Widespread deformation of basin-plain sediments in Aysén fjord (Chile) due to impact by earthquake-triggered, onshore-generated mass movements

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ABSTRACT

The 2007 Mw 6.2 earthquake in Aysén fjord caused widespread basin-plain deformation and has had prehistorical predecessors. Both superficial and buried deformed basin-plain deposits are mapped using multibeam bathymetry and seismic-reflection (sparkler) profiling. The seismic signature of the sediment was ground-truthed with short cores on key locations. Deformed basin-plain deposits induced by the 2007 earthquake can be divided in frontally emergent and confined deposits, with both a deep and shallow basal shear surface. All deformed basin-plain deposits with a deep basal shear surface are induced by the weight and impact of a slope-adjacent mass-flow wedge. The frontally emergent – most mobile – basin-plain deformation is triggered by mass flows originating from onshore mass movements (i.e. debris flows, rock slides and avalanches) propagating into the fjord. This basin-plain deformation results in vertical seafloor offsets of up to 20 m. Therefore it might be even more important for far-field tsunami propagation than the impact of the onshore mass movements on the sea surface. In the depressions created by the basin-plain deformation, megaturbidites occur, while more distally, sandy density-flow deposits cover the seafloor. The data also indicates that these density flows propagate slower than the basin-plain deformation. Based on correlations with the two main eruptions of the Hudson Volcano, we hypothesize that during the Holocene three to four similar events have struck the fjord. The small variability of the structural characteristics of the Liquiñe-Ofqui Fault Zone in the northern Patagonian fjordland and historical seismic swarms in this area make us conclude that similar hazards should be taken into account for most of the fjords in this region.

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1. Introduction

Offshore mass movements have repeatedly proven their value in paleoseismological research; both in marine and lacustrine settings (Dawson, 1999; Strasser et al., 2006; Moernaut et al., 2007; Tripinas et al., 2008; Moernaut et al., 2009; Masson et al., 2011). They have been successfully correlated to historical earthquakes and were used to identify prehistorical events. Especially in fjords or bays, mass movements can be catastrophic (Syvitski and Schaefer, 1990; Bae et al., 2000; Urgeles et al., 2002) and potentially induce tsunamis (Tocher and Miller, 1999; Ward, 2001; L’Heureux et al., 2011; Suleimani et al., 2011). These, in turn, represent great hazards for coastal villages and harbors, which in many cases provide the best access to remote inland areas.

The northern Chilean fjordland is a tectonically active region. Not only is it crossed from north to south by the Liquiñe-Ofqui Fault Zone, an active dextral strike-slip structural lineament, it is also located along the subduction zone that produced the largest earthquake ever instrumentally recorded (i.e. the 1960 Great Chilean earthquake (Fig. 1A; Melnick et al., 2009). Nevertheless, except for the study of St-Onge et al. (2012), in which mass-wasting events in Reloncaví fjord were linked to megathrust earthquakes, and Araya Vergara (2011), who detected buried submarine mass-transport deposits in Aysén fjord, no studies have had the objective to map and date submarine mass-movement deposits or basin-plain deformation in the Chilean fjordland. However, the 2007 Aysén fluid-driven seismic swarm (Fig. 1B; Legrand et al., 2011) and 21 April 2007 Mw 6.2 main shock clearly highlight that research on the characteristics and the recurrence of such catastrophic events is essential for establishing reliable hazard assessments in the region. After the extensive onshore mass movement triggered by the 2007 main earthquake, studies were performed on these mass movements and the tsunami induced by them (Fig. 1C; Naranjo et al., 2009; Sepúlveda and Serey, 2009; Sepúlveda et al., 2010). Unfortunately, older (i.e. historical and prehistorical) onshore mass movements are often hard to detect and to date (Jibson, 1996), complicating any study of such older events and of possible recurrence rates between events. Ward and Day (2010) propose that the 1958 seismically induced onshore rockslide in Lituya Bay (Alaska) triggered a submarine landslide. Introducing this submarine slide into their tsunami model was necessary to fit the high run-ups – derived from eyewitness and trimline accounts – at the more remote locations.

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Bozzano et al. (2009) and Watt et al. (2012) have successfully shown that seismic-reflection profiling and/or multibeam bathymetry can be utilized to detect deposits of onshore landslides propagating into the sea or a lake. They also show that these onshore landslides have the potential to trigger offshore mass movements, thereby demonstrating that on- and offshore mass movements can be related. Such offshore mass-movement deposits are more efficiently preserved in the sedimentary record, and can potentially be dated much easier.

Finally, Schnellmann et al. (2005) proposed a model in which mass movements trigger basin-plain deformation with fold and thrust belts, accompanied by a lake tsunami or seiche.

In this study, we link offshore basin-plain deformation to onshore mass movements triggered by the $M_w$ 6.2 2007 earthquake, using seismic-reflection profiling and multibeam bathymetry data. This strategy allows identifying and mapping not only the recent, but also older (buried) basin-plain deformation. Furthermore, short sediment cores are used for ground-truthing the superficial deposits and in this way better understand the seismic signature of older deposits.

Fig. 1. Setting of Aysén fjord: A: shaded relief map of the inner fjord area with indication of the various types of onshore mass movements that were triggered by the 2007 $M_w$ 6.2 earthquake (Sepúlveda et al., 2010). In the fjord, location of reflection-seismic (sparker) profiles and short sediment cores are shown. B: region around Aysén fjord with indication of the Liquiñe–Ofqui Fault Zone (after Cembrano et al., 2000; Melnick et al., 2009), volcanoes, tephra and pumice fallout isopachs (after Naranjo and Stern, 1998, 2004; MEN = Mentolat, MAC = Macá, H = Hudson) and the region of the 2007 seismic swarm and main shock (after Mora et al., 2010; Legrand et al., 2011) and C: northern Patagonian fjordland with Liquiñe–Ofqui Fault Zone and the epicenter of the 1927 earthquake, which – similarly to the 2007 event – triggered onshore mass-movements and tsunamis in the Chilean fjordland (Naranjo et al., 2009).
A hypothesis, based on seismic stratigraphy, for the age of these older deposits is presented.

2. Setting

2.1. Geographical setting

Aysén fjord is one of the many fjords in the Chilean part of Patagonia, located between Puerto Montt in the north and Cape Horn in the south. It is carved in the Andean Cordillera, which reaches altitudes of approximately 2000 m asl (meters above sea level) in this region. At the eastern extremity of the fjord lies Puerto Chacabuco, the largest harbor of the Aysén District and through Puerto Aysén the main access to Coyhaique, the district capital (Fig. 1).

2.2. Geological, tectonic and volcanic setting

The geology of this area is dominated by the Patagonian Batholith. This granitic to dioritic batholith outcrops from the Llaima volcano (38°40′S) in the north to Cape Horn (56°00′S) in the south. Therefore, most of the catchment of Aysén fjord consists of this granitic to dioritic bedrock, locally interrupted by some Quaternary volcanic centers with associated basaltic to rhyolitic volcanic rocks that cover the batholith. The eastern part of the Aysén River catchment consists mainly of Mesozoic volcanic rocks (basaltic–rhyolitic composition) with some intercalated sedimentary sequences (Sernageomin, 2003).

A second important geological feature in the study area is the Liquiñe-Ofqui Fault Zone (LOFZ; Fig. 1). The LOFZ is a 1000 km long two overlapping, NNE-striking, master faults, connected by a series of NE-striking, right-lateral en echelon faults (Legrand et al., 2011). In Aysén fjord, one of these en echelon faults joins the easternmost master fault (Fig. 1A and B).

During the first months of 2007 a seismic swarm with more than 7000 recorded earthquakes affected the region around Aysén fjord. The series of seismic events started on 22 January 2007 and reached a maximum on 21 April 2007 with an Mw 6.2 earthquake. The epicenter of this Mw 6.2 earthquake was located in the fjord (Fig. 1B) and the northern part of the Patagonian fjordland (i.e. Aysén to Chiloé). On 21 April 2007 several short gravity cores were collected in the inner part of the fjord, three of which, taken from key locations in the fjord, were used in this study (AY11, AY12 and AY20; Fig. 1A). The cores were taken at the eastern end of the fjord (Fig. 1A and B). The series of seismic events struck by a seismic swarm in May–June 2008; during this swarm magnitudes reached Mw 5.3 (Servicio Sismológico, 2012).

Several major Pleistocene–Holocene volcanoes are present close to Aysén fjord. The Macá and Cay volcanoes, 20 and 30 km north of the fjord, respectively, and the Hudson Volcano, 50 km south of the fjord, are the three southernmost volcanoes of the Andean Southern Volcanic Zone (SVZ) (Fig. 1B). The Hudson Volcano (45°54′S, 72°58′W, 1905 m asl) is an active volcano, with a 10 km wide, ice-filled caldera. The largest known eruptions of this basaltic–dacitic volcano are prehistorical and date from ~3600 14C yr BP (H2) and ~7430 14C yr BP (~8260 cal yr BP) (H1), the latter probably having formed the caldera (Naranjo and Stern, 1998; Stern and Weller, 2012). The basaltic to dacitic Macá Volcano (45°06′S, 73°10′W, 2960 m asl) is the highest volcano in the region, but has only one known eruption: in 410 ± 50 AD. The Cay Volcano (45°3′33″S, 72°59′3″W, 2090 m asl) is a basaltic to dacitic stratovolcano with no historical records of eruptions (Naranjo and Stern, 2004).

3. Materials and methods

The data used in this study (multibeam bathymetry, high-resolution reflection-seismic profiles and short sediment cores) were acquired in December 2009 on board of RV Don Este.

3.1. Multibeam bathymetry

The bathymetry of the inner part of the fjord (i.e. east of Tortuga Point; Figs. 1 and 2) was mapped using an ELAC Seabeam 1050 multibeam sonar. The two 50 kHz transducer arrays were installed on a pole at the bow of RV Don Este, and operated with 120° swath, transmitting and receiving 108 beams of 3° by 3° beam angle. In order to correct for roll, pitch and heave an IXSEA Octans motion sensor was installed right above the transducers, to minimize errors. The sound-velocity profile was made using the pressure, temperature and salinity data from an IDROMAR IP039D CTD. Continuous CTD measurements at the surface were made with a Valeport CTD. The data was first cleaned with HDpedit and tide information was added in HDPpost. Final data cleaning and interpretation were done using the IVS 3D Fledermaus™ software. Grid resolutions are depth-dependent and range from 4 m for the shallow areas to 20 m for the deeper areas.

3.2. Seismic-reflection data

High-resolution seismic-reflection data was acquired using RCMG’s in-house designed Centipede multi-electrode sparker source. The sparkers, operated at 300–400 J, produces a broad-spectrum seismic signal, with a mean frequency of ~1.3 kHz. A single-channel streamer with 10 hydrophones and a total group length of 2.7 m, was used as a receiver. Data was acquired at an average survey speed of 4 knots. Seismic and GPS data were digitally recorded and converted to SEG-Y format with the IXSEA™ Delph Seismic Acquisition system. Interpretation of the data was done using The Kingdom Suite™ software. Up to 35 seismic profiles with a total length of 133 km were acquired in the inner part of the fjord (Fig. 1A). Thickness of sediment packages was calculated based on a sound velocity of 1500 m/s.

3.3. Sediment cores

Several short gravity cores were collected in the inner part of the fjord, three of which, taken from key locations in the fjord, were used in this study (AY11, AY12 and AY20; Fig. 1A). The cores were taken with a Swiss gravity corer and have a length ranging between 50 and 130 cm. Magnetic susceptibility (MS) of the sediment was measured every 2.5 mm with a Bartington MS2E point sensor (3.5 mm resolution). Grain size was measured with a Malvern Mastersizer 2000 at 1–5 mm spatial resolution, depending on the sediment, without pretreatment.
4. Results

4.1. Basin morphology

The combination of multibeam bathymetry and seismic stratigraphy provides insights into both the morphology of the bottom and the sedimentary infill of the fjord. The primary morphology of the fjord consists of steep slopes, presenting a continuation of the onshore mountainous morphology, with sharp lower slope breaks (Fig. 2). The basin plain is almost flat and has a depth of 150 m bsl (meter below sea level) in the inner fjord (i.e. at the slope break of the Aysén River delta), gradually increasing towards the outer fjord, reaching a depth of 217 m bsl east of the Cuervo Ridge (Fig. 2). West of the Cuervo Ridge, the fjord bottom is flat and has a depth of 343 m bsl. This primary morphology, however, is interrupted by several discontinuities.

4.2. Superficial deformed basin-plain deposits (DBPDs)

The flat topography of the fjord floor is interrupted by large depositions with a characteristic chaotic to transparent facies and a positive topography. Based on their resemblance with deposits described by e.g. Schnellmann et al. (2005) and Moernaut and De Batist (2011), these deposits are interpreted as mass-movement deposits and/or deformed basin-plain deposits (DBPDs). Since a fjord can be regarded as a system intermediate between lacustrine and marine (i.e. relatively small basins, high sedimentation rate, but marine setting), we adopt the terminology used by Moernaut and De Batist (2011) to describe the morphology of the DBPDs and the characteristics of their basal shear surface (BSS). However, we use deformation distance instead of run-out distance, since this term better explains that in this specific case of DBPDs, it is the deformation that propagates, rather than a mass movement that slides and slumps. The DBPDs can be subdivided into four groups: i) long, frontally confined DBPDs with a shallow BSS (Type 1), ii) frontally emergent DBPDs with a shallow BSS (Type 2), iii) frontally confined DBPDs with a deep BSS (Type 3) an iv) frontally emergent DBPDs with a deep BSS (Type 4; Fig. 3).

Type 1 DBPDs are frontally confined DBPDs with a deformation distance of up to 3 km. The BSS is not everywhere at the same stratigraphic level, but corresponds in most cases to the first high-amplitude reflection which occurs at a depth of <10 ms bsf. These DBPDs develop distally of Type 2 DBPDs on slopes of 0.5°–1.5° and have a slope angle at the toe of 0.3° (Figs. 3, 4 and 5).

Type 2 DBPDs are frontally emergent DBPDs with a deformation distance of <1 km. The BSS is not everywhere at the same stratigraphic level, but corresponds in most cases to the first high-amplitude reflection which occurs at a depth of <10 ms bsf. Proximally, most of these DBPDs have a hummocky morphology with poor seismic penetration. Some of these DBPDs occur where onshore rock slides and small debris flows propagated into the fjord (Fig. 3).

The Type 3 DBPD is a frontally confined DBPD with a deformation distance of ~2 km. The used BSS (i.e. SL-D) is the second high-amplitude reflection and occurs at a depth of ~30 ms (~22.5 m). The proximal fjord slope gradients range between 0 and 45°. On the northwestern slopes, a headscarp is present at a depth of ~60–80 m. At the foot of the slopes, sidecarps merge with the frontal thrust of the DBPD and a wedge with chaotic facies occurs. The proximal zone of the DBPD forms a depression with a similar depth as the original sea floor and consists of a chaotic-transparent seismic facies, which is partly covered by a ponding, transparent unit. The distal part of the DBPD has a positive morphology and rises gradually towards the borders with maximum heights of 13 m above the undisturbed sea floor. In this part thrust faults cut through the still recognizable original stratigraphy (Figs. 3, 6 and 7A).

Type 4 DBPDs are frontally emergent DBPDs with a deformation distance of >1 km. The DBPDs have three BSSs, which are connected by two step-ups. These DBPDs occur where onshore rock avalanches or debris flows propagated into the fjord. No head- or sidecarps are observed in the usually steep fjord slopes. Proximally, a wedge with chaotic facies and sometimes hummocky upper surface occurs. The rest of the
4 TYPES OF BASIN-PLAIN DEFORMATION

Fig. 3. Schematic illustration of the 4 types of basin-plain deformation encountered in Aysén fjord.

- The outer thrust belt has a positive topography, although lower than the one of the inner thrust belt. The BSS that is used in this part of the deposit is SL-F (Figs. 6 and 7B).
- The unconfined part of the deposit emerges distally or laterally from the outer thrust belt (Fig. 6).

4.2.1. Acantilada Bay

Acatilada Bay is characterized by the Aysén River delta, situated at the eastern extremity of the fjord. The subaqueous topset of the delta is only a few meters deep and therefore not covered by the bathymetry map. The delta slopes extend from 20 to 120 m bsl and have a mean slope angle of 6°–8° (Fig. 4). The seismic penetration in the delta front sediments is limited to only a few meters and the seismic facies is chaotic. The north-facing slopes are interrupted by multiple small channels, which are up to 5 m deep and extend straight towards the gently SW-inclined basin floor (1.5° to 0.22° from northeast to southwest, respectively). These channels converge into a larger channel that originates at the northern foot of the delta slope, circumventing the delta to the west and then following a SW course adjacent to Partida Island, ending up at Mano Point. This channel has a width and depth of 100–250 m and 0–4 m, respectively (Figs. 4 and 5). Along the Aysén River delta and the western and northern shores of Acatilada Bay, multiple Type 2 DBPDs and one Type 4 DBPD occur. The most characteristic for this area, however, are several Type 1 DBPDs, originating at the distal parts of the other DBPDs and covering a large part of Acatilada Bay (Figs. 4 and 5). Adjacent to these Type 1 DBPDs, also channels occur with widths and depths of 50–200 m and 0–3 m, respectively. These channels locally cut through the Type 1 DBPDs and the western channel only cuts through the upper sediments without disturbing deeper stratigraphic levels; they merge south of Mano Point (Figs. 4 and 5).

4.2.2. Cola Point–Mentirosa Island area

In the area comprised between Mano Point in the east and the Quitralco Fault in the west, multiple DBPDs are present. The depth of the flat basin floor ranges between 180 and 210 m bsl. Along the shores, several Type 2 DBPDs occur, but the most prominent DBPDs are of Type 3 (Cola Point NW) or Type 4 (Cola Point SE, Mentirosa Island W, Mentirosa Island SE, Fernández Creek and Aguas Calientes; Figs. 6 and 7). All these DBPDs are named after onshore geographic features where also subaerial debris flows and rock slides and avalanches occurred during the 2007 earthquake (Figs. 4). In the Cola Point SE DBPD, both the inner and the outer thrust belts are incised by a NNE-SSW striking channel (Fig. 6).

4.2.3. Cuervo Ridge and surroundings

In the southern part of the northwest-facing slopes of the Cuervo Ridge a Type 2 DBPD occurs. Also distally from Playa Blanca, a Type 2 DBPD with a hummocky morphology is present. Lacking seismic-reflection data in this area prevents us from making further observations concerning that DBPD (Fig. 8).

4.2.4. Ground-truthing

Southwest of Mentirosa Island, a transparent, ponding deposit can be observed on the seismic-reflection profiles. The deposit is filling a morphological depression delimited by the confined part of the DBPDs in the north, west and south, and by the gently rising undisturbed basin floor in the east (Figs. 6, 7 and 9). Core AY11 was taken in the central part of this deposit (Figs. 6 and 7C), and penetrates a 106 cm thick, normally graded unit (Fig. 10). The lower 18 cm consist of normally graded coarse to very fine sands, followed by 12 cm of normally graded fine sands to silts and finally 76 cm of homogenous silts. The upper 2 cm of the core consist of silts similar to the sediments underneath this deposit (Fig. 10).

Core AY20, which was taken south of the Fernández Creek DBPD (Fig. 6), sampled a 12.5 cm thick sandy deposit. The lower 5 cm of
this deposit consists of predominantly normally graded medium to fine sands with intercalated and silty laminae. The upper part is characterized by normally graded fine to very fine sands. On top of this unit, a 1.2 cm thick gray-brown, poorly-sorted silt layer occurs. Finally, the top of the core is formed by a 1.8 cm thick, brown, poorly sorted–sorted silt layer that is similar in composition to the bioturbated (mottled, but with horizontal stratification) sediment below the sandy unit and to the homogenous silts in AY11 (Fig. 10).

Fig. 4. Shaded relief map with occurrence of onshore mass movements (after Sepúlveda et al., 2010) and bathymetry of the Acantilada Bay area (location: Fig. 2); A: bathymetry and location of sparker profiles; B: grayscale bathymetry with interpretation of superficial deformed basin-plain deposits and channels. SL: stratigraphic level; BSS: basal shear surface.

Fig. 5. Sparker profiles crossing Acantilada Bay (location: Fig. 4A) and showing SL-F and 2007 deformed basin-plain deposits and channels; A: proximal; B: distal. Uninterpreted profiles: see SI Fig. 1.
Core AY12 was taken in the outer thrust belt of the Cola Point SE DBPD (Fig. 6) and clearly shows the deformations in the sediments. The lower 45 cm of the core (i.e. 6–51 cm) are characterized by folded and faulted silts, in which the original stratification (AY20) can still be recognized. The upper limit of this unit is irregular and filled by a 3–5 cm thick normally graded unit of coarse sand to coarse...
silt. The upper 2.5 cm of the core consist of homogenous silt similar to the top of AY20 (Fig. 10).

4.3. Buried DBPDs

The upper part of the sedimentary infill of the fjord consists of draping sediments with a continuous, stratified facies (Fig. 9). Within these draping sediments, seismic-reflection data reveal several buried DBPDs (Fig. 9), some of which share the same stratigraphic level (SL), all with high reflection amplitudes (i.e. SL-A, SL-C, SL-D and SL-F). Three stratigraphic levels (i.e. SL-A, SL-C and SL-F) comprise all except one of the buried DBPDs. The geographical extent of DBPD-occurrence at the three levels is similar as for the superficial DBPDs: from the Cuervo Ridge in the west to the Aysén River delta in the east (Fig. 11).

The largest DBPD occurs at SL-A, in the area between the Condor River mouth and Mano Point. This deposit is accompanied by a large transparent, ponding deposit, covering the northeastern part of the DBPD and the area comprised by this deposit and the Aysén River delta. Other, smaller, DBPDs occur at this level in the northwestern part of Acantilada Bay, in the Mentirosa Island–Aguas Calientes area and south of the Cuervo River delta (Figs. 9 and 11).

Fig. 8. Shaded relief map with occurrence of onshore mass movements (after Sepúlveda et al., 2010) and bathymetry of the Cuervo Ridge area (location: Fig. 2); A: bathymetry and location of sparker profiles; B: grayscale bathymetry with interpretation of superficial deformed basin-plain deposits. SL: stratigraphic level; BSS: basal shear surface.

Fig. 9. Sparker profile showing the upper part of the Aysén fjord sedimentary infill and the different stratigraphic levels (SL), deformed basin-plain deposits, transparent lens-shaped megaturbidites, the Condor River fan, the associated ponding unit and sediment cores. H1 = ~7430 14C yrs BP Hudson eruption; H2 = ~3600 14C yrs BP Hudson eruption, the H1 probably formed the caldera (Naranjo and Stern, 1998; Stern and Weller, 2012; location: Fig. 1). Uninterpreted profiles: see SI Fig. 3.
At SL-C, DBPDs are present in the same areas as for SL-A, with the largest one south of Mano Point. Probably some more, shallow excavating DBPDs occur in Acantilada Bay, but the resolution of the seismic-reflection data does not allow mapping these deposits (Fig. 11).

Fig. 10. Three gravity cores taken in Aysén fjord showing the folded and faulted upper part of the DBPDs (AY12 and AY11), distal density-flow deposits (AY20 and AY12) and a megaturbidite (AY11). The dotted correlation lines between AY20 and AY12 show how the original stratification can still be recognized in the deformed sediments. For each core: left: unprocessed core photograph; middle: photograph processed with histogram equalization; right (only for AY20 and AY11): grain-size distribution (location of the cores: Figs. 1 and 6A); and the magnetic susceptibility of the sediments on a logarithmic scale (yellow line).

At SL-D, only a single DBPD was found west of the Condor River mouth. This stratigraphic level only has a high-amplitude reflection in the area close to this deposit (Figs. 9 and 11).

At SL-F, DBPDs were found in Acantilada Bay, Mentirosa Island and in the Playa Blanca–Cola Point area. East of the DBPD at Playa Blanca, a...
transparent lens-shaped deposit with a maximum thickness of 3 ms, occurs. This semi-ponding deposit is plastered against the southwestern slopes (Figs. 9 and 11).

Furthermore, due to the incomplete coverage of the seismic-reflection grid and to blanking below younger DBPDs, an underrepresentation of buried DBPDs is likely.

4.4. Condor River fan

At the Condor River mouth, at SL-B, a large fan-like feature is intercalated between the sediments of the draping unit (Figs. 4 and 9). The buried fan is composed of a chaotic to transparent facies, and is deposited on an area covering a quarter of a circle with a radius of approximately 2.5 km. The maximum thickness of this buried fan with a slope of 4–10° is estimated to be 130–200 ms (~100–150 m). The corresponding volume of this fan can be calculated as the volume of a quarter cone and would be ~0.16–0.25 km³.

Towards the basin, the Condor River fan merges into a distally thinning, ponding unit. SL-B and SL-B’ represent bottom and top of this unit, respectively. This unit is characterized by a basin-focusing morphology, without any draping, and high-amplitude reflections, especially close to the fan (Fig. 9). It has a thickness of ~20 ms (~15 m) near the fan and gradually thins out to a thickness of ~10 ms (~7.5 m) east of the Cuervo Ridge (Fig. 9). The thickness of this deposit gradually increases up to ~30 ms (~22.5 m) at the base of the Aysén River delta. The estimated total volume of this unit is in the order of ~0.9 km³.

SL-E is a prominent reflector with high reflection amplitudes, which are strongest close to the Condor River fan (Fig. 9).

5. Discussion

5.1. 2007 basin-plain deformation

The superficial DBPDs in Aysén fjord, mapped with seismic-reflection profiling and multibeam bathymetry, are assigned to the 2007 Mw 6.2 earthquake. The three short sediment cores (i.e. on top of one of the DBPDs, in the transparent unit and in a distal area) all show an evidently very recent event deposit with a maximum of 2 cm of draping sediment covering the top of the deposit (Fig. 10). In AY12 and AY20, these sandy deposits are interpreted as density-flow deposits (Mulder and Alexander, 2001). Moreover, the largest of these superficial DBPDs (i.e. CPSE, MIW, MISE, FC and AC in Fig. 6 and the DBPDs along the northern and western shores of Acantilada Bay; Fig. 4) can be spatially linked to subaerial debris flows and rock avalanches that entered the fjord after the 2007 Mw 6.2 earthquake. These large deposits form a jigsaw puzzle with the smaller deposits rather than covering
them (Fig. 6), indicating that all the superficial deposits mapped using the seismic-reflection data and the multibeam bathymetry are triggered by a single event: the 2007 Mw 6.2 earthquake. The transparent ponding units that fill the depressions created by the DBPDs (Figs. 7 and 9) and are cored by AY11 (Fig. 10), are interpreted as megaturbidites (Bouma, 1987).

All the DBPDs along the northwestern shores of Acatintalada Bay occur where onshore debris flows, rock slides and avalanches entered the fjord. In the east, where the basin slopes are very steep (i.e. 15–35°), the deposits are likely mainly composed of material from these onshore mass movements, as suggested by the hummocky morphology. In the western part of Acatintalada Bay, where the slopes are less steep (i.e. 3–20°) and therefore possibly sediment-bearing, the onshore mass movements likely triggered offshore DBPDs, which have a higher mobility. The DBPDs at the base of the west-facing delta slopes and the absence of large channels on these slopes, suggest that these slopes have indeed failed during the 2007 earthquake.

A major part of Acatintalada Bay basin plain is covered by Type 1 basin-plain deformation (Fig. 4). In the east, this basin-plain deformation developed adjacent to a preexisting channel, which channelized density flows that originated at the delta and the northeastern shores of Acatintalada Bay. In the west, a channel incises the Type 1 DBPD and continues to Mano Point, where it merges with the eastern channel (Fig. 4). This channel was most likely formed by density flows postdating the DBPD, indicating that the deformation of the basin-plain deposits propagated faster than the density flows.

All the large DBPDs (i.e. Type 3 and 4) occur at locations where large subaerial debris-flows and rock slides/avalanches entered the fjord, and vice versa (Fig. 6). Only the Type 3 DBPD at Cola Point is caused by a mass flow of slope sediments, in turn triggered by an onshore rock avalanche that entered the fjord. For all the large and most of the Type 2 DBPDs, the model proposed by Schnellmann et al. (2005), with some minor adaptions, can be used to explain their formation and geometry. This model assumes that the increasing load at the slope break marking the transition between the fjord slope and basin floor induces gravity spreading of the basin-plain sediment leading to propagating overthrusting and sediment deformation, with the latter causing a completely transparent facies. The proximal mass-flow wedge (i.e. the load) was recognized on top of some of our DBPDs (Fig. 7).

The structure of the Type 3 DBPD at Cola Point can be explained by this model. However, the proximal deformation that is characteristic for the Cola Point DBPD does not occur in the deposits studied by Schnellmann et al. (2005). A possible explanation for this difference is the larger height drop of the mass flow in Aysén fjord (>150 m versus ~100 m). This difference can cause higher impact velocities, resulting in a higher energy transfer, resulting in the stronger basin-plain deformation.

In the Type 4 DBPDs, not only the central depression is deeper, but also proximal deformation penetrates deeper in the fjord sediments (Fig. 7). A succession of step-ups and a distal unconfined part is typical for these deposits. This enhanced displacement is probably caused by a larger height drop and steeper slopes (Moernaut and De Batist, 2011). All the Type 4 DBPDs occur where onshore rock avalanches and debris flows entered the fjord, resulting in an offshore height drop of >195 m. Moreover, the higher density of the rocks compared to the slope sediments will increase the inertia and energy transferred during impact with the basin-plain sediments. Types 3 and 4 of basin-plain deformation in Aysén fjord could thus be considered to be the next two steps in the model of Schnellmann et al. (2005), in which energy transfer during impact becomes increasingly important for the deformation. Alternatively, the different morphology of the Aysén DBPs compared to the DBPDs in the Swiss peri-alpine lakes (Schnellmann et al., 2005) could also be related to different mechanical properties of the basin-plain sediments in the two settings. Also, apart from the mechanical properties of the muddy sediments, the occurrence of the high-amplitude reflectors SL-A and SL-C (interpreted in Section 5.2 as density-flow deposits of prehistoric 2007-like events) can also have had an influence on the deformation. These sandy density-flow deposits are used as a BSS by the 2007 DBPDs and thus have the right mechanical properties to act as a shear surface.

We can therefore conclude that the prehistoric event-deposits influenced the formation of the 2007 event deposits and that without the occurrence of these density-flow deposits, the deformation in 2007 would have penetrated less deep into the basin-plain sediments and would therefore have caused less vertical deformation of the seafloor.

The large height difference between the inner depressions and the thrust belts (up to 20 m) suggests that the Types 3 and 4 basin-plain deformation could have the potential to be tsunamiogenic. Similarly as in Lituya Bay (Alaska), such a tsunami sensu stricto can potentially propagate further than an impact wave produced by onshore mass movements impacting on the fjord water (Ward and Day, 2010). The tsunamiogenic potential is therefore also influenced by the mechanical properties of the basin plain sediments and the occurrence of sand layers such as density-flow deposits. However, all types of deformation and resulting waves should be modeled to confirm these hypotheses.

Similarly to Acatintalada Bay, the data also clearly shows how the basin-plain deformation of the Type 4 deposits propagates faster than the overriding density flow. The channel that incises the inner thrust belt of the southeastern Cola Point frontally emergent DBPD, is aligned with the on- and offshore mass-flow pathways (Fig. 6). Hence, we interpret that the density flow broke through the freshly deformed and up-thrusted basin-plain sediments. This delay can be logically explained by the rigid basin-plain sediment, in which all deformation occurs quasi-instantaneously through grain-to-grain contact, compared to the density flow, which has to travel the entire distance. On top of the deformed basin-plain sediments, these density flows only leave a thin (i.e. a few centimeters) normally graded gravel to silt layer, while, distally, these density flows are preserved as ~10 cm thick sandy deposits (core AY20; Fig. 10).

5.2. Prehistoric events and chronology

Also the older DBPDs coincide with a distal density-flow deposit. The older DBPDs distally correspond to basin-wide high-amplitude reflections in the case of multiple DBPDs on one stratigraphic level, originating from different slope segments of the fjord (i.e. SL-A, SL-C and SL-F), or to local high amplitudes in the case of a single DBPD on one stratigraphic level (i.e. SL-D). Some of these stratigraphic levels can be tied to the MD-07 3117 core, a Calypso core which was taken from the RV Marion Dufresne, in the context of the PACHIDERME project, 2 months prior to the Mw 6.2 earthquake. In this ~21 m long core, sandy layers at depths of ~5 m, ~17 m and ~20 m are interpreted as tephra layers, and at a depth of ~10 m a 50 cm thick tephra and pumice layer occurs (Kissel and The shipboard scientific party, 2007). The tephras at ~5 m and ~20 m depth correlate with SL-F and SL-D, respectively (Fig. 9). Based on our seismic-stratigraphic mapping and interpretation of the DBPDs in the fjord, we suggest that these high-amplitude reflections represent distal density-flow deposits, similar to the 2007 density-flow deposit (i.e. ~10 cm thick sandy layers in AY20, at approximately the same location; Fig. 6A), containing reworked tephra. The pumice-terephra layer correlates with SL-E (Fig. 9).

Based on its intercalated, local nature, we interpret the Condor River fan and its coeval ponding unit to be deposited in a relatively short time period (Fig. 9) and tentatively attribute it to the ~8260 cal yr BP H4 Hudson eruption. The dimensions at sea level of the Condor River fan (i.e. ~1.1 km²) are of the same magnitude as the fan deposited in Chaitén Bay (Fig. 1C) after the 2008 Chaitén eruption (i.e. ~1.3 km²) presented by Lara (2009) (Fig. 8), indicating that it is possible for such a fan to be deposited during a single event. The H4 eruption was the largest Holocene eruption of the Hudson Volcano and the characteristics of the related tephras are consistent with a caldera formation during...
this eruption (Naranjo and Stern, 1998). The Condor River is draining some of the southern slopes of the Hudson Caldera (Fig. 1), hence, caldera formation could produce pyroclastic flows potentially reaching the fjord, and/or provide the Condor River with large amounts of volcaniclastic sediment. The thickening of the ponding unit towards the Aysén River delta is consistent with the fact that also the Aysén River drains a part of the Hudson Volcano (Fig. 1). However, studying volcanic outcrops in the Condor River valley is necessary to confirm this hypothesis.

Correlating the Condor River fan with the ~8260 cal yr BP H1 Hudson eruption, results in an average sedimentation rate of about 0.3 cm/yr at the location of the MD-07 3117 Calypso core. This sedimentation rate is very similar to the sedimentation rates of 0.19–0.30 cm/yr that Salamanka and Jara (2003) determined in the upper 25 cm of sediment on two locations in the inner Aysén fjord. Hence, these sedimentation rates support our correlation and by assuming a constant sedimentation (in TWT ms) – and in the absence of any direct age control on these deposits – we make a simple, back-of-the-envelope assessment of the ages of the prehistoric eruptions and earthquakes. The resulting age for the pumice-tephra layer at SL-E – which is best developed close to the Condor River fan – would be ~3170 yr BP and can be tentatively attributed to the 3600 ±6C yr BP Hudson eruption. This H2 pumice-tephra layer was deposited in large areas south of Aysén fjord (Naranjo and Stern, 1998), and is the only known pumice deposit that close south of the fjord (Naranjo and Stern, 1998, 2004; Stern, 2008; Fig. 1B). The ages of SL-A, SL-C, SL-D and SL-F would be ~9160 yr BP, ~7160 yr BP, ~5830 yr BP and ~1500 yr BP, respectively. On stratigraphic levels SL-A, SL-C and SL-F basin-wide mass-movements occurred on a scale comparable to that of the 2007 event (Fig. 11). We infer that these mass-movement events were triggered by 2007-like earthquakes along the LOFZ, with epicenters very close to Aysén fjord. The SL-D DBPD was probably formed during a smaller earthquake in the fjord or a similar earthquake, but with an epicenter further away from the fjord. Hence, three to four catastrophic events comparable in effect to the 2007 earthquake and tsunami have struck Aysén fjord during the Holocene.

In the studied inner part of the fjord we do not see signs of major DBPDs that could be attributed to the 1927 seismic swarm and associated M 7.1 earthquake. The estimated epicenter of this major earthquake is located about 100 km to the north of the fjord, probably along one of the northernmost en echelon faults connecting the two overlapping master faults of the LOFZ. By using the formulas of Keefe (1984) and Rodriguez et al. (1999) to extrapolate the maximum epicentral distances for onshore disrupted landslides and flows obtained by Sepúlveda et al. (2010) for the 2007 earthquake, we find maximum epicentral distances of ~90 km and ~50–70 km for disrupted landslides and flows, respectively. Hence, it could be expected that no major subaerial landsliding affected Aysén fjord.

5.3. Activity along the Liquine-Ofqui Fault Zone (LOFZ)

2007-like events can occur regularly in the Patagonian fjordland. Scarce historical and instrumental records indicate that seismic swarms, both with and without a large main shock, are common along the southern part of the LOFZ (i.e. south of Puerto Montt). The 1927, 2005, 2007 and 2008 seismic swarms in the Patagonian fjords (Lange et al., 2008; Naranjo et al., 2009; Servicio Sismológico, 2012) demonstrate that these swarms, which are often accompanied by large main shocks, are common along the LOFZ. The low recurrence rate of on average 2500–3000 yr in Aysén fjord shows that at a specific location along the LOFZ, these catastrophic events are scarce. We expect, however, that similar events occur relatively frequently along the LOFZ in the Patagonian fjords between Reloncavi fjord in the north (41.5° S) and San Rafael Lake in the south (46.75° S), and should be taken into account for hazard assessments. North of Reloncavi fjord, the nature of the LOFZ becomes more segmented and displacement decreases (especially north of 40° S, dying out at 38° S) (Rosenau et al., 2006), reducing the risk of large earthquakes in this area.

6. Conclusions

This study gives an overview of the impact of the 2007 seismic swarm and Mw 6.2 main shock on the sedimentary record of the fjord, and provides a hypothesis on the recurrence of such events. The main conclusions are listed hereafter:

1) The onshore mass movements that were triggered by the 2007 Mw 6.2 earthquake and propagated into the fjord, all triggered deformation of basin-plain sediments. Multiple small offshore mass movements have been triggered directly by the earthquake.

2) The largest offshore deformed basin-plain deposits were produced by the impact and load of the onshore mass movements, resulting in unique morphologies of the basin-plain deformation, characterized by proximal depressions and high inner and outer thrust belts.

3) Some of these DBPDs might have been more important for far-field tsunami propagation than the impact of the onshore mass movements themselves.

4) Density flows originating from the Acanthalada Bay and Cola Point mass flows, formed or enlarged channels, occasionally cutting through the deformed basin plain, indicating that the latter propagate faster.

5) The data shows that three to four similar events have struck the fjord during the Holocene. A similar frequency pattern should be taken into account for hazard assessments in the other fjords between 41.5° and 46.5° S.

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