Control of high oceanic features and subduction channel on earthquake ruptures along the Chile–Peru subduction zone

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\textbf{Abstract}

We discuss the earthquake rupture behavior along the Chile–Peru subduction zone in terms of the buoyancy of the subducting high oceanic features (HOF’s), and the effect of the interplay between HOF and subduction channel thickness on the degree of interplate coupling. We show a strong relation between subduction of HOF’s and earthquake rupture segments along the Chile–Peru margin, elucidating why these subducting features play a key role in seismic segmentation. Within this context, the extra increase of normal stress at the subduction interface is strongly controlled by the buoyancy of HOF’s which is likely caused by crustal thickening and mantle serpentinization beneath hotspot ridges and fracture zones, respectively. Buoyancy of HOF’s provide an increase in normal stress estimated to be as high as 10–50 MPa. This significant increase of normal stress will enhance seismic coupling across the subduction interface and hence will affect the seismicity. In particular, several large earthquakes ($M_w \geq 7.5$) have occurred in regions characterized by subduction of HOF’s including fracture zones (e.g., Nazca, Challenger and Mocha), hotspot ridges (e.g., Nazca, Iquique, and Juan Fernández) and the active Nazca–Antarctic spreading center. For instance, the giant 1960 earthquake ($M_w = 9.5$) is coincident with the linear projections of the Mocha Fracture Zone and the buoyant Chile Rise, while the active seismic gap of north Chile spatially correlates with the subduction of the Iquique Ridge. Further comparison of rupture characteristics of large underthrusting earthquakes and the locations of subducting features provide evidence that HOF’s control earthquake rupture acting as both asperities and barriers. This dual behavior can be partially controlled by the subduction channel thickness. A thick subduction channel smooths the degree of coupling caused by the subducted HOF which allows lateral earthquake rupture propagation. This may explain why the 1960 rupture propagates through six major fracture zones, and ceased near the Mocha Fracture Zone in the north and at the Chile Rise in the south (regions characterized by a thin subduction channel). In addition, the thin subduction channel (north of the Juan Fernández Ridge) reflects a heterogeneous frictional behavior of the subduction interface which appears to be mainly controlled by the subduction of HOF’s.

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1. Introduction

Along the Chile–Peru Trench the oceanic Nazca plate subducts beneath the South American plate at a current rate of $\sim 6.5$ cm/year (Khazaradze and Klotz, 2003). The Nazca plate hosts a large population of high oceanic features (HOF’s) including hotspot tracks and oceanic fracture zones (FZ’s). Perhaps the most striking feature of the oceanic plate is the Chile Rise, an active spreading center that marks the boundary between the oceanic Nazca and Antarctic plates and is currently subducting at $\sim 46^\circ$S beneath the South American plate. Bathymetric data show four main oceanic hotspot tracks: Nazca, Iquique, Copiapó and Juan Fernández Ridges (Fig. 1). The Nazca Ridge (NR) was formed at the Easter Island hotspot on the Pacific–Farallon/Nazca spreading center (Pilger, 1984), while the Juan Fernández Ridge (JFR) is an off-ridge hotspot formed at the JFR hotspot onto 27 Myr old oceanic crust (Yáñez et al., 2001). The oceanic Nazca plate is further segmented by several oceanic fracture zones formed at the Pacific–Nazca and Antarctic–Nazca spreading centers (Tebbens et al., 1997).

Most of the seafloor features hosted by the oceanic Nazca plate are in the throes of being subducted (Herron et al., 1981; Yáñez et al., 2001; Rosenbaum et al., 2005). Subduction of a HOF may modify the geodynamics and tectonic setting of the outer forearc region, and disrupt and erode material from the overriding plate. In addition, subduction of a HOF is expected to influence dramatically the degree of coupling across the subduction interface and may affect seismicity, in particular the size and frequency of large

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The slip distribution at the subduction interface is highly heterogeneous in terms of the strength or frictional characteristics of the material in the fault zone. These stress heterogeneities have been recognized as asperities (Kanamori, 1994) and barriers (Das and Aki, 1977) to earthquake rupture which control the seismic moment release and rupture area. A seismic asperity is an area with locally increased friction and exhibits little aseismic slip during the interseismic period relative to the surrounding regions. Once the shear yield stress $\tau_o$ (critical shear stress required for failure) along these heterogeneities is reached by the accumulated interseismic shear stress $\tau_1$, the asperity concentrates the coseismic moment release and slip during the earthquake. The regions with little or no slip during rupture propagation are generally called barriers and control the size of the earthquake rupture area (ERA) (Aki, 1979; Kanamori and McNally, 1982).

Once the earthquake initiates ($\tau_1 \geq \tau_o$), the rupture front propagates through regions where the dynamic stress associated with the rupture process is larger than $\tau_o$. If the rupture front encounters an obstacle where the dynamic stress is lower than $\tau_o$, the fault motion slows down and eventually stops (barrier zone). In other words, the interplay between the amount of dynamic stress and $\tau_o$ will define whether a region behaves as an asperity with large slip or behaves as a barrier with no slip. Therefore, it is fundamental to investigate the shear yield stress distribution along the subduction interface. A heterogeneous distribution of $\tau_o$ might be useful to predict the location of local barriers and asperities before occurrence of an earthquake. $\tau_o$ varies over time and depends on frictional properties, fluid pore-pressure, and state stress anomaly. For example, subduction of a rigid seamount will result in an increase of normal stress $\sigma_n$ and hence $\tau_o$, preventing eventual rupture (strong coupling). Likewise, buoyancy of HOF’s will enhance seismic coupling and anomalous increase of $\tau_o$. In contrast, an increase of fluid pore pressure will reduce $\tau_o$ facilitating rupture propagation.

Since HOF’s dramatically modify the geodynamics of the outer forearc and interplate boundary conditions, they are probably the most obvious candidates for asperities and/or barriers. The frictional behavior of HOF’s is clearly influenced by the increase of normal stress at the subduction interface caused by the excess buoyancy of these features (Scholz and Small, 1997). Furthermore, Scholz and Small (1997) claim that the “geometric” normal stress effect due to the need of the overriding plate to accommodate the shape of the HOF is a significant additional source of increasing normal stress at the subduction interface. Thus, the anomalous strong coupling associated with HOF’s (buoyancy and geometry) might...
play an important role on seismicity. This appears to be the case for South America, where many large earthquakes occur in regions where these large Nazca plate features enter the trench.

We study the spatial relationship between large underthrusting earthquakes (\(M_w \geq 7.5\)) and bathymetric heterogeneities on the subducting Nazca oceanic plate along the Chile–Peru subduction zone. Such heterogeneities are manifested by subduction of bathymetric anomalies (HOF's), which are likely to affect the seismic coupling across the subduction interface due to their buoyancy. For instance, hotspot seamounts usually host thick oceanic crust and magmatic material that may have erupted, ponded beneath or intruded the base of the oceanic crust (magmatic underplating) (Koppers and Watts, 2010). On the other hand, oceanic fracture zones are characterized by a high degree of hydration in both the crust and mantle, and they commonly comprise bodies of serpentinities deep into the oceanic upper mantle (Contreras-Reyes et al., 2008). Thus, buoyancy is linked directly to structural or chemical alteration of the crust or mantle beneath the oceanic feature that stores rocks less dense than its surroundings (Henderson et al., 2009). We use published seismic velocity models across the Nazca Ridge (NR), JFR and Mocha FZ in order to estimate the anomalous normal stress associated with the buoyancy of these HOF's. In addition, several HOF's are identified from bathymetric data and compared with published coseismic slip models of large underthrusting earthquakes in order to explore if subducting HOF correlate with zones of large slip (asperities) and/or little slip (barriers).

In addition, the amount of subducted sediment may also play an important role on megathrust seismicity since the subduction channel might smooth the subduction interface resulting in a homogenous coupled region. It can be expected that a large homogenous coupled region facilitates long lateral rupture propagation (Ruff, 1989; Scherwath et al., 2009; Contreras-Reyes et al., 2010) and hence the earthquake size. Therefore, a thin subduction channel above a subducting HOF may reflect the presence of a seismic barrier while a thick subduction channel would tend to reduce the effect of the anomalous coupled region caused by the subducting HOF. In order to test this hypothesis for some earthquakes recorded along the Chilean subduction we use a compilation of geodetic data and seismic data constraining the subduction channel thickness and we discuss the interplay between subduction channel thickness and the dual asperity-barrier behavior. Identification of both subduction channel thickness variation and magnitude of buoyancy forces associated to HOF's would provide a very useful context for future seismological works.

2. Relation between high oceanic features and earthquake rupture areas along the Chile–Peru subduction zone

The Chile–Peru subduction zone is an extremely active convergent plate margin producing large earthquakes (\(M_w > 8\)) about every 10 years and capable of generating at least one tsunami-generating mega-thrust earthquake per century. This is documented in historical accounts and instrumentally well-recorded events that date back to the 14th century (Darwin, 1851; Beck et al., 1998; Bilek, 2010). Fig. 1A shows the location of the rupture areas of the largest events documented over the 19th to 21st centuries. These ERA's are spatially linked to the locations of the main oceanic incoming features of the Nazca plate. One of the most important observations is that at least one rupture limit for each event coincides approximately with a subducting HOF, and about half of the ERA's are bounded by HOF's. In the following we discuss some examples:

(a) The southern Peru–northern Chile region (15.5–23°S): this region is characterized by two active seismic gaps that mark the zones of the (\(M_w = 8.7\)) 1868 southern Peru and the (\(M_w = 8.9\)) 1877 northern Chile mega-thrust earthquakes (Comte and Pardo, 1991). The southern Peru 1868 rupture segment broke previously in 1604 and 1784 in a similar, approximately 450 km long rupture. The northwest part of that segment broke again in an Ms = 8.4 earthquake in 2001, filling at least partially the south Peru gap (Ruegg et al., 2001). This event propagated for \(\sim 70\) km before encountering a 6000 km² fault area that acted as a barrier. Approximately 30 s after the start of rupture, this barrier began to break when the main rupture front was \(\sim 200\) km from the epicenter (Robinson et al., 2006). In this case, the subducted feature acted as both a barrier and an asperity during a single event and was associated with the oceanic Nazca FZ (Robinson et al., 2006). Further to the south, the northern Chile seismic gap (18.5–23°S) is roughly coincident with the subduction of the Iquique Ridge (IR). Furthermore, the 2007 Tocopilla earthquake (\(M_w = 7.7\)) spatially correlates with the southern part of the IR, while the 1995 Antofagasta event (\(M_w = 8.0\)) is bounded by the IR and Tal–Tal ridge in the north and south, respectively (Figs. 1A and 1B).

(b) Central Chile (31–37.5°S): the most recent large Chilean earthquake occurred on February 27th, 2010 (\(M_w = 8.8\)) and propagated northward and southward achieving a final rupture length of about 450 km (34–38°S; Lay et al., 2010). In the same region, 176 years ago in 1835 a large earthquake with an estimated magnitude of \(M_w \sim 8.5\) was reported by Darwin (1851). North of the rupture area of the 2010 earthquake, two megathrust earthquakes occurred: the 1906 (\(M_w \sim 8.4\); Beck et al., 1998) and 1985 (\(M_w = 7.8\); Barrientos, 1995) events. The southern limit of the 2010 earthquake is coincident with the Mocha FZ, while the northern limits of the 1906 and 1985 earthquakes are coincident with the subduction of JFR (Fig. 1). Similarly, the southern limit of the 1943 (\(M_w = 7.9\)) earthquake correlates with the JFR.

(c) The 1960 mega-thrust event (37.5–46°S): the largest instrumentally recorded earthquake, which occurred in 1960 with a magnitude of \(M_w = 9.5\), ruptured over a length of almost 1000 km from \(\sim 37.5°S\) to \(46°S\) near the Chile Triple Junction (CTJ) of the Nazca, Antarctic and South American plates (Moreno et al., 2009). The Chile Rise is currently subducting at the CTJ, which is a region characterized by a shallow trench associated with the buoyancy of this active spreading ridge (Herron et al., 1981). The 1960 earthquake ruptured across six major fractures zones and it stopped near the Mocha FZ to the north and at the CTJ to the south. In this case, the fractures zones involved did not act as barriers, except for the Mocha FZ and Chile Rise which are the northern and southern earthquake rupture limits, respectively.

Other examples showing spatial correlations between ERA locations and incoming HOF's can be inferred from Fig. 1A. In addition, such spatial correlations also appear to apply for the historical seismic segmentation of the last 500 years (see Fig. 1B). Considering the historical ERA estimates, not every limit of an ERA coincides with the subduction of an HOF, but at least one limit of each ERA is coincident with a prominent HOF as is shown in Fig. 1A, and half of the events are bounded at both limits by HOF's. Thus, the large earthquakes of the Chile–Peru margin appear to be highly segmented by bathymetric features. Obviously, the uncertainty of earthquake rupture areas are large for historical reports and the correlation between HOF and ERA position is far from perfect for old events (Fig. 1). Nevertheless, events instrumentally recorded during the 20th to 21st centuries present a reasonably good correlation between HOF and ERA position (Fig. 1). Additionally, geodetical data provide further constraints for the size of ERA's and fault slip distribution as we discuss in the following section.
3. Geodetical observations

Geodetical data provide further and crucial evidence for seismic coupling at the subduction interface, allowing identification of regions with high coseismic slip (asperities) or no slip (barriers). In this section, we show a compilation of fault slip models for (1) the active seismic gap of the southern Peru-north Chile region (15.5–23°S), (2) Central Chile (31–38°S), and (3) the region of the largest ever recorded 1906 earthquake (~38–46°S).

3.1. The active seismic gap of south Peru-north Chile

The region between the Ilo Peninsula (15.5°S, South Peru) and the Mejillones Peninsula (23.5°S, North Chile) represents a major seismic gap that has not experienced a significant megathrust earthquake since the 1868 earthquake (Mw = 8.7) and the 1877 earthquake (Mw = 8.9) (Comte and Pardo, 1991, Fig. 2A). The 1995 Antofagasta (Mw = 8.1) and the 2001 Arequipa (Mw = 8.4) earthquakes appear to have provided an extra load at both extremities of the remaining ~500 km length unruptured segment. After the 1877 event and before the 2007 Tocopilla earthquake, a few events (Mw < 7.5) have been reported in the region (Comte and Pardo, 1991; Tichelaar and Ruff, 1991) but they were not large enough to release a significant part of the 9 m of slip deficit accumulated in the gap in the last 130 years (Béjar-Pizarro et al., 2010). Fig. 2B shows the slip distribution with 3 m contour interval of the 2001 Arequipa earthquake (Mw = 8.4) based on the coseismic slip model of Pritchard et al. (2007). This event only partially filled the rupture zone of the 1868 earthquake (Fig. 2A). The rupture front propagates southward and encounters the Nazca FZ at 70 km from the hypocenter. Although, this HOF apparently acted as barrier, the rupture continued around this barrier, which remained unbroken for ~30 s and then began to break when the main rupture front was ~200 km from the epicenter (Robinson et al., 2006). The 1995 Antofagasta earthquake (Mw = 8.1) ruptured a 180 km-long segment just south of the main 1877 gap. Geodetic slip inversions give a consistent picture of the coseismic slip distribution during the earthquake with a single asperity slipping 5–10 m (Chlieh et al., 2004; Fig. 2B). The earthquake rupture area is limited by the southern part of the Iquique Ridge in the north and the Tal–Tal ridge in the south (Fig. 2B). Thus, these features appear to act as barrier for rupture propagation.

The 2007 Tocopilla event (Mw = 7.7) ruptured the deeper part of the seismogenic interface (30–50 km) and did not reach the surface, indicating that the shallow part of the seismogenic interface (from the trench to 30 km depth) remains unbroken with the exception of the southern edge of the rupture (at ~23°S), where coseismic slip seems to have propagated up to ~25 km depth (Béjar-Pizarro et al., 2010). The fault slip model shows slip concentrated in two main asperities with maximum values lower than 2 m (Béjar-Pizarro et al., 2010). The rupture area of this event is roughly coincident with southern region of the IR which may have arrested the updip rupture propagation (Fig. 2B).

3.2. Central Chile (31–37.5°S)

The rupture area of the most recent megathrust earthquake; the 2010 event (Mw = 8.8), is characterized by two regions of high seismic slip with maximum slip of 10–15 m (Lay et al., 2010; Moreno et al., 2010). Between these asperities, the rupture bridged a zone that was creeping interseismically with consistently low coseismic slip. The bilateral rupture propagated towards the south, where one...
as a second asperity failed (Moreno et al., 2010).

The southern limit of the rupture area is fairly coincident with the subduction of the Mocha FZ (Fig. 3A), and hence this HOF might have acted as seismic barrier. In the northern limit of the 2010 earthquake, the rupture apparently stops some 100 km south of the subduction JFR according to the teleseismic fault slip model of Lay et al. (2010) (Fig. 3B), Moreno et al. (2010) proposed that pre-stress was lowered in the northern periphery of the 2010 ERA by previous earthquakes such as 1906 and 1985 events resulting in a low-stress barrier. Nevertheless, the JFR appears to be the northern barrier for both the 1906 and 1985 events, and further north the southern limit of the 1943 earthquake (M = 7.9) (Fig. 3). Thus, the JFR plays a crucial role on seismic segmentation in south central Chile. In fact, at ~72.5°/32.5°S, magnetic anomalies reveal the subduction of a large seamount belonging to the Juan Fernández hotspot track (Yáñez et al., 2001). The size of this seamount is about 40 km in diameter and 4 km high, and therefore equivalent to the size of the O’Higgins guyot which host a wide crustal root and a magmatic underplating body beneath the Moho (Kopp et al., 2004, see Section 4). Thus, the subduction of the JFR might considerably affect seismicity in this segment of the Chilean margin.

3.3. The 1960 megathrust earthquake (Mw = 9.5)

Fig. 4 shows the fault slip model inverted by Moreno et al. (2009) using the geodetic data compiled by Plafker and Savage (1970). The northern limit of the 1960 ERA is coincident with both the Peninsular de Arauco and the subducting Mocha FZ, while in the southern fault slip ceased a few kilometres north of the CTJ. The most constrained region of the model is north of Isla de Chiloé, where most geodetical data were acquired (Moreno, personal communication). Here, the two main slip patches are bounded by fractures zones, and these HOF’s present a reduction of fault slip or increase of coupling.

4. Seafloor morphology and buoyant zones of the oceanic Nazca plate

Individual features on the subducting Nazca oceanic plate are large, reaching elevations of up to 5 km above the surrounding regional seafloor depth and widths up to 70 km (Fig. 1). The buoyancy of these features is a key factor controlling interplate seismic coupling and depends on the local mode of isostatic compensation. Fig. 5 shows the seismic structure of the NR (Hampel et al., 2004), easternmost portion of the JFR (Kopp et al., 2004), and the Mocha FZ (Contreras-Reyes et al., 2008) derived from wide-angle seismic transects. The NR comprises of an anomalously thick crust (~14 km), twice the thickness of the typical oceanic Nazca crust, and characterized by underplating (Hampel et al., 2004) (Fig. 5A). Both the thick crust and the underplated subcrustal material represent unusual bodies with density similar to gabbro, but lower than the surrounding mantle peridotite. The easternmost portion of the JFR corresponds to the O’Higgins seamount group (3–4 km high) which shows a moderately thick crust according to the seismic model of Kopp et al. (2004) (Fig. 5B). However, the major buoyancy source is owed to reduced mantle velocities and hence densities. That is, low upper mantle velocities in the range of 7.5 ≤ Vp ≤ 8.0 km/s are commonly attributed to underplating of melt bodies beneath the crust which are usually centered at the volcanic edifice, and an asymmetric distribution is rather uncommon. Fig. 4B shows that the reduced mantle velocities extend further trenchward suggesting mantle serpentinization in the outer rise region (Kopp et al., 2004). Thus, the sereniptized mantle also represents an extra buoyant force since the bulk density of peridotite decreases with the degree of hydration. Regardless if low mantle seismic velocities and densities are caused by mantle serpentinization or magmatic underplating, both processes are an important source for buoyancy of the oceanic lithosphere.

Contreras-Reyes et al. (2008), reported both low crustal and upper mantle velocities beneath the Mocha FZ which were interpreted in terms of an intensely fractured and hydrated upper oceanic lithosphere. The maximum depth of mantle serpentinization is up to 6–8 km within the uppermost mantle, providing an important buoyancy source (Fig. 5C). Thus, buoyancy due to crustal and upper mantle hydration may contribute an important mechanism for increased coupling at the subduction interface.

Table 1 and Fig. 5 summarize the resulting anomalous normal stress, Δσn, associated with the buoyancy of the NR, JFR and Mocha FZ. Δσn ranges between 10 and 50 MPa providing a significant

Fig. 3. (A) Earthquake segmentation along the south central Chile margin is indicated by ellipses that enclose the approximate rupture areas of historic earthquakes (Beck et al., 1998; Ruegg et al., 2009). Starts denote epicenter location. (B) Fault slip distribution of the 2010 Maule earthquake obtained by inversion of teleseismic P waves, SH waves, and resulting R1 source–time functions after Lay et al. (2010). Slip distribution at 0.5 m contour interval of the 1985 earthquake based on the dislocation model of Mendoza et al. (1998).
increase of coupling at the plate interface compared to the apparent stress drop of about 30 MPa reported as an average for Chilean interplate earthquakes (Leyton et al., 2009). Our \(\Delta\sigma_n\) estimates are based only on the buoyancy of HOF’s and represent a minimum of the actual anomalous coupling. Actually, \(\Delta\sigma_n\) is likely to be higher due to the additional normal stress associated with the accommodation of a subducting high relief topographic feature (topographic effect). Scholz and Small (1997) roughly estimated that a typical seamount 4 km high and 60 km wide results in an extra normal stress of 100 MPa. Thus, the resulting \(\Delta\sigma_n\) is expected to be higher than estimates shown in Table 1 and Fig. 5 for large seamounts such as the NR and JFR.

5. Discussion

5.1. Role of subducting HOF’s on seismic coupling

The spatial correlation between ERA locations and incoming HOF’s indicates that seamounts and fracture zones play a key role in the dynamics of the earthquake rupture along the Chile–Peru subduction zone. For instance, part of the southern Peru seismic gap and the entire northern Chile seismic gap is coincident with the subduction of the IR, which likely controls at least in part the degree of coupling in the seismogenic zone. The IR is composed by a number of seamounts (1.0–2.0 km high) and it is surrounded by a smooth and broad region of shallow seafloor (swell). This swell is 200–300 km wide and more than 500 m in elevation (Fig. 2). The gray region representing the IR’s swell shown in Fig. 2B is based on the shallow topography of this HOF inferred from the bathymetric data. The extension and distribution of the swell is not clear due to presence of the prominent outer rise topography. McNutt and Bonneville (2000) proposed that the swell is mainly a consequence of the buoyancy of magmatic underplating beneath the Moho or crustal thickening, as imaged by seismic refraction experiment. Unfortunately, the deep structure of the IR has not yet been imaged by seismic refraction data and a high resolution image of the IR deep structure is unknown. Nevertheless, the 3D density model of Tassara et al. (2006) shows that the IR hosts an anomalously thick crust \(\sim 15\) km thick providing an important source of buoyancy. A smooth and uplifted topography (like the IR swell) might provide a

### Table 1

The anomalous normal stress \(\Delta\sigma_n\) is calculated as \(\Delta\sigma_n = \Delta\rho g H + \rho g \Delta H\), where the \(\Delta\sigma_n\) (buoyancy force) depends on the thickness of the corresponding anomalous crustal thickness \(\Delta H_c\), thickness of the underplated magmatic material beneath the crust and/or thickness of the serpentinized mantle \(\Delta H_m\). \(\Delta\sigma_n\) also depends on the mantle–crust density contrast \((\Delta\rho_m = 530\, \text{kg/m}^3)\), “normal” mantle-serpentinized mantle density contrast \((\Delta\rho_m = 230\, \text{kg/m}^3)\) and/or “normal” crust-altered crust density contrast \((\Delta\rho_c = 50\, \text{kg/m}^3)\). \(g = 9.81\, \text{m/s}^2\) is the acceleration due to gravity. \(\Delta H_m\) and \(\Delta H_c\) are based on the information shown in Fig. 5.

<table>
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<th>Oceanic feature</th>
<th>(\Delta H_c) (km)</th>
<th>(\Delta\rho_m) (kg/m³)</th>
<th>(\Delta H_m) (km)</th>
<th>(\Delta\rho_m) (kg/m³)</th>
<th>(\Delta\rho_c) (kg/m³)</th>
<th>(\Delta\sigma_n) (MPa)</th>
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<td>50</td>
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<tr>
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<td>530</td>
<td>3.0</td>
<td>230</td>
<td>0</td>
<td>10</td>
</tr>
<tr>
<td>Mocha FZ</td>
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<td>–</td>
<td>8.0</td>
<td>230</td>
<td>50</td>
<td>20</td>
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Fig. 5. (upper) Structure and interpretation of the (A) Nazca Ridge, (B) easternmost portion of the Juan Fernández Ridge, and (C) Mocha FZ based on the 2D seismic velocity model of Hampel et al. (2004), Kopp et al. (2004), and Contreras-Reyes et al. (2008), respectively. Map locations of seismic profiles are shown in Fig. 1A. (below) Direct comparison of these HOF’s structure with typical Nazca oceanic crust (6.5 km thick). The anomalous normal stress $\Delta \sigma_n$ (buoyancy force) depends on the thickness of the corresponding anomalous crustal thickness ($\Delta H_c$) and on thickness of the underplated magmatic material beneath the crust and/or thickness of the serpentinized mantle ($\Delta H_m$). $\Delta \sigma_n$ also depends on the mantle–crust density contrast ($\Delta \rho_{mc} = 530 \text{ kg/m}^3$), “normal” mantle–serpentinized mantle density contrast ($\Delta \rho_{m} = 230 \text{ kg/m}^3$) and “normal” crust–altered crust density contrast ($\Delta \rho_{c} = 50 \text{ kg/m}^3$).

See further details in Table 1. Density values are taken from the density model of Tassara et al. (2006).
high degree of coupling homogeneously distributed. Furthermore, the IR subduction is highly oblique beneath South America, effectively spreading the anomalously coupled zone over a wider region. Similarly, the obliqueness of other subducted HOF's can extend the anomalously coupled region along strike (e.g., Nazca FZ, Challenger FZ, Iquique FZ and Mocha FZ). For large HOF's such as the IR, oblique subduction can explain the large size and recurrence interval of great earthquakes.

The impact of subducting seamounts on the degree of coupling has been broadly documented (e.g., Scholz and Small, 1997; Cloos and Shreve, 1996) in contrast to the effect of fracture zones on seismic coupling which has been given less attention. Even though the importance of oceanic fracture zones on seismic segmentation has been proposed by some authors (e.g., Ruff and Miller, 1994), the mechanism responsible for such segmentation has not been discussed much. We proposed here that the deep serpentinitized bodies within the uppermost mantle hosted by oceanic FZs affect dramatically the buoyancy and hence the degree of coupling (Fig. 5). Consequently, their impact on seismicity is of similar importance to hotspot seamounts. For instance, the subduction of the Nazca FZ, Iquique FZ, Challenger FZ and Mocha FZ appears to strongly correlate with the northern and southern limits of several ERA's (Fig. 1A). In most cases these FZ's coincide with the ERA's borders and hence act as seismic barriers.

5.2. The interplay between HOF and subduction channel thickness and its impact on rupture propagation

An additional and fundamental element affecting the coupling degree along the subduction interface is the subduction channel thickness (e.g., Ruff, 1989). For example, a rough subducting HOF covered by a thin subduction channel translates into highly heterogeneous strength distribution, and between regions of high relief would behave aseismic due to its weak coupling, which will result in small earthquake generation (Ruff, 1989). This process is illustrated in Fig. 6A, which shows the typical structure of the northern erosional Chile margin characterized by a thin subduction channel. A thin subduction channel would result in a potential asperity and large earthquake nucleation if the dynamic stress is larger than the yield shear stress, otherwise it will behave as a seismic barrier. On the other hand, a thick subduction channel (Fig. 6B) will allow a smoother strength distribution along the plate interface resulting into highly homogeneous strength distribution (Ruff, 1989). It can be expected that a thick subduction channel would reduce coupling above topographic heterogeneities and it would increase plate coupling between high topographies (Fig. 6B). Thus, these two processes together will favor a homogeneous plate contact and consequently earthquake rupture propagation and generation of giant earthquakes.

Recently, seismic studies have reported large variations of the subduction channel thickness along the Chile–Peru margin (Schwath et al., 2009; Contreras-Reyes et al., 2010). North of the JFR, the erosional northern Chile and Peruvian margin are characterized by a thin subduction channel, typically thinner than 300 m (Reichert et al., 2002). On the other hand, the subduction channel south of the JFR and north of the Mocha FZ is relatively thin (< 1.0 km thick) compared to the subduction channel south of the Mocha FZ and north of the CTJ (> 1.5 km) (Contreras-Reyes et al., 2010). Thus, we would expect that south of the Mocha FZ, the thick subduction channel has a strong influence on earthquake rupture propagation. In contrast, north of the Mocha FZ where the subduction channel becomes thinner we would expect that the HOF's mainly control the size of the earthquake. Although the limits of several earthquakes do not perfectly correlate with HOF subduction, this may be explained by local anomalous regions of thick subduction channels that would tend to homogenize the subduction interface facilitating rupture propagation.

The Mocha FZ is a remarkable oceanic feature controlling the size of both the 1960 and 2010 earthquakes. In fact, the highly coupled region where the Mocha FZ subducts appears to have arrested the southward propagation of the recent Maule 2010 earthquake (Fig. 3). On the other hand, the Mocha FZ stopped in the north the
rupture propagation of the megathrust 1960 event (Fig. 4). This earthquake then propagated all the way south to just north of the CTJ. The thick subduction channel along the rupture zone may have provided enough smoothness for a long rupture propagation across the fracture zones involved. The thin subduction channels northern of the Mocha FZ and near the CTJ were too thin to reduce plate coupling, and rupturing ceased before crossing the Mocha FZ and the Chile Rise in the north and south, respectively (Scherwath et al., 2009).

In addition, the fault-slip model shows that the slip patches (Moreno et al., 2009) are approximately bounded by the subduction of fracture zones, which suggests that these HOF’s tend to act as barriers, but both the large amount of energy carried by the rupture front and the thick subduction channel have allowed further rupture propagation (Fig. 4).

A giant earthquake can also be generated in a region with a thin subduction channel, if the topography of the subducting feature is rather smooth and wide enough as for instance an oceanic plateau. This case could be the active seismic gap of north Chile, where the IR subducts and no major earthquake (M> 8.5) has occurred since more than 120 years. The smooth topography and deep crustal root of the IR (Tassara et al., 2006) favors efficiently the plate contact, more than 120 years. The smooth topography and deep crustal root of the IR (Tassara et al., 2006) favors efficiently the plate contact, and hence reduce plate coupling and an accretionary wedge that has a large degree of coupling and hence reduce plate coupling and an accretionary wedge that has a large degree of coupling. We propose that two important factors affect the earthquake rupture behavior: (1) the buoyancy of the subducting HOF, and (2) the effect of the interplay between HOF and subduction channel thickness (Kopp, 1991). The IR is the HOF and subduction channel that is the thickest in the area, and hence has more buoyancy (Mansinha et al., 1978). The thick subduction channel smooths the degree of coupling and hence reduce the yield shear stress at the subduction interface allowing lateral earthquake rupture propagation. Future work should focus on more seismogenic HOF’s and subduction channels, in order to better understand their internal structures to provide better constraints on rupture earthquake dynamics modelling and seismic hazard assessment.

6. Summary

We propose that two important factors affect the earthquake rupture behavior: (1) the buoyancy of the subducting HOF, and (2) the effect of the interplay between HOF and subduction channel thickness (Kopp, 1991). The IR is the HOF and subduction channel that is the thickest in the area, and hence has more buoyancy (Mansinha et al., 1978). The thick subduction channel smooths the degree of coupling and hence reduce the yield shear stress at the subduction interface allowing lateral earthquake rupture propagation. Future work should focus on more seismogenic HOF’s and subduction channels, in order to better understand their internal structures to provide better constraints on rupture earthquake dynamics modelling and seismic hazard assessment.

Acknowledgments

We thank Javier Ruiz and Sergio Ruiz for fruitful discussions. Eduardo Contreras-Reyes acknowledges the support of the Chilean National Science Foundation (FONDECYT) project # 11090009. We thank Marta Béjar-Pizarro and Marcos Moreno for providing the digital coseismic slip models. We also thank M.A. Gutscher, an anonymous reviewer, and the Editor George Helfrich for comments on the manuscript. We appreciate comments from the English revision task group. We appreciate comments on the English manuscript. We appreciate comments on the English manuscript.

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