and Figure 6 shows perspective views of the seafloor in selected regions.

3.1 The oceanic Nazca Plate

The most prominent morphologic feature on the oceanic Nazca Plate offshore Northern Chile is the Iquique Ridge (Fig. 1a; also compare Bello-Gonzáles et al., 2018). Between ~19.5-20.8°S the ridge is characterized by a seafloor depth that is ~1 km shallower compared to the oceanic plate farther to the south (Fig. 2). Within this area, multiple individual ridges are elevated up to 1000 m above the surrounding seafloor. Farther to the south, between ~20.8°S-22°S, the character of the Iquique ridge changes. Here, it is characterized by multiple individual seamounts that sit on oceanic crust in ~4000 m of water. The presence of the seamounts may result from a time of reduced magmatic activity and crustal formation at the Foundation hotspot between 48-52 Ma. The seamounts have diameters of 1000-5000 m and are elevated 1000–1500 m above the surrounding seafloor. Some of the seamounts show well preserved conical calderas a few hundred meters deep. The seamounts are surrounded by magmatic edifices. The backscatter data (Fig. 5a) images the steep flanks of the seamounts and magmatic edifices while their flat tops are correlated with low backscatter intensity. Where the Iquique Ridge enters the trench, the trench is about 1 km shallower than farther to the north and south. This shallower relief indicates that the northeastern end of the Iquique Ridge extends beneath the North Chilean forearc and is currently subducting under the South American Plate.

The oceanic plate off Northern Chile is characterized by abyssal hills that strike NW-SE, observed in bathymetric (Figs. 2 and 6a) and backscatter (Fig. 5a) data. With a strike direction parallel to the paleo-spreading center, these features are commonly interpreted as a seafloor spreading fabric, resulting from the interaction of magmatism and faulting during the generation of new oceanic lithosphere at mid-ocean ridges (Carbotte and Macdonald, 1994). Where not overprinted by the Iquique Ridge, the spacing, length, topography and strike direction of the spreading fabric on the oceanic plate seaward of the trench does not vary significantly throughout the study area. The spacing between individual abyssal hills is usually <5000 m and the vertical offset, from the crests of the hills to the surrounding seafloor, ranges from 50 to 200 m.

South of ~20°S, multiple ~N-S striking horst-and-grabens characterize the oceanic plate up to 50 km seaward of the deformation front (Figs. 2b and 6a). The horst-and-grabens induce vertical seafloor offsets of up to 800 m. Seismic line SO104-26 images a prominent horst-and-graben structure around the deformation front (Fig. 7b). Aligned parallel to the trench axis and the coastline, the horst-and-graben structures result from bending-induced normal
faulting of the subducting oceanic plate (Savage, 1969; Bodine and Watts, 1979). The surface outcrops of the faults that bound the horsts and grabens show up as high amplitude features in the backscatter map due to their steep slopes (Fig. 5a). North of 20°S, where the trench and the South American coast swing to strike northwest, the structural pattern of the oceanic plate changes. Here, horsts-and-graben are no longer observed. Instead, closely spaced (< 5000 m) trench parallel half-grabens, which seem to correspond to the above described spreading fabric, are observed all the way to the trench axis. The latter is located in a water depth of about 6500 m, approximately 1500 m shallower than the trench axis south of 20°S.

Seismic line SO104-26 images 100-200 m of sediment within the above-mentioned graben close to the deformation front (Fig. 7b). However, in all other seismic sections a sedimentary trench fill is absent (at the scale of seismic resolution of ~50 m) (Fig. 7). Due to the limited sediment cover, the structural pattern of the oceanic plate can be traced nicely to the deformation front (Fig. 2). This is often not the case, including in other erosive margins where a thin hemipelagic sediment cover often obscures the structural fabric of the subducting plate (e.g. Costa Rica, Shipley et al., 1992; Japan, von Huene and Lallemand, 1990).

3.2 The marine forearc

3.2.1 Northern study area

Throughout the northern study area, prominent seafloor scarps, indicative of upper plate faulting, are mostly absent on the lower continental slope (> 5000m water depth; bluish to greenish colors in Fig. 3a). South of 20°S the seafloor is characterized by broad synclines and anticlines striking predominantly in NW-SE direction. Seismic line SO104-26 crosses one anticline, which develops above subducting lower-plate topography (Fig. 7b). The other two seismic lines from the northern study area image a small frontal prism of 2 km (SO104-22, Fig. 7a) to 8 km (SO104-27, Fig. 7c) width with little internal structure. Farther landward, the latter seismic line images a reflection in the upper plate ~1 s TWT beneath the seafloor that possibly indicates the seaward-most extension of the continental basement.

Towards the northernmost extent of our bathymetric dataset (19.2°S - 19.9°S) the lower slope is affected by multiple seaward dipping scarps, forming multiple amphitheater-like lower slope embayments that open to the trench (Figs. 3 and 6b). The embayments have diameters of 10–30 km. Up slope of all embayments, a combination of landward and seaward dipping seafloor scarps is observed (Figs. 3 and 6b). Some of the scarps host steep (up to 30°) slopes. At multiple locations throughout the northern study area, the deformation front, which marks the transition from the trench to the marine forearc, is affected by small slope failures, each of them leaving a characteristic spoon-shaped embayment within the lowermost slope.

The middle continental slope (2000-5000 m water depth, yellowish colors in Fig. 3a) is dominated by multiple seaward dipping scarps that result from normal faulting. The normal faults run in groups with a strike direction parallel to the deformation front. For most scarps, vertical offsets are ~10-20 m. They do not, however, show up as distinct features in the backscatter data (Fig. 5a), which may be due to their short along-strike extent (often < 5 km), in combination with the small vertical seafloor offset. At multiple locations, the middle slope forms a long-wavelength (~50 km) bulge (e.g. ~19.3°S and 20°S) as expressed by depth contour lines kinking towards the deformation front.

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The structural complexity of the upper slope (above 2000 m water depth, orange to reddish colors in Fig. 3a) increases southwards. North of 19.5°S, fault scarps at the seafloor rarely occur in water depths above 1000 m. However, multiple NW-SE striking escarpments indicate the presence of seaward dipping normal faults between 1000–2000 m water depths. Farther south, normal fault scarps extend to water depths above 1000 m, and south of 20.5°S, where the spatial coverage of the bathymetric data extends to water depths around 500 m, the upper slope hosts a rather blocky and rough morphology.

The intensity of the backscatter signal generally increases in the region of the upper slope but distinct patches of high backscatter are only visible at some locations (e.g. 20.7°S and 20.9°S; Fig. 5a). The 4 kHz sub-bottom profiler data collected across these patches (Fig. 5b) indicate that the high-backscatter corresponds to 1-2 km wide plateaus not covered by sediments. In-between these topographic highs, small basins with some tens of meters thick-layered sediments show reduced backscatter intensity. The sediment starved topographic highs host multiple pinnacles up to 10 m high. Considering the steep dip angles of some of the underlying strata (Fig. 5b), the pinnacles may represent comparatively harder layers cropping out at the seafloor.

Between 1000-2000 m water depth a N-S striking ridge is visible at multiple locations (Fig. 3a). Seismic lines SO104-22 (Fig. 8a) and SO104-27 (Fig. 8b) image the ridge to be bound by large (hundreds of meters vertical displacement) normal faults which dip seaward to the west of the ridge and seaward and landward to the east of the ridge. Towards the north, the impact of the ridge on seafloor morphology is reduced due to an up to 0.8 s TWT thick section of slope sediment (Fig. 8a). The ridge can, however, be well identified in seismic reflection data (Fig. 8). Around 20.6°S the ridge is most prominent in the multibeam bathymetric data and rises 500-1000 m above the surrounding seafloor. Here the seafloor to the seaward side of the ridge appears rather smooth over a distance of about 10 km across the slope and 40 km along the slope. The smooth seafloor is caused by the accumulation of slope sediment (Fig. 8b). However, normal faulting continues under the slope sediments as evidenced by up to 100 ms TWT large vertical offsets in the top of the continental basement (Fig. 8b).

Submarine canyons dissect the northern study area at multiple locations. The backscatter map (Fig. 5a) outlines the course of the canyons due to their steep flanks and possibly coarse beds, which both usually generate high backscatter intensities. Around 19.2°S a prominent canyon, hereafter referred to as Camarones Canyon, can be traced from the upper slope to water depths of ~5000 m (Figs. 3 and 6b). Camarones Canyon has a width of 500-1000 m and cuts up to 150 m into the surrounding seafloor. In the area of the middle slope, the canyon kinks to the north around one of the mid-slope bulges mentioned above. Within the area of the bulge, multiple paleo-canyons are visible to the south of Camarones Canyon. The paleo-canyons seem to be disconnected from the active thalweg of Camarones Canyon. The lower paleo canyon hosts multiple meanders and cuts deep (up to 250 m) into the surrounding seafloor. The thalweg of Camarones Canyon dips gently over a wide area of the upper and middle continental slope to a water depth of about 3000 m (see inset with canyon profiles in Fig. 3). Below 3000 m water depth the dip of the thalweg increases and the canyon forms a straight line perpendicular to the deformation front. Around 19.9°S another canyon, hereafter referred to as Pisagua Canyon, can be traced from the upper slope almost to the deformation front. Incision depth varies between 50–300 m. The thalweg profile of the Pisagua Canyon dips gently along the upper continental slope to a water depth of about 2000 m, where the dip angle increases and the canyon forms a straight line perpendicular to the deformation front.
Farther to the south between 20°S - 20.5°S, multiple smaller canyons are visible at the upper and the middle slope. These canyons terminate downslope in the area of the active normal fault scarps at the middle slope.

3.2.2 Southern study area

Within the southern study area (Fig. 4), multiple slope failures are mapped at the deformation front between 21°S–22°S. Here, the lower continental slope is dissected by multiple seaward dipping normal fault scars, which are not present in similar water depth in the northern study area. Seismic line SO104-07 shows that in the area of the normal faults at the lower slope, a frontal prism as imaged elsewhere is absent (Fig. 7). Instead, multiple seafloor scarps above a thin upper-plate support the interpretation of normal faulting at the lower slope (Fig. 7d). Seismic line SO104-09, located just south of 22°S, documents a 6 km wide frontal prism (Fig. 7c). South of ~22.5°S the resolution of the bathymetric data in the area of the lower slope decreases (area not covered by new multibeam data) making it difficult to judge whether the absence of distinct fault scarps is real or is simply unresolved by the older, lower resolution bathymetric data in this area.

The middle slope is dominated by N-S striking fault scars that represent seaward-dipping normal faults and that vertically offset the seafloor <50 m, similar to the northern study area. Between 21.7°S and 22.4°S the middle slope hosts a 50 km wide embayment that opens to the lower slope. The water depth within the embayment is ~1000 m deeper compared to the adjacent mid-slope areas to the north and south. The embayment is rimmed up-dip and along-strike by steep scarps with slopes locally exceeding 30°. The seafloor within the embayment shows a reduced morphological complexity with only a few small N-S striking fault scars. Seismic line SO104-09 shows that the smooth seafloor results from slope sediments, which mask the basement topography (Fig. 8c). A second similar embayment, although smaller in diameter, is visible around 22.6°S.

Upslope of the embayments, the N-S trending ridge discussed in section 3.2.1 is visible in multibeam (Fig. 4) and seismic reflection (Fig. 8c) data. Although the landward flank of the ridge is partially buried under thick slope sediment (Fig. 8c) it stands out as a prominent morphologic feature at the seafloor throughout the entire study area. Similar to the northern study area, the ridge is bound by huge landward and seaward dipping normal faults (Fig. 8c). The ridge likely represents a continuous structural feature over a latitudinal distance of at least 500 km, extending from around 19.5°S (Fig. 8a) to the Antofagasta Ridge south of the Mejillones Peninsula (around 24°S; von Huene and Ranero, 2003).

Backscatter intensities are slightly higher at the upper-slope compared to the region of the lower-slope (Fig. 5a). Within the upper-slope area, patches of high backscatter up to 2 km wide are visible. Where covered by sub-bottom profiler data, the high backscatter anomalies correspond either to the flanks of east-west trending ridges, which are often associated to canyon flanks, or to thethalweg of upper-slope canyons and channels (Fig. 5c). The ridges (canyon flanks), which are elevated up to 80 m above the surrounding seafloor, do not host a sedimentary cover (Fig. 5c).

Between 21°S – 22°S the upper slope is dissected by multiple canyons that are likely related to the Loa River estuary at 21.4°S (Fig. 1). The canyons are up to 1 km wide and cut up to 200 m into the surrounding seafloor. All canyons terminate downslope at the large N-S
striking seafloor scarps, indicating the active nature of the underlying normal faults. Around 22°S, two canyons extend across the upper slope into the area of the middle slope. These canyons, which we refer to as northern and southern Tocopilla Canyon, terminate in the area of the large embayment at the middle slope. Farther to the south, a couple of canyons are visible at the upper slope, but again all terminate around water depths of 2000 to 3000 m.

4 Discussion

In the following we discuss geologic structures and processes that may cause the observed variations in the structure of the convergent margin and evaluate how the structural variations may impact marine forearc deformation over short (earthquake cycle) and long (millions of years) timescales. The different regions (oceanic plate, northern and southern study area) are not treated individually in the discussion.

Figure 9 shows a representative structural cross-section for the North Chilean margin and close-up views of key bathymetric features, and integrates many of the structural elements discussed below. Trench-parallel upper-plate normal-faults represent the most pervasive structural element of the North Chilean marine forearc (also compare Geersen et al., 2018). In some areas the normal faults extend from the upper-slope to the deformation front (Fig. 9b). The normal faults in the offshore forearc are aligned with extensional structures onshore between the coastline and the West Andean Thrust (Victor et al., 2004; Allmendinger and González, 2010; Armijo et al., 2015). The ~1000 m high-coastal scarp (Fig. 9a), which may represent the morphologic expression of a large, west-dipping normal-fault (Armijo and Thiele, 1990), resembles the morphologic expression of some of the large (hundreds of meters vertical displacement) normal faults observed in the marine forearc (e.g. Fig. 8c).

4.1 Along-trench variations in bending-related deformation and possible implications

Off Northern Chile, bend faulting is well expressed in the bathymetry by the progressive increase in the roughness of the oceanic plate approaching the trench axis (Figs. 1-2). However, the style and amount of bending-related deformation changes fundamentally around 20°S, where the trench axis strike bends from NW-SE to N-S (Fig. 2b). South of 20°S, bending related faulting is manifested in N-S striking horst-and-graben structures that result from the combination of landward and seaward dipping new bending-induced faults unrelated to pre-existing structures. Vertical seafloor offsets above the N-S faults reach up to 800 m and increase towards the trench axis, indicating the active nature of the faults. North of 20°S, trench-parallel horst-and-grabens are not observed in the trench region. Here, the NW-SE striking spreading fabric represents the only structural lineaments at the seafloor.

Different authors have shown evidence that pre-existing normal faults that formed at the spreading center and are oriented <30° oblique to the axis of bending can be reactivated during plate-bending without the development of new bend-faults (Masson, 1991; Ranero et al., 2005; Graindorge et al., 2008). For northern Chile (north of 20°S) and possibly wide areas off southern Peru, which are not yet covered by multibeam bathymetric data, it seems that bending-related relief along the trench is formed only by reactivation of the NW-SE trending fabric formed at the spreading center.
Bending-related faulting has been proposed to promote hydration of the oceanic plate by allowing seawater to infiltrate the mantle (Ranero et al., 2003). The degree of mantle hydration may be one of the main parameters determining the intensity of intermediate depth seismicity. The extent of mantle hydration, and its impact on seismicity, however, may depend on the intensity of bend-faulting prior to subduction. For the Alaska margin, Shillington et al. (2015) show changes in seismic velocity supporting increased hydration of the oceanic plate in an area of comparatively more intense deformation, where inherited spreading fabric is reactivated. In this context the change from the formation of new bend-faults south of 20°S to reactivation of spreading-related faults north of 20°S may imply that more water penetrates into the oceanic uppermost lithosphere in the northern region. This, however, requires confirmation with seismic velocity measurements from the northern margin segment.

It is interesting to note that the lack of horst-and-graben structures, and thus new bending faults in the trench, between 18°S and 20°S correlates with a segment of the trench with seafloor depth of ~6500 m (Fig. 2a). This is up to 1500 m shallower than to the south, where nascent bend faults occur, and to the north, where coverage by swath bathymetric data is not adequate for detailed mapping of bending fault orientation. The well-displayed tectonic fabric of the oceanic plate at the trench axis in our study area indicates that the trench depth is not controlled by basement uplift rather than variations in sediment infill (Fig. 2). A similar shoaling (arching) of the subducting plate is observed along the Cascadia subduction zone off western Washington and southern British Columbia (Crosson and Owens, 1987; Tréhu et al., 2002; Preston et al., 2003) where the arch is interpreted to develop due to the geometric effect induced by the bend in the subduction-zone. Alternatively, the shallowing of the trench within the bend may be due to the presence of a thicker and more buoyant crust associated with the Iquique Ridge. Both mechanisms may be active in our study area, a hypothesis that is currently being tested through analysis of large-aperture marine seismic data acquired across the Iquique Ridge during MGL1601 (Tréhu et al., 2017; Myers et al., 2018).

4.2 Tectonic control on structural variations along the lower continental slope

Most seismic lines (except for SO104-07, Fig. 7d) reveal the presence of a ~5 km wide frontal prism at the lowermost continental slope (Figs. 2 and 7). Given the very limited amount of sedimentary trench fill, which could be used to build the prism, it is likely that the prism is comprised of milled rock that is sourced from the toe and base of the upper-plate, which is recycled into a prism-forming mélangé. This may also explain the lack of internal structure imaged within the frontal prism (Fig. 7). The only significant along-strike variation in the structure of the lowermost continental slope occurs between 21°S and 22°S, where a thin and faulted upper plate is visible and where a frontal prism is absent (Figs. 4 and 7d). Geersen et al. (2018) used this observation in combination with critical taper analyses, seismicity, and regional plate-coupling to suggest that the normal faults at the lower slope may have resulted from shallow, possibly near-trench breaking earthquake ruptures in the past. In their model, a reduction in basal friction together with a positive Coulomb stress change, both induced by seismic rupture of the underlying shallow plate-boundary, can bring the outermost forearc wedge to extensional failure (see Cubas et al., 2013).

Other prominent morphological features at the lower continental slope are the embayments in the northernmost part of the study area that open to the trench (Figs. 3 and 6b). Although we cannot ultimately resolve the processes that formed the embayments, subduction of lower-
plate topography is a possible candidate. The main argument for this is the detailed seafloor morphology in the area of the embayments, which shares many aspects usually related to seamount subduction. The embayments are outlined by radial, predominantly seaward dipping seafloor scarpus. Elsewhere similar radial structures at the lower slope are often interpreted to have formed due to collapse of the upper-plate in the wake of excess topography on the subducting plate (Dominguez et al., 1998; Masson et al., 1990; Ranero and von Huene, 2000; Collot et al., 2001). Upslope of the embayments, landward dipping seafloor scarpus are also observed, leading to the formation of small but distinct ridges at the seafloor. If caused by a subducting seamount, the landward dipping scarpus upslope of the embayments could result from back-thrusting in front of the leading flank of the seamount or from gravitational collapse of the upper-plate and associated normal faulting behind the trailing flank of the seamount (Dominguez et al., 1998; Ruh et al., 2016). That the embayments are correlated with the projection of the Iquique Ridge under the marine forearc also supports the hypothesis that they formed due to subducting plate topography. From seismic data there is evidence for the presence of excess topography on the subducting plate in the larger area (Geersen et al., 2015). However, the embayments are up to 30 km wide, which is wider than the lower-slope furrows attributed to seamount subduction at other erosional margins. For example, a 15-20-km-wide breaches in the lower slope were caused by the 20-30 km wide and 3 km tall under-thrust seamounts that cause the large Jaco and Parrita Scars and slides off Costa Rica (figure 8 in Harders et al., 2011). The spatial extent of such reentrant structures may depend both on the size of the subducting topography and on the rheology of the overriding plate. In addition, the large embayments off Northern Chile may result from the combined effect of multiple subducting positive-relief elements located in close vicinity.

4.3 Absence of indications for focused fluid-seepage

The North Chilean marine forearc is pervasively cut by faults, which might support fluid flow and related fluid-seepage at the seafloor (Figs. 3-4) as described in numerous other margins (e.g. Sahling et al., 2008; Ranero et al., 2008). However, the new multibeam data and the sub-bottom profiler lines (Figs. 3-5) do not reveal clear indications for abundant fluid seepage along the margin. Large mud mounds hundreds of meter wide and tens to >100 m tall, as observed at the erosional Central American forearc (Klaucke et al., 2008; Sahling et al., 2008), have not been observed in our data, which are high enough resolution to display such structures if they were present. Patches of high backscatter are present at the upper continental slope (Figs. 3-4). In the case of the Central American margin, similar features generally correspond to authigenic carbonate pavements at the seafloor. The sub-bottom profiler data, however, indicate that off Northern Chile the high backscatter patches often represent either east-west striking ridges that are exposing indurated material, possibly brought to the surface by a combination of tectonic and erosion processes, or topographic plateaus that are starved of sediment and that possibly represent metamorphic continental basement that crops out at the seafloor (Fig. 5). The steeply dipping underlying strata may create outcrops of indurated layers at the seafloor. It remains unclear whether the pinnacles that rest on these plateaus (Fig. 5c) are of tectonic origin or whether some represent fluid-seepage related features or deep-water coral bioherms. During cruises SO244 (Behrmann et al., 2016; Kopp et al., 2016) and MGL1610 (Tréhu et al., 2017), no acoustic evidence of gas flares in the water column was observed in either in the EM 122 multibeam or Parasound data.
Combining these observations, there is little evidence for focused fluid seepage along the North Chilean marine forearc. It should be noted, however, that the spatial resolution of the hull-mounted backscatter data collected with R/V SONNE and R/V Marcus G. Langseth (25 m at maximum) does not equal the high-resolution deep-towed side-scan sonar data (1 - 6 m pixel size) collected along large parts of the Central American forearc, which was used to detect some of the smaller structures (Klaucke et al., 2008; Sahling et al., 2008; Buerk et al., 2010).

The paucity of seep sites off Northern Chile might be caused by several factors. One is the comparatively limited number of fluid sources. The seismic data shown in Figure 7 underline the sediment-starved nature of the trench. Along vast parts of the margin, the thickness of the sediments that rest on the igneous oceanic basement is <100 m, and this sediment is likely of pelagic origin with low porosity and limited organic matter. Slope sediments are present at the upper and middle slope but their thickness does not exceed 1 s TWT and usually thins to < 0.5 s TWT downslope (Fig. 8). Compared to Central America there is much less sedimentary material in the forearc (including the plate-boundary at depth) that could act as a source for fluids.

What may also contribute to the lower abundance of focused seep sites off Northern Chile (compared to Central America) is the fact that the North Chilean forearc is more affected by normal faulting. The pervasive faulting and fracturing may result in highly distributed fluid flow, and diffuse seepage is harder to detect in marine geophysical data.

4.4 Nature of the basement under the upper slope

From its dimension, the large N-S trending ridge at the upper continental slope that extends over a latitudinal distance of ~500 km (also compare von Huene and Ranero, 2003), represents the most prominent upper-plate morphologic feature off Northern Chile. Between the cities of Arica and Iquique (~18-20°S) the ridge was already mentioned in the seminal papers of Coulbourn and Moberly (1976) and Moberly et al. (1982), who noted it as structural high.

The large N-S trending ridge is bounded by large (hundreds of meters) vertical displacement normal faults (e.g. Figs. 3 and 8c). Interestingly, these faults dip both landward and seaward. This stands in contrast to the predominantly seaward dip of normal faults along the rest of the marine forearc. Sallares and Ranero (2005) and Contreras-Reyes et al. (2014) used wide-angle seismic data to show that the seismic velocity of the basement rock in the marine forearc of Central and Northern Chile is consistent with a basement composed of arc magmatic rocks. At 32°S, offshore Valparaiso, analysis of magnetic anomalies found evidence of a submerged, possibly pre-Jurassic, volcanic arc (Yañez et al., 2001), and a Jurassic to Cretaceous volcanic arc crops out along the coastal in both regions (e.g. Rutland, 1971). The position of these extinct volcanic arcs relative to the active arc is attributed to landward arc migration in response to progressive removal of the continental margin by subduction erosion over long (mainly Mesozoic) periods of time (von Huene and Scholl, 1991).

Following this argument, a possible explanation for the origin of the large N-S striking ridge at the upper continental slope is that it represents the remnants of a Jurassic or pre-Jurassic magmatic arc which is now collapsing in response to extensional faulting. The ridge is
located about 40-80 km seaward of the coast. If it represents a pre-Jurassic magmatic arc, this will offer the opportunity to calculate rates for subduction erosion and coastal retreat over longer (since Jurassic) timescales.

### 4.5 Uplift of the marine forearc

Around 19.2°S a prominent bulge extends into the mid-slope region. Recent uplift in this area is evidenced by the northward migration of the Camarones Canyon and the presence of at least three paleo-canyons to the south of the present day Camarones thalweg (Fig. 3). A similar bulge, although smaller in size, is located at 20°S (Fig. 3). Farther towards the south, the offshore forearc does not show indications for recent uplift, and south of 21°S upper-plate normal faults actually extend to the lower slope.

Clift and Hartley (2007) reported indications for recent (last 2 Ma) uplift of the coastal region of Northern Chile and Southern Peru, which they attributed to tectonic underplating. In their model, much of the material that is eroded from the outer wedge is underplated under the coastal forearc. Instead of being underplated under the coastal forearc the eroded material may also be added to the upper plate farther seaward below the middle continental slope. However, uplift in the marine forearc seems to be localized with distinct mid-slope bulges at 19.2°S and 20°S. It is difficult to explain localized underplating under the middle slope at 19.2°S and 20°S whereas elsewhere the eroded material is underplated about 100 km farther landward.

Another process that may cause localized uplift of the offshore forearc is the subduction of seamounts and volcanic ridges on the oceanic plate (Domínguez et al., 1998; Ranero and von Huene, 2000). The area between 19°S-20°S correlates with the projection of the Iquique Ridge under the marine forearc (Fig. 1a) and from seismic data there is evidence for the presence of excess topography on the subducting plate in the larger area (Geersen et al., 2015). As discussed above, the embayments at the lower continental slope between 19.2°S – 19.9°S may result from the subduction of Iquique Ridge related topography on the oceanic plate. Figure 2 indicates that the Iquique Ridge already collided with the marine forearc and started to subduct. Uplift of the upper plate at 19.2°S and 20°S may therefore result from the subduction of Iquique Ridge related topography. A 3D tomography experiment in combination with a dense net of seismic reflection lines collected in 2016 during Langseth cruise MGL1610 (Tréhu et al., 2017; Davenport et al., 2018) will shed further light on the nature of the subducting part of the Iquique ridge and the mechanisms causing the mid-slope bulges.

### 5 Conclusions

We combined new high resolution multibeam bathymetric data from two recent research cruises with older legacy data, seismic reflection lines and sub-bottom profiler data to derive a tectonic map for the North Chilean marine forearc and the adjacent oceanic Nazca Plate. The new map shows along-strike changes in the structure and the deformation pattern of the upper and lower-plates over an along-margin distance of ~400 km. In contrast to the erosive Central American margin, we find little evidence for focused fluid seepage in Northern Chile. Due to a series of recent (1995, 2001, 2007, 2014) and anticipated (seismic gap) earthquakes,
the North Chilean margin is in the focus of many active research projects. The new tectonic map and interpretations presented here will be useful for informing ongoing and future studies within this margin segment.

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Access to the field data: Bathymetric data from German research cruises (e.g. R/V SONNE SO104, SO244) can be requested through the German Bundesamt für Seeschifffahrt und Hydrographie (BSH) (http://www.bsh.de). The bathymetric data collected during R/V Marcus G. Langseth cruise MGL1610 are available from the Rolling Deck to Repository (R2R) website (http://www.rvdata.us/). Seismic reflection data from R/V SONNE cruise SO104 are stored at the Bundesanstalt für Geowissenschaften und Rohstoffe (BGR) and can be requested through the Geo-Seas data portal (http://www.geo-seas.eu/).

Access to processed data: The following data used in this manuscript are available from the PANGAEA data archive (https://doi.pangaea.de/10.1594/PANGAEA.893034): (I) A gridded DEM (100 m) that combines the bathymetric data from R/V SONNE cruises SO104, SO244, and R/V Marcus Langseth cruise MGL1610. (II) A GeoTIFF that contains the backscatter information from the multibeam data collected during R/V SONNE cruise SO244. ArcGIS shape files that contain the fault traces and other tectonic features shown in Figures 2, 3, and 4 are available from the main author upon request.

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**Figure 1.** (A) Structure of the oceanic Nazca Plate off Northern and Central Chile and Southern Peru. Data from SRTM15+ grid (Becker et al., 2009, Sandwell et al., 2014) overlain with shipborne multibeam bathymetric data (as described in Figure 1B). Red lines on the oceanic plate indicate the age of the oceanic basement (Müller et al., 2008). (B) Multibeam bathymetric data collected during R/V SONNE cruise SO244 (2015) and R/V Marcus G. Langseth cruise MGL1610 (2016) complemented with older data collected over the last two decades during seagoing campaigns with German research vessels (R/V SONNE, R/V METEOR) and masked data from the Global Multi-Resolution Topography (GMRT) synthesis (Ryan et al., 2009). The topographic data are from the Shuttle Radar Topography Mission (Farr et al., 2007). The thick red lines in the marine forearc encircle the rupture areas of the 1995 Antofagasta (Pritchard et al., 2007), 2001 Arequipa (Pritchard et al., 2007), 2007 Tocopilla (Schurr et al., 2012), and 2014 Iquique (Schurr et al., 2014) earthquakes. Dashed black lines are the available seismic reflection lines from the area. See Figures 7-8 for vertical images of the seismic data.
Figure 2. (A) Multibeam bathymetric map of the trench and the oceanic plate. Dashed black lines show the full extent of the seismic reflection lines shown in figures 7-8. (B) Tectonic interpretations as derived from the bathymetric and seismic reflection data. Dashed red lines show the exact spatial extent of the seismic lines shown in figure 7.
Figure 3. (A) Multibeam bathymetric map of the northern study area. Dashed black lines show the full extent of the seismic reflection lines shown in figures 7-8. (B) Tectonic interpretations as derived from the bathymetric and seismic reflection data. Dashed red lines show the exact spatial extent of the seismic lines shown in figures 7-8. Small inset shows thalweg profiles of Camarones and Pisagua canyons.
Figure 4. (A) Multibeam bathymetric map of the southern study area. Dashed black lines show the full extent of the seismic reflection lines shown in figures 7-8. (B) Tectonic interpretations as derived from the bathymetric and seismic reflection data. Dashed red lines show the exact spatial extent of the seismic lines shown in figures 7-8.
Figure 5. (A) Backscatter information derived from the SO244 multibeam data (high backscatter = white) shown on-top of the bathymetric compilation. (B-C) 4 kHz sub-bottom profiler lines across some high backscatter patches at the upper continental slope. While the high backscatter corresponds to sediment starved topographic highs, reduced backscatter intensity is found in the areas of small sedimentary basins located in-between the topographic highs.
Figure 6. Perspective views of the seafloor. (A) The oceanic Nazca Plate prior to subduction with ~N-S striking horst-and-grabens and NW-SE striking spreading fabric. Note that the map is rotated (north is oriented to the bottom-right of the figure). (B) The lower continental slope in the northernmost study area. Seaward and landward dipping scarps at the seafloor form multiple embayments that open to the trench. The deformation front is characterized by multiple slope failures which may result from collision and subduction of lower plate relief.
Figure 7. Seismic reflection data from the lower continental slope. The maps in the right column show the detailed morphology of the seafloor around the seismic sections. TOB = Top Oceanic Basement. (A) Time section of seismic reflection line SO104-22. (B) Time section of seismic reflection line SO104-26. (C) Time section of seismic reflection line SO104-27. (D) Time section of seismic reflection line SO104-07. (E) Time section of seismic reflection line SO104-09.
Figure 8. Seismic reflection data from the upper continental slope. The inset maps show the detailed morphology of the seafloor around the seismic sections. TCB = Top Continental Basement. (A) Time section of seismic reflection line SO104-22. (B) Time section of seismic reflection line SO104-27. (C) Time section of seismic reflection line SO104-09.
Figure 9. (A) Structural cross-section across the North Chilean continental margin. (B-D) Seafloor morphologic expression of some of the structural elements discussed in the manuscript.