Glacial and tectonic control on fjord morphology and sediment deposition in the Magellan region (53°S), Chile

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A B S T R A C T
In the Patagonian Andes erosion by temperate Pleistocene glaciers has produced a deeply incised fjord system in which glacial and non-glacial sediments were deposited since the Late Glacial glacier retreat. So far, fjord bathymetry and structures in the sediment infill were widely unexplored. Here we report the results of an investigation of morphology and sediment characteristics of a 250 km long fjord transect across the southermost Andes (53°S), using multibeam and parametric echosounder data, and sediment cores. Subaquatic morphology reveals continuity of on-land tectonic lineaments mapped using field and remote sensing data. Our results indicate that glacial erosion and fjord orientation are strongly controlled by three major strike-slip fault zones. Furthermore, erosion is partly controlled by older and/or reactivated fracture zones as well as by differential resistance of the basement units to denudation. Basement morphology is regionally superimposed by Late Glacial and Holocene subaquatic moraines, which are associated to known glacial advances. The moraines preferentially occur on basement highs, which constrained the glacier flows. This suggests that the extent of glacier advances was also controlled by basement morphology. Subaquatic mass flows, fluid vent sites as well as distinct Late Glacial and Holocene sediment infills have furthermore modified fjord bathymetry. In the western fjord system close to the Strait of Magellan subaquatic terraces occur in 20 to 30 m water depth, providing an important tag for proglacial lake level during the Late Glacial.

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1. Introduction

The morphology of the previously glaciated southermost Andes (52°–54°S, Fig. 1A) is characterised by an up to 2000 m deep fjord system, mainly formed by Pleistocene glacial erosion. Orientation of these fjords is more or less controlled by a neotectonic strike-slip fault system (Glasser and Ghiglione, 2009; Waldmann et al., 2011), related to the oblique convergence of the Antarctic and South American plates (Diraison et al., 1996; Ramos, 1999; Lodolo et al., 2003; Costa et al., 2006; Oliver and Malumian, 2008; Ghiglione et al., 2010). However, only a very crude bathymetry existed for most of the fjords, and very little information has been published regarding subaquatic moraine systems and other sediments (Kilian et al., 2007a).

We have investigated the bathymetry as well as sediment types and fjord orientations. Present day glaciers originate from the 200 km² large Gran Campo Nevado ice cap (GCN; Fig. 1C), located at the ice divide of the southermost Andes at 53°S (Schneider et al., 2007). The investigated fjords are aligned along a 250 km long E–W fjord transect that crosses the main lithological units or morphostructural provinces of the Andean mountain belt (trending N–S; Fig. 1B). We focus on four areas to the west and to the east of the GCN, covering different morphostructural provinces: (1) fjords west of the GCN within the Patagonian Batholith, (2) the Gajardo Channel east of the GCN incised into a Jurassic–Cretaceous metamorphic basement, (3) the western Seno Skyring located in the Magallanes fold-and-thrust belt and (4) the eastern Seno Skyring and Seno Otway in the Magallanes foreland basin. For the surveys we used parametric echosounding and high-resolution multibeam bathymetry. Additionally, twenty sediment cores were recovered from the fjord sediments.

2. Regional setting

The major orogeny of the southermost Patagonian Andes started in the Middle to Late Jurassic, forming an island arc and associated backarc basin. Extensional faulting and subsidence lead to the evolution of the Rocos Verde marginal basin, with an infill of Jurassic volcanioclastics and Cretaceous turbidite sequences. In the late Cretaceous and the Palaeogene crustal shortening and thickening caused the closure of the Rocos Verde basin, and the formation of the Magallanes fold-and-thrust belt (Fildani et al., 2003; Menichetti et al., 2008; Oliver and Malumian, 2008; Ghiglione et al., 2010; Romans et al., 2010). The Magallanes foreland basin developed east of the Andean orogen, filled with an up to 7 km thick sediment sequence (Ramos, 1989; Ghiglione et al., 2009). From Neogene onwards the region was affected by NW–SE sinistral wrench tectonics (Diraison et al., 1996). The left-lateral...
Magallanes–Fagnano fault zone (MFZ; Fig. 2) represents the major transform fault system (Lodolo et al., 2003). Some minor NNE trending right-lateral strike-slip faults can be interpreted as associated R’ Riedel shear zones (Maffione et al., 2010). Elongated pull-apart basins, horst and graben structures, pressure ridges, uplifted and faulted flower structures and drag folds are due to releasing or restraining bends in the fault system (Lodolo et al., 2003).

During the Last Glacial Maximum (LGM) an 80 km wide ice field extended from the GCN to the west and east (Kilian et al., 2007a). It formed part of a large (2500 km N–S) Patagonian Icefield (Glasser and
Ghiglione, 2009). Extent of the ice cover is well documented in the area of Seno Skyring and Magellan Strait (Benn and Clapperton, 2000; McCulloch et al., 2005a; Kilian et al., 2007a). Between 18 and 15 kyr BP a major ice recession led to the formation of extended proglacial lakes, like the Seno Skyring, Seno Otway and the fjord systems of the Magellan Strait (Glasser et al., 2008; Rabassa, 2008). Proglacial lakes were successively transformed to fjords due to the global sea level rise (Kilian et al., 2007b). The orientation of most fjords follows the structural geological lineaments (Glasser and Ghiglione, 2009; Waldmann et al., 2011). The Seno Otway, Seno Skyring and the Eastern Strait of Magellan cross-cut the fold-and-thrust belt following extensional structures (Diraison et al., 2000) oriented perpendicular to the Andean orogen. The fjords and bays carved out by the glaciers (McCulloch et al., 2005a) act as depositional environments for glacial, glaciomarine, and glaciolacustrine sediments.

Five different deglacial ice retreat stages (A to E) are evident in the Magellan region (Clapperton et al., 1995; McCulloch et al., 2005b; Rabassa, 2008). In the western fjord systems the related moraines are in general subaqueous, like that of stage D in Seno Skyring and stage E within the Gajardo Channel (Kilian et al., 2007a). In the western Magellan region the glaciers retreated to present-day extents between 12.6 and 11.8 kyr BP (e.g. Kilian et al., 2007a). Pollen records from the GCN area indicate an evolved Magellanic rainforest after 11.2 kyr (Lamy et al., 2010).

Seen in a global context, the Chilean fjord system belongs to the Southern Fjord Belt (Syvitski et al., 1987). The present-day GCN ice cap reaches 1700 m (a.s.l.) and feeds temperate valley and tidewater glaciers (Warren and Aniya, 1999; Benn and Clapperton, 2000; Rabassa et al., 2008; Koppes et al., 2009). The Chilean tidewater glaciers extend into the warmest settings, in terms of air and water temperatures (Dowdeswell et al., 1998). In temperate fjords, like in the southernmost Chilean fjord system, the postglacial deposition is dominated by high organic input (Syvitski and Shaw, 1995; Howe et al., 2010), which can be as well measured in lake sediment cores fed by streams coming from the GCN region (Breuer et al., 2013). The sediment from the GCN-derived glaciers enters the fjords mainly as hypopycnal flows (Fig. 3B). Several Holocene tephras layers from volcanoes of the Austral Volcanic Zone (AVZ; Stern and Kilian, 1996) form important age markers in sediment cores of the investigated fjord area (Kilian et al., 2007a), with the most prominent one originating from the 4.15 kyr BP Mt. Burney (Figs. 1, 2) eruption (Kilian et al., 2003; Stern, 2008).

Fig. 2. False colour Landsat ETM+ satellite image from 2000 of the Magellan fjord region between 52°S to 53.3°S with morphological lineaments representing the three main strike-slip fault directions (MFZ, SCFZ, GFZ) and their associated trendlines. The orientation of all faults and trendlines are plotted in a rose diagram. The fold axes and thrust fault derived from the official geological map (SERNAGEOMIN, 2003; modified). Epicentres of relevant earthquakes in this region.

After USGS Earthquake Hazards Program.
3. Material and methods

3.1. Core data

Stratigraphies and sedimentological properties deduced from piston and gravity cores were used to improve the interpretation of sediment layers seen in the seismic profiles. The sediment cores include: CG-1 (Baeza, 2005; Kilian et al., 2007a); Sky-1 (Kilian et al., 2007ab; Lamy et al., 2010). For cores BA-1, LO-1, SG-1 and MD07-3126 a chronology was developed based on tephra chronology, 14C dating from macroplant remnants (Poznan Radiocarbon Laboratory; Table 1) and 210Pb dating (Table 2; Appleby and Oldfield, 1992). All 14C ages were converted to calibrated ages before present (BP 1950) using the Calib 6.0.1 software and SHCal04 curve for the Southern Hemisphere (Stuiver and Reimer, 1993; McCormac et al., 2004; Stuiver et al., 2011). All depicted ages are means of 1-sigma values.

3.2. Acoustic data

Along the fjord transect 1600 km of echosounder profiles and around 500 km of multibeam data were recorded, processed and interpreted. We used the Parametric Echosounder System (SES 2000) from Innomar (Wunderlich and Müller, 2003), which provides high and variable low (4–12 kHz) frequency signals at water depths of 0.5 to 800 m with a vertical resolution of less than 5 cm. The pulse length ranges between 0.08 and 1 ms, and the beam width is 1.8° (Wunderlich and Müller, 2003). The fjord bathymetry was calculated from the high frequency signal. To obtain optimal visualization of sediments, we tested low frequency signals between 4 and 12 kHz to obtain an optimal trade-off of penetration and resolution. This allowed a penetration of up to 55 m. However, in some areas gas in the sediment hampered penetration, resulting in transparent layers (Fader, 1997). The ELAC-Nautic Seabeam 1180 multibeam sonar system was used for high-resolution seafloor bathymetry in selected areas. The system is designed for water depths up to 600 m. However, deeper basins can be imaged with a reduced swath width. Using a frequency of 180 kHz, it is able to record 126 single depth values with a swath of max. 153°.

3.3. Data editing and processing

The echosounder data were edited using the post processing software ESE (version 2.9) from Innomar. A sound velocity correction was applied by using several CTD records taken in the fjord system. Acoustic velocities of water-rich sediments may vary by ±2% compared to marine water (Hamilton, 1979). In our case, comparison of the depth of the prominent 4.15 kyr BP tephra reflector in the echosounder profiles and related sediment cores indicated sound velocities around 1500 m/s. The most important layers were digitised and surfaces of them were calculated by using different geostatistical interpolation models. The ‘Ordinary Kriging’ method was used for restricted areas, because it relies on the spatial correlation structure of the data to determine the weighted values. For the extended Seno Skyring with less coverage of seismic lines the ‘Inverse Distance Weighted’ method (IDW) was used, which weighted the nearby data points. The multibeam data were processed at the IFM-GEOMAR (Kiel, Germany). This included checking and editing of the navigation, removing of erroneous measurements and artefacts by automatic and manual cleaning, conversion of recorded travel times to positions and depths by complete ray-tracing through the water column taking into account measured sound velocity, depth profiles, and calculation of digital terrain models.

3.4. Structural and lineament analysis

Lineament analysis includes terrestrial erosional incisions, fjord orientations, as well as distinct, straight lines in the morphology which may represent fractures and tectonic faults. We mainly interpreted GeoCover Landsat satellite images, recorded in 2000 by NASA, with a pixel size of 14.25 m. The Landsat 7 ETM + comprises the bands 7, 4 and 2 and the mosaic is projected in UTM coordinates in the World Geodetic System WGS 84. For the structural overview map (Fig. 2) only the main linear surface features were taken into account, like fjord valleys, coastlines and fracture zones. Straight continuous landforms are related to strike-slip faults (Glasser and Ghiglione, 2009). For the detailed maps of the Seno Glaciar (Fig. 3A) and the Gajardo Channel (Fig. 1B) the same principles were applied. Smaller faults, fractures or trendlines were analysed on the ice-polished bedrock. Ground-truthing of most faults was done during field work in the Seno Glaciar area and in the Gajardo area in 2001. We also included information, like fold axis and thrust faults, from the geological map 1:100,000 (SERNAGEOMIN, 2003) and major strike-slip faults determined by Glasser and Ghiglione (2009). A total of 642 lineaments were mapped and plotted in three different rose diagrams (GEORient; version 9.5.0).

4. Results

4.1. On-land morphology

Remote sensing data and field mapping show three sets of first-order fault zones (Fig. 2): (1) The sinistral strike-slip Magallanes Fault Zone (MFZ) trending 136°. Northward of the Strait of Magellan this fault fabric is less pronounced (Fig. 3); (2) A sinistral fault system oriented NNW (165°) along the continental margin northward until at least 50’S. In the study area it appears primarily to the east of the GCN and parallel to the Swett Channel. Thus, we name it the Swett Channel Fracture Zone (SCFZ; Fig. 3); (3) A NNE-trending fault system (44°) around the GCN, running parallel to the northern Gajardo Channel (Fig. 6). It is termed, therefore, the Gajardo Fault Zone (GFZ). These particularly dextral strike-slip faults are associated to the MFZ (Maffione et al., 2010).

4.2. Fjords to the west of the GCN (Swett Channel and adjacent bays)

The multibeam-based bathymetry indicates up to 350 m water depth in the Swett Channel and other parallel fjords further to the east. Most subaquatic structures and canyons are related to the SCFZ (Fig. 3A). The geology of the Swett Channel region is characterised by alternating zones of granitoid rocks, mylonitic rocks and orthogneisses (Fig. 3A). The mylonites (Menichetti et al., 2008) crop out along two NNW-trending zones parallel to the SCFZ direction, and their easy erodibility leads to the formation of canyons and fjords. The granitoids, resistant to erosion and denudation (Breuer et al., 2013), form a distinct NNW-trending ridge east of the SCFZ between the mylonitic zones. This ridge formed a morphological barrier capable of damming the GCN glaciers and hampering their westward flow. Only narrow passages cross this morphological barrier. The bathymetric high at the southern end of the Swett Channel (Fig. 3A, B) separating it from up to 340 m deep basin (Quenca Luca) is part of the granitoid rock ridges. One of the major palaeo-glacier streams coming from the Glacier Noroeste migrated first southward through the Swett Channel, before crossing into the Luca Basin, and flowing westward into the Seno Glaciar basin (red arrows in Fig. 3B).

Fig. 3 A) Georeferenciated ortho-image (Schneider et al., 2007) combined with high resolution bathymetry. Morphologically analysed fault and trendlines, core locations and profiles are shown. Moraines, deltas and fluvial sediments are distinguished within the Quaternary sediments. The rose diagram shows the orientation of faults and trendlines in this region. B) False colour Landsat ETM + satellite image from 2000 and black-and-white ortho-image (Schneider et al., 2007). The bathymetry is derived from a geostatistical interpolation (Ordinary Kriging). The white dots show the localities of the SG-1 and MD07-3126 core. The red dashed lines indicate the former glacier flow directions. The suspension load and transport direction can be seen clearly in the ortho-image.
In the south-eastern sector of the Bahía de los Glaciares (Fig. 3A) three glaciers enter this fjord area at present. Fig. 4A shows a cross section between the glacier tongues of Bahía de los Glaciares along the ancient glacier flow until Seno Glaciar (A–A'; Fig. 3A). The profile crosses several well-preserved subaquatic moraine systems, faults, basement highs and intervening small sediment basins (Fig. 4A). The glaciers could only erode two narrow and shallow passages into the above described granitoid ridge. The passages constricted the glacial streams and thus hindered the westward ow during the Neoglacial advances, so that moraine formation occurred particularly in these narrow channels (Fig. 3A; multibeam-derived bathymetry map). In the echosounder profiles the moraines show typical cram-and-tail structures with chaotic internal reflectors (Fig. 4A). In small depressions on the top younger, stratified sediments with a draped channel ﬁll pattern were deposited. A very pronounced left-lateral strike-slip fault (SCFZ) is characterised by a deeper fjord incision.

Sedimentation in the Swett Bay, west of the GCN, is affected by the inﬂow of up to 500 m³/s fresh water from two large rivers. One of them drains Lago Muñoz Gamero with its 810 km² catchment. The other one is a meltwater river originating from the proglacial lake of Glacier Noroeste (Fig. 3A). The latter river carries a high suspension load into the fjord, causing south-westward hypopycnal ow within the low-salinity surface water layer over several kilometres along the Swett Channel (Fig. 3A, B) and into the adjacent bays (e.g. Bahía Lobo). The wind-controlled distribution of the suspended load can be seen in several satellite images and aerial photographs (Fig. 3B), and was also observed during our cruises at different seasons of the year. The on-land deltaic fan continues a few hundred metres into the fjord, as can be seen in the high-resolution bathymetry (Fig. 3A). Related to the above described glacial clay plumes, relatively high sedimentation rates of 1.0 (BA-1 sediment core) to 1.2 mm/yr (SG-1 sediment core) were found in basins within the western GCN fjord system. The echosounder data also show up to 30 m thick sediment deposits (Fig. 4B) in basins between subaquatic ridges and/or moraines, and in shallow coastal bays or platforms.

Fig. 4B shows an echosounder proﬁle across the Swett Bay into the Swett Channel. Three seismic sequences can be differentiated. The basement is characterised by chaotic internal reﬂectors and shows a pronounced morphology. The basin of Swett Bay is separated from a basin in the Swett Channel by a granitoid ridge. Both basins are ﬁlled with >20 m of well stratified sediments, interpreted as ice distal deposits. This indicates sedimentation under more dynamic conditions with current activity. The uppermost sequence has a draped channel ﬁll pattern with subparallel internal reﬂectors documenting suspension sedimentation.

Bahía Arévalo (Fig. 4C) is a shallow (5–6 m water depth) small bay located 2.1 km NNW of the mouth of the river coming from Glacier Noroeste (Fig. 3A). It was investigated by echosounding and by drilling (core BA-1). The 8.8 m long piston core documented a continuous organic matter bearing clayey sedimentation throughout the past 5 kyr. The age constraints include the 4.15 kyr Mt. Burneys tephra at 4.2 m core depth and a calibrated 14C age of 1.7 kyr BP obtained from macroplant remnants found at 1.6 m core depth (Table 1), indicating relatively constant sedimentation rates of 1 mm/yr. Below 5 m core depth (approx. 5 kyr BP), a transition towards a coarser grained clastic ﬂuvial sedimentation is recorded, probably documenting a less elevated coastline which enabled the formation of a shallow lake. This suggests coastline uplift by at least 10 m after ~5 kyr BP (Kilian et al., 2011). The echosounder proﬁle in Fig. 4C is next to the BA-1 core location. The reﬂectors of the uppermost sequence are even parallel and have a draped pattern. The lowermost lake facies with variable and partly coarse sediment shows clinoformal reﬂector patterns (Fig. 4C), typical for prograding deltaic sediment systems (Stoker et al., 1997). At the top of the lowermost deltaic sequence there is an erosional unconformity. The upper deltaic facies II filled the depressions and levelled the subaquatic morphology. The deltaic sequences have an Early to Middle Holocene age. The tephra layer is conformal on the deltaic sequence, and is overlain by an organic-rich, marine sediment. At present a river enters the bay and satellite images show a recent subaerial delta.

Bahía Lobo is a larger fjord bay (3 × 0.8 km) 2.5 km south of the glacial river coming from the Glacier Noroeste (Fig. 3A). In the aerial images parts of the hypopycnal ow that directly enter the bay can be seen. The whole 8.17 m long LO-1 piston core retrieved at 83 m water depth consists of silty clays with low amounts of organic material. A calibrated radiocarbon age of 0.73 kyr BP was obtained from macroplant remnants at 761 cm core depth (Table 1) indicating very high sedimentation rates of 10 mm/yr during the past centuries.

4.2.1. Seno Glaciar, Seno Icy and Bahía Beaufort

During the last Glacial Maximum there was an extended south-westward ice stream from the GCN along the Swett Channel into Seno Glaciar, Bahía Beaufort and Seno Icy, where it merged with the large westward-migrating ice stream of the western Strait of Magellan north of Tamar Island (Figs. 2, 3A, B). The bathymetry of Seno Glaciar does not show pronounced ridges. The fjord includes a >5 km wide sediment basin with a water depth of 520 m, which was ﬁlled by >50 m thick sediments during the Late Glacial and Holocene. Sediment thickness is constrained by extrapolating seismic reﬂectors from the basin margin towards its centre where a thick IRD layer hampers acoustic penetration. Two short sediment cores MD07-3126 and SG-1 could also not penetrate this IRD layer. The 0.48 m long QASQ box core (MD07-3126 obtained with RV Marion Dufresne in 2007) and the 0.56 m long gravity core (SG-1 obtained with RV Gran Campo II) were retrieved from the basin. Six 210Pb ages give an age frame for the last ~0.55 kyr (Table 2). The base of the silty to clayey sediment sequence contains ice-rafted debris. 210Pb ages for the past 75 years indicate average sedimentation rates of 1.2 mm/yr and suggest that the massive IRD

### Table 1

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<th>Error (±yr)</th>
<th>Calibrated age (yr BP)</th>
<th>Sed-rate (cm/kyr)</th>
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### Table 2

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</table>

Average sedimentation rate: 0.12.
layer was deposited during an early maximum of the Little Ice Age between ~0.6 and 0.5 kyr BP.

Further downstream, the ancient ice flow turned to SSW, running 15 km in Bahia Beaufort and Seno Icy parallel to the GFZ (Fig. 1C). The crude bathymetry of Seno Icy indicates several ridges perpendicular to the ancient ice flow, which is related to cross-cutting fractures of the MFZ (Figs. 2, 3). A detailed bathymetry was obtained across this up to 600 m deep U-shaped fjord-valley (Figs. 1C, 5). In the north of this bathymetrical map several subaquatic canyons are visible, parallel to the MFZ and GFZ directions. They are more pronounced on the shallow fjord margins and may have been formed by fluvial erosion after the regional ice retreat after 17 kyr BP (Kilian et al., 2007a), when the coastline was much lower than at present, due to >100 m lower global sea level. The bathymetric data also show subaquatic terraces in 20 to 30 m water depth on both sides of the valley (Fig. 5). Their formation requires relatively constant shoreline elevations during the Late Glacial and Early Holocene. This observation can be only explained, if the local glacial rebound was similar to the global sea-level rise for some millennia, especially between 14 and 9 kyr BP. This is compatible with previous regional coast levels estimated by Kilian et al. (2007a) and Lamy et al. (2010). No clearly preserved subaquatic moraine remnants have been detected bathymetrically along the ancient glacier stream between Swett Channel and the Seno Icy area, indicating a relatively fast and continuous glacier retreat in this sector.

4.3. Gajardo Channel

The 55 km long Gajardo Channel (Figs. 1, 6) was formed by glacier streams originating mainly from the GCN and from mountains on western Riesco Island (Kilian et al., 2007a). Along the ancient glacier flow direction...
the NNE-trending northern section of the Gajardo Channel stretches over 30 km from the Angostura de Témanos at the GCN towards the Seno Skyring (Fig. 6).

Multibeam bathymetry and selected echosounder cross sections (Fig. 6) show a U-shaped glacial valley with a total relief of up to 2000 m. The present day water depth is up to 680 m. On-land mapping in the north-eastern sector around the GCN, interpretation of aerial photos, as well as false colour satellite images show NNE-trending active dextral strike-slip faults parallel to the GFZ direction. Faults parallel to the MFZ (NW direction) have been also mapped in the field. The rose diagram shows a third minor E-W trending fault direction shaping the Bahamondes area and Chandler Island. Parallel to this on-land fault zone several lineaments could be detected, which cross the northern Gajardo Channel diagonally (Fig. 6).

The high-resolution bathymetry shows areas with irregular sub-aquatic ridges. Partly these are basement highs of less erodible rocks, which can be related to terrestrial features, and partly they are sub-aquatic moraines. The latter are seismically characterised by chaotic to contorted reflectors (Kilian et al., 2007a). They were probably formed at around 14 kyr BP during Late Glacial glacier readvance Stage E (Fig. 6; McCulloch et al., 2005b; Kilian et al., 2007a). In front of the north-eastern entrance of the Gajardo Channel within the Euston Channel, an older subaquatic moraine system was likely being formed between 17.5 and 15 kyr BP during Stage D (Kilian et al., 2007a). To the north of the Angostura Témanos and in the southward MFZ-oriented deeply incised fjord, moraines of the Little Ice Age have been previously detected (Koch and Kilian, 2005; Kilian et al., 2007a). The high-resolution multibeam map shows several curved terminal moraines (Fig. 6), which are partly related to on-land moraine remnants.

The Gajardo Channel includes several basins (Fig. 6), which are separated by subaqueous moraine systems and/or bedrock ridges. The main Holocene sedimentation took place in these basins and in small depressions located on top of the morainal ridges. The deepest basin of the fjord (680 m water depth) is located ca. 6 km southwest of the entrance of the Gajardo Channel. A 1.73 m long gravity core (CG-1) was taken in the 250 m deep basin for which a fjord-parallel echosounder profile is shown in Fig. 6. It is characterised by parallel reflectors with a draped channel fill pattern. The 4.15 kyr Mt. Burney tephra layer in 5.3 m depth in the core produces a distinct reflector within the Holocene sequence. The underlying ice distal sediments (Fig. 6) show ponded fill patterns with sub-parallel internal reflectors. The CG-1 core shows an overall clayey to silty sediment composition (Kilian et al., 2007a).

A Late Holocene to recent debris fan was detected in the bathymetric map south of Chandler Island (Fig. 6). The associated steep (48°) hill slope on land shows clear evidence of a recent debris flow. The sub-aquatic deposits cover an area of 1.0 × 1.2 km and have a volume of around 80 million m³. Two major debris flow events are distinguished, indicated by the dashed lines in Fig. 6; B–B’. The most recent abrasion surface is not visible in the Landsat TM 345 false colour image of 1986, but was observed in the field in 1999. It is possible that the mass flow is related to the M = 5.0 earthquake which occurred on 31.08.1996 around 38 km to the SSW, or the 4.7 magnitude earthquake which occurred on 19.11.1988 around 32 km to the SSW (Fig. 1C).

4.4. Western Seno Skyring

At present, the former proglacial lake Seno Skyring (Figs. 1C, 2) represents an estuarine fjord system with restricted connection to the Strait of Magellan through the Gajardo Channel in the west, and Seno Otway, via the Fitz Roy Channel in the east. The E–W oriented fjord has a length of around 80 km and is up to 15 km wide. In the west Seno Skyring passes into the Euston Channel, where its deepest part of the basin (660 m water depth) is located. Further to the east, sediment basins
are developed in-between perpendicular oriented ridges, which belong to the Magallanes fold-and-thrust belt.

A bathymetric map of the Seno Skyring (Fig. 1C) was generated by geostatistical interpolation from the echosounder profiles shown in Fig. 7. In some areas, e.g. the Euston Channel, not enough data were available for a good interpolation.

In general, the water depth decreases towards east. The 16 km² sized Escarpada basin with a maximum water depth of 440 m is located in the central sector of Seno Skyring. Echosounder profiles show a more than 35 m thick sediment fill with parallel reflectors, which belong to Holocene and Late Glacial ice distal sequences. Escarpada Island has a U-shape (Fig. 1C), due to selective erosion of a shallowly northward plunging fold of the Cretaceous Cerro Toro Formation (Romans et al., 2011; KS1m in Fig. 7), which was more resistant to glacial erosion. Sediment core ES-1 was retrieved at 78 m water depth in the 2.5 km wide bay within U-shaped coastline of Escarpada Island. The core shows the 4.15 kyr BP Mt. Burney layer at 3.2 m sediment depth, indicating sedimentation rates of 0.78 mm/yr (Baeza, 2005).

Eastward of Escarpada Island the basin of the former Skyring glacier incised the Magellanic fold-and-thrust belt (Figs. 1, 2, 7), where Upper Cretaceous and Lower Tertiary rock units of the former Magellan Basin crop out (SERNAGEOMIN, 2003). Erosionally resistant sedimentary rock layers form N-S-trending ridges, which can be easily identified in satellite images, are subaquatically preserved. Some of them define small islands or peninsulas. The most important subaquatic ridge extends from Altamirana Bay southward to Riesco Island across the

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**Fig. 6.** Landsat ETM+ image combined high resolution bathymetric map of the Gajardo Channel. The two main fault and trendline directions are plotted in the rose diagram. Several moraine systems, basement highs, basins and one pronounced debris flow could be detected within the fjord bathymetry. The profiles (1–3, upper left corner) show the typical U-shaped form. The echosound profile [A–A’] shows the typical sediment succession of the isolated basins within the channel. The profile [B–B’] shows the suspected original seafloor, and the dashed lines indicating the assumed border between the two events.
whole Seno Skyring (Fig. 7). The ridge stands up to 210 m above the glacially abraded basement, and subdivides Seno Skyring into two main basins.

Fig. 8 shows a coast-parallel echosounder E–W profile across Seno Skyring, including the above described subaquatic ridge. Four different sediment sequences can be distinguished. A lowermost Sequence I of ice proximal deposits (5–10 m thickness) with subparallel internal reflectors and partly chaotic reflectors documents coarse sedimentation (IRD), above the folded basement. Above this layer, a non-stratified Sequence II occurs in isolated lenses often related to fault-like structures (Fig. 8). These may represent local subaquatic debris flows either triggered by fault activity, or by reworking of IRD deposits of the Late Glacial (Kilian et al., 2007a). We prefer an interpretation of the wedge-shaped acoustically transparent bodies, lenticular in cross-section, as debris flows (Syvitski et al., 1987; Whittington and Niessen, 1997; Gipp, 2003; Howe et al., 2003). The well-stratified Sequence III further up-section is interpreted as ice distal deposits (4–10 m thickness) above a basal unconformity. The uppermost sequence consists of well stratified, organic-rich Holocene sediments (2–8 m thickness) and includes the 4.15 kyr Mt. Burney tephra layer. The transition to the underlying ice distal sediments is mostly concordant. The sediment thickness of each sediment sequence reflects basement morphology, with higher sediment thickness in basins. Faulting leads to block rotations of the basement (Fig. 8). The sediment cover is only partly faulted, with low displacements. Most of these faults cause a sediment deformation and the minor faults die out upwards. Analysis of all seismic profiles across Seno Skyring indicates that Holocene sediment thickness decreases towards the east.

The shallow (<80 m water depth) eastern part of Seno Skyring is the area of the ancient glacier tongue. Core Sky-1 was recovered in this sector from 76 m water depth. Four tephra layers with well-known ages (Kilian et al., 2007a,b; Lamy et al., 2010) were identified. IRD layers in this core indicate ice retreat as early as 17 kyr BP. The sedimentation rates of 0.19 mm/yr throughout the Holocene and late Glacial are much lower than further west.

Fig. 9 shows a representative 6.5 km long echosounder profile (A–A’) from the eastern part of the Seno Skyring. Local reflectors are disturbed, indicating gas within the sediments that could be derived from deeper sources (Fig. 9A). This layer blanking mainly affects the deepest sequence of the ice proximal deposits. We suspect that the gas ascends along faults from the subjacent hydrocarbon-bearing Cretaceous and Lower Tertiary rock units of the Magallanes Basin. The lowermost ice proximal deposits have an uneven reflector signature; in gas-influenced areas the reflectors are disrupted. The top of the sequence has a subdued relief, but the transition to the ice distal sediments is concordant. In general, the ice distal facies is characterised by even sub-parallel reflectors, in some areas (from 2.0 to 2.5 km profile length) with a wavy parallel pattern. Occasional synsedimentary faults cause offsets in the range of some decimetres up to 1 m. This indicates that some of the faults were active during the Holocene. In a few cases an erosional unconformity separates the Holocene from the underlying ice distal sediments. One pronounced reflector within the Holocene represents the 4.15 kyr BP Mt. Burney tephra layer. In the SKY-E1 core this tephra layer occurs in 1.18 m core depth with a thickness of 22 cm. Within the seismic profile (Fig. 9A) the tephra layer appears at around 1.5 m depth.

A gas-related feature appears at 5 km of the profile and extends towards the east from there. It is characterised by dome-shaped structures, which deformed and uplifted the whole overlying sedimentary
sequence (Fig. 9A). Westerly-induced currents in the upper 70 m of the water column (Kilian et al., 2007b) eroded the uplifted sediments by bottom transport processes. These eroded sediments were deposited further east, above the normally deposited Holocene sediments with a clear unconformity (Fig. 9A). The reflectors of the redeposited sediments have an oblique tangential pattern typical for prograding systems. In contrary to this, in debris flows no internal reflectors can be seen (Fig. 9A). In the eastern part of the Seno Skyring the thickness of the Holocene sequence is on average 10 m in basin areas, and thins to 2 m at morphological highs.

4.5. Seno Otway

Seno Otway (Fig. 9B) is an 87 km long and 30 km wide estuarine bay and former proglacial lake, connected to the Magellan Strait in the west by the Jeronimo Channel. Most of the measured seismic data were obtained in the north-eastern and central parts (Fig. 1).

Where the Magellanic fold-and-thrust belt intersects Seno Otway, pockmarks occur frequently. Strata below the pockmarks are strongly disturbed, likely due to gas expulsion. Here we concentrate our description on the easternmost sector of Seno Otway (Fig. 1) to the northeast of the major pockmark field. In the echosounder profile (Fig. 9) four sequences are evident. The oldest one is weakly reflective and chaotic, probably representing the bedrock or older glacial deposits (Glacial I; Fig. 9; B–B’). The transition to the ice proximal deposits is clearly visible as well as the basin-like structure. In this sequence (till) the internal reflectors are chaotic and contorted (DaSilva et al., 1997). Above, there is a concordant transition into well-stratified ice distal sediments. The internal reflectors are even parallel with a draped pattern and indicate homogeneous, fine-clastic sedimentation under fairly calm conditions. This Glacial II sequence is secondarily deformed by water-escape structures, which equally affected the underlying Glacial I sediments. The water-escape features occur in the profile as V-shaped structures (Fig. 9B), refilled with sediment. The water-escape features start within the ice proximal sequence (Fig. 9; B–B’), where coarser clastic sediments occur. After Syvitski (1997) these events are typical for ice proximal areas. Water-escape events may have been triggered by periodical earthquakes in this region (Mills, 1983), but the mechanisms are not obvious by analysing the morphologies (Morretti and Ronchi, 2010). Rapid sedimentation can be also a trigger mechanism (Postma, 1983; Syvitski, 1997). It is obvious that the water-escape structures only occur up to the unconformable transition to the uppermost sediment sequence (Fig. 9B). This erosional unconformity cuts the reflectors of the Glacial II sequence (at 5.5 km profile length). The overlying relatively thin (1.5 to 2 m thickness) Holocene sequence has sub-parallel reflectors and levels the surface between the basin and the structural highs of the bedrock (at 3.3 km profile length).

5. Discussion

Cenozoic subduction and associated upper-plate deformation (Ramos, 1999; Diraison et al., 2000; Polonia et al., 2007; Glasser and Chiglione, 2009) have strongly impacted on the basin and mountain topography, and the structure and erodibility of rocks in the southernmost Andes. Our investigation of fjord and on-land lineament orientation also indicate a predominant control of the erosion by a complex framework of at least three major fault zone systems, producing a complex fjord network. In this framework we have attempted to identify other important mechanisms that have modified on-land morphology and bathymetry of the fjord system. These include formation of lateral and/or terminal moraines as well as debris flow deposits. Irregular morphological structures like basin ridges, bathymetric highs, or narrow passages profoundly influence fjord currents and related sediment transport. In the
following chapters we discuss these aspects in the light of all the data presented.

5.1. Tectonic control

A tectonic control on the fjord orientation has been previously proposed by Glasser and Ghiglione (2009), based on the investigation of fjord orientation and fault zones without bathymetrical information. The detailed fjord bathymetry presented here, and the investigated relationship to mapped on-land tectonic lineaments, however, improve the understanding of subaquatic erosion, sediment transport and mass movements.

The rose diagram in Fig. 2 shows the three main directions of the MFZ, SCFZ and GFZ fault systems. From the detailed maps of Swett Channel (Fig. 3A) and Gajardo Channel (Fig. 6) it becomes obvious that the MFZ direction is present in each area, but the SCFZ direction is lacking in Gajardo Channel area and vice versa. In the Swett Channel area zones of mylonitic rocks occur parallel to the SCFZ, probably originating from strike-slip faulting in Late Cretaceous or Tertiary times (Menichetti et al., 2008). These present-day sinistral faults of SCFZ orientation may represent inverted or reactivated older faults.

Beside the SCFZ orientation no clear other preferred direction can be seen in the rose diagram of the Swett Channel (Fig. 3A). When taking a closer look at the satellite image (Fig. 3A) it is obvious that almost every peninsula has its own main trendline direction. Whether this relates to fabric genesis of the different rock units (e.g. Patagonian Batholith, Tobifera Formation) separate in time and space, and a later juxtaposition at terrane boundaries during the Tertiary, is not entirely clear. An improved interpretation would require extensive radiometric dating and structural mapping. The tectonically controlled erosion features are partly superimposed by a Quaternary detritus cover (Fig. 3A), often represented by subaerial moraines or deltas. In some areas, like around Mt. Burney volcano (Fig. 1) the Holocene volcaniclastic deposits mask the glacially shaped fjord morphology. Such postglacial deposits complicate the interpretation of the satellite image.

Fjord orientations are consistently linked to the main tectonic directions, as described above. Rock lithology, however, has a strong local influence on landscape and fjord morphology. The basement of the Swett Channel area is dominated by silicic magmatic rocks of the Patagonian Batholith and the Tobifera Formation (SERNAGEOMIN, 2003), which are relatively resistant to glacial erosion. However, fracturing along fault zones and tectonic lineaments greatly facilitates erosion, leading to a fjord landscape with steep slopes. The bathymetric map shows several subaquatic basement highs, which can be directly related to the subaerial morphology of continuing ridges along strike. These basement highs and the subaquatic moraine systems subdivide the area into uneven-sized basins.

The Gajardo region is mainly affected by dextral strike-slip faults (Fig. 6) and some thrust faults, which belong to the Magallanes fold-and-thrust belt (Fig. 2). The uniform basement of the Gajardo Channel area consists of allochthonous rocks of the Tobifera Formation, and metamorphosed sedimentary rocks of Devonian and Carboniferous age (SERNAGEOMIN, 2003). In contrast to the magmatic rocks of the Patagonian Batholith, metasedimentary rocks are generally more erodible (Fig. 1). Therefore the Gajardo Channel was formed as a uniform U-shape glacial valley with less subaquatic basement highs, but was later bathymetrically modified by subaquatic moraine systems (Fig. 6).

Further to the east, the Seno Skyring is subdivided in two main basins by the N–S trending folds of the Magallanes fold-and-thrust belt (Figs. 1C, 7). These steeply dipping and weathering resistant rocks
form typical ridges in the subaerial landscape (Fig. 7), which continue into the fjords. These structures and active faults may have played a role in creating very irregular coast lines (Figs. 8, 9A) and there is evidence that the Glacial and Holocene sediments in the fjord were affected by very young faulting (Fig. 9A). The locations of gas expulsion structures may also be controlled by faults and their activity.

The whole research area is a tectonically active region, documented by recent earthquakes, with epicentres shown in Fig. 2. Along the SCFZ northwest (Fig. 2) of our research area the epicentres of the three earthquakes of magnitudes 5.3 to 6.6 are located. These events document active wrenched faulting in this area. Other earthquakes of the last decades (Fig. 2) are also located close to the main fault zones, or within the foreland fold-and-thrust belt. Generally, the earthquakes in southernmost South America and Tierra del Fuego are characterised by shallow hypocentres (~40 km) and by intermediate magnitudes (M = 36), with higher magnitudes along MFZ and including the Lago Fagnano area (Lodolo et al., 2003; Waldmann et al., 2011). One of the recent earthquakes in the GCN area may have triggered the large debris flow in the Gajardo Channel.

5.2. Subaquatic moraines

In addition to the basement highs, subaquatic moraine systems are responsible for the development of isolated basins and therefore for the specific sediment distribution and deposition.

The cross section A–A’ through the Bahía de los Glaciares shows Neoglacial moraine ridges which were formed during the most extended Holocene glacier advance at around 2.3 to 2.2 kyr (Arz et al., 2011). These moraines are preferentially located on top of basement highs (Fig. 4A), suggesting that fjord morphology is also controlled by the extent of glacier advances. Post-neoglacial sedimentation is restricted to small depressions with a higher thickness (~11 m) or on top of the moraines with a lower thickness (~3 m).

The stage E moraines in the Gajardo Channel (Fig. 6) are not located on basement highs. They were formed on the nearly flat basement floor, like the other Neoglacial moraine systems in the Gajardo Channel. Within the irregular moraine belts sedimentation is mainly restricted to small basins and depressions in the moraine system. The Late Glacial and the Neoglacial moraines show the same sedimentation characteristics. The breached character of the Holocene sediments and the predominant hypopycnal sediment flow with following suspension sedimentation lead to a uniform sediment deposition. Seismic character of the morainal systems is dominated by chaotic internal reflectors with a lower amplitude and often characterised by strong sea bottom multiples (Carlson, 1989; Howe et al. (2003) described similar features from the Kongsfjorden area, and Sexton et al. (1992) characterised moraines as structureless seismic bodies.

Some small debris or mass flows were detected at the steep slopes of subaquatic moraines. Following Ottesen et al. (2008), Carlson (1989) or Cai et al. (1997) we interpret these as glacialic debris flows. These mass movements affected the older ice distal sediments and partly the Holocene sediments. The older debris flows affecting the ice distal deposits are embedded in the normal sediment succession. Slumping at morainal ridges and at sediments deposited at basement highs and along the steep fjord walls is often caused by an unstable sediment configuration due to rapid sediment accumulation (Solheim and Pfarman, 1985; Syvitski and Shaw, 1995; Stoker et al., 2010). The slumping processes modified the original shape and structure of sedimentation. The slumps and subaqueous gravity flows (Figs. 6B, 8) may be generated by earthquakes (Gipp, 2003; Stoker et al., 2010) or by expulsion of biogenic gas (Fig. 9). The marine sediments are often unconsolidated, and when shaken by large earthquakes, they thought to move en masse (Powell and Molnia, 1989). High sedimentation rates create slope over-steepening, and often lead to gravitational mass transfer (Powell and Molnia, 1989). In southernmost Chile the debris flows are likely triggered by the same factors and the water-saturated soils enhance the subaerial debris flows (Fig. 6B).

5.3. Sediment structures in fjord basins

In general, we distinguish two main glacial units due to their seismic expression: the ice proximal deposits and the ice distal deposits. The heterogeneous composition of the ice proximal sediments often has no clear internal seismic reflection pattern. Ice proximal sediments occur in different thicknesses, but often the base of the units cannot be clearly identified owing to the decreasing quality of the echosounder data with depth. Above a mainly concordant transition the ice distal deposits succeed with a variable thickness. The ice distal sediments often fill depressions or small basins and smooth the submarine morphology (Figs. 6, 9B). In other cases, like the examples of the Skyring (Figs. 8, 9A), the ice distal sediments closely follow the seafloor morphology. Their draped character suggests sedimentation under more tranquil conditions, like suspension sedimentation. The seismic character of the ice distal sediments is given by continuously parallel to sub-parallel, medium to high-amplitude reflections (Cai et al., 1997; Stoker et al., 1997).

In most areas the transition to overlying organic-rich Holocene sediments is conformal, like e.g. Skyring (Figs. 8, 9A) and Gajardo Channel (Fig. 6). In the Seno Otway (Fig. 9B) a hiatus occurs between the ice distal and the Holocene sediments. The hiatus contains an unknown timespan and is associated with an erosional unconformity. Probably the hiatus marks the catastrophic outflow event of the former proglacial lake of the Seno Otway. The Holocene sequence in the entire research area is consistently characterised by even and parallel internal reflectors and a draped pattern, which is typical for suspension sedimentation.

In the larger embayments such as Seno Skyring and Seno Otway, we found a homogeneous and conformable sedimentation of both sequences. The highest sediment thicknesses were found in large basin-like structures like e.g. in the 440 m deep Escarpada and in the 660 m deep Euston basin within the Seno Skyring area (Fig. 1) and in the 680 m deep basin of the north-northeastern Gajardo Channel. In these depressions the reference layer of the 4.15 kyr BP Mt. Burney tephra is overlain by up to 8 m of sediments (Figs. 4B, 6, 8). The Holocene sequence thins out to the basin margins. In most of the smaller basins and shallower coastal bays of the inner fjord region the 4.15 kyr Mt. Burney tephra layer appears at 1–3 m sediment depth. In the narrow fjords and basins in areas with high seafloor relief (basement highs or subaquatic moraines) the sediments are not spatially continuous.

5.4. Comparison with other fjord regions

The Chilean fjord system belongs to the world’s southern fjord belt (Svendsen et al., 1987; Howe et al., 2010). The southernmost Chilean fjords described here are located in an area with a temperate climate. The annual mean temperature is about 5.7 °C at sea level (Schneider et al., 2007) and the annual precipitation rates are about 2000–8000 mm yr⁻¹. The precipitation shows a positive correlation with altitude in the GCN region and a negative correlation with the distance from the Pacific towards the steppe landscapes in the east. Despite the relatively mild climate, the glaciers in the GCN region reach the sea level and form tidewater glaciers. The high precipitation rates lead to high glacier accumulation rates in the GCN region (Möller et al., 2007). In some cases the glacier retreat was stronger in the last years, so that the glaciers do not longer calve into the fjords and the fjords are now under a glacio-fluvial regime. It is important to note, that the salinity of water at the glacier terminus has no influence on the calving rates of the glacier (e.g. Venteris, 1999).

The first order controls on glacimarine sedimentation named by Powell and Molnia (1989) are firstly tectonism, which creates accumulation areas for glaciers and provides sources for glacial debris. From their convergent tectonic setting the fjord regions of Alaska and Chile are correlatable. Tectonism at convergent margins causes mountain
building and under favourable continental orientation to ocean water masses and air masses, glaciations can be predominately accumulation controlled (Warren and Sugden, 1993). This setting allows glaciers in SE Alaska and in Chile at relatively low latitudes to extend down to sea level (Powell and Molnia, 1989). Furthermore, the tectonic movements have a strong influence on the amount of sediment production, and therefore for high sedimentation rates in the accommodation area (Hallet et al., 1996). For example, the tectonic uplift combined with locally fractured and metamorphised rocks creates continuously renewable sources for glacial erosion and debris production (Powell and Molnia, 1989). This is reflected by glacial erosion at the Bahía de los Glaciares and at the Swett Channel (Figs. 3, 4A). The investigated fjord area is an excellent example how the tectonic setting controls the orientation of fjord systems and in particular changes in flow direction through time (Gipp, 2003). Powell and Molnia (1989) describe from the southeast Alaskan margin those active and dormant strike-slip faults and their alignments separate tectonostatigraphic terranes of the margin. Thus major valleys and fjords are aligned with these fault systems and other tectonic elements such as thrust faults. Pfiirman and Solheim (1989) have shown that the tidewater glaciers from Nordaustlandet, Svalbard, follow topographic depressions.

The second first order control mentioned by Powell and Molnia (1989) is the climate. It produces high rates of snow accumulation and/or high rainfall and glacier ablation, which cause high runoff and high sediment discharge. In southeast Alaska precipitation increases with altitude, like in the southeastern Patagonian Andes. High freshwater discharge entering the fjord systems of southeast Alaska and southernmost Chile is due to a combination of high precipitation and glacial meltwater (Powell and Molnia, 1989). Second-order controls are composed of the glaciers themselves and the marine-fluvial interaction (Powell and Molnia, 1989).

The fjords of southernmost Chile are dominated by siliciclastic depositional systems, and thus comparable to the fjords of Alaska (Powell and Molnia, 1989). Sediment supply in the southernmost Chilean fjords is carried by glacial meltwater, like in the fjords of Spitsbergen (Elverøi et al., 1980), of the north-western Barents Sea (Dowdeswell et al., 1998), and of southeast Alaska (Cowen, 1992). In contrast, in the fjords of East Greenland and of the Antarctic Peninsula the high numbers of icebergs play a more significant role in transport and sedimentation (Dowdeswell et al., 1998). In temperate glacier setting like in Chile and in SE Alaska the precipitation is an important factor in glaciomarine sedimentation (Cowen, 1992; Dowdeswell et al., 1998), while in more river-influenced fjords the circulation and sediment transport have stronger ties to the hydrological cycle (Syvitski et al., 1987).

Among the former processes fine-grained meltwater plume sedimentation is dominant in these temperate environments (Fig. 3B). Two types of sedimentation from meltwater plumes can be distinguished: first, slow sedimentation from mainly fine-grained sediment derived from the meltwater plume (Syvitski and Murray, 1981; Pfiirman and Solheim, 1989); and, secondly, faster deposition of the coarse-grained bed load of the glacier streams forming ice proximal deltas (Fig. 3A; Dowdeswell et al., 1998). The ice distal sediments in our research area are mainly characterised by fine grained input, like silts and clay transported by glacial meltwater, a fact that is documented in different sediment cores taken in the research area (Kilian et al., 2007a,b). Such suspension-controlled deposits are reported from the fjord system of Svalbard during the Holocene (Elverøi et al., 1980; Dowdeswell et al., 1998). Beyond the major depocenters of tidewater fronts and deltas, most sedimentation in the Chilean fjords is controlled by the hypopycnal fronts and deltas, most sedimentation in the Chilean fjords is controlled by the hypopycnal fronts and deltas. The ice distal sediments in the research area (Kilian et al., 2007a,b). Such suspension-controlled deposits are reported from the fjord system of Svalbard during the Holocene (Elverøi et al., 1980; Dowdeswell et al., 1998). Beyond the major depocenters of tidewater fronts and deltas, most sedimentation in the Chilean fjords is controlled by the hypopycnal fronts and deltas.

6. Conclusions

The on-land and subaquatic morphology of the investigated area in the southernmost Andes is controlled by three main strike-slip fault zone directions. These are the WNW trending Magallanes Fault Zone, the Swett Channel fault zone (SCFZ), and the right-lateral Gajardo Channel Fracture Zone, which is conjugated to the Magallanes Fault Zone (Maffione et al., 2010).

The basement highs consisting of erosion-resistant rocks partly controlled the extent of glacier advances, and thereby the formation of the subaquatic moraine systems. The Late Glacial and Neoglacial moraine systems prevent a spatially continuous sedimentation, which is mainly restricted to isolated fjord basins. Due to the predominant suspension sediment transport in the upper freshwater-dominated water column with wind-induced currents, the subaquatic geomorphology has no direct influence on the sediment deposition. However, fjord morphology may have controlled locally high salinity bottom currents and their mixing with low-salinity surface water. This could have enhanced flocculation and settling of clay particles. Besides areas with stronger water currents, like in the Strait of Magellan or directly connected fjords, where sediments could be eroded, the basement highs form sediment traps with increased sediment accumulation. Local sediment redistribution processes like debris flows, water escape structures and gas expulsion along faults are important and may have been triggered by seismic events in the southernmost Patagonian Andes.

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