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## Late Holocene current patterns in the northern Patagonian fjords recorded by sediment drifts in Aysén Fjord

Wils Katleen <sup>1,\*</sup>, Wermersche Marlies <sup>1</sup>, Van Rooij David <sup>1</sup>, Lastras Galderic <sup>2</sup>, Lamy Frank <sup>3</sup>,  
Arz Helge W. <sup>4</sup>, Siani Giuseppe <sup>5</sup>, Bertrand Sebastien <sup>1</sup>, Van Daele Maarten <sup>1</sup>

<sup>1</sup> Renard Centre of Marine Geology (RCMG), Department of Geology, Ghent University, Ghent, Belgium

<sup>2</sup> GRC Geociències Marines, Universitat de Barcelona, Barcelona, Spain

<sup>3</sup> Alfred Wegener Institut für Polar- und Meeresforschung, Bremerhaven, Germany

<sup>4</sup> Leibniz Institute for Baltic Sea Research Warnemünde (IOW), Rostock-Warnemünde, Germany

<sup>5</sup> GEOPS, UMR 8148 Université de Paris-Saclay, Orsay, France

\* Corresponding author : Katleen Wils, email address : [katleen.wils@ugent.be](mailto:katleen.wils@ugent.be)

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### Abstract :

Present-day circulation patterns in the southeastern Pacific Ocean are driven by the Antarctic Circumpolar Current, directing subantarctic surface water into the Patagonian fjords since at least the early Holocene. In this way, bottom current patterns in the area are regulated by the regional climate, although the complex bathymetry of the fjords has a significant impact as well. To understand the potential interplay of climate, seafloor topography-and circulation patterns, we study the sedimentary infill of Aysén Fjord (~45°S) and reveal the first active sediment drifts in the region. These allow constraining the present-day circulation patterns in northern Patagonia and show an incoming (southward) as well as returning (northward) flow direction. While the general sedimentary evolution of the fjord (and thus also the sediment drifts) is climate-driven (i.e., it reflects variability in southern westerly wind strength), the onset of drift formation at ~3.7 ka does not seem to have originated from an abrupt change in regional climate. Instead, we propose that a megathrust earthquake described in paleoseismic records in the area could have resulted in subsidence of one (or more) of the many bathymetric highs in the Patagonian fjords, thus contributing to enhanced spilling of subantarctic water into the fjord. This study underscores the importance of multidisciplinary research to understand past and present bottom current circulation patterns and disentangle different possible feedback mechanisms.

### Highlights

► First active contourite deposits recognised in the Patagonian fjords. ► Sediment drift morphology allows bottom current reconstruction. ► Interplay of tectonic and climatic factors on present-day circulation patterns.

**Keywords** : Patagonia, Sediment drift, Bottom current circulation, Climatic variability, Tectonic control

## 30 **1 Introduction**

31 The present-day oceanic surface circulation pattern near the west coast of southern South America is  
32 dominated by the northern margin of the Antarctic Circumpolar Current (ACC), carrying cold, relatively  
33 fresh and nutrient-rich subantarctic water to the Chilean coastline (Strub et al., 1998). The ACC is  
34 considered to be mostly wind-driven, depending on the belt of westerly winds between 45 and 55°S  
35 (Barker and Thomas, 2004). These Southern Westerly Winds (SWW) also largely determine the existing  
36 climate patterns in southern South America, characterized by drastically stronger precipitation on the  
37 western side of the Andes compared to the eastern flank (Garreaud et al., 2013). Consequently,  
38 numerous studies have aimed at reconstructing the late glacial and Holocene climate evolution of the  
39 southern mid-latitudes by revealing variability in the strength and position of the westerlies and/or  
40 ACC through time (e.g., Haddam et al., 2018; Heirman et al., 2012; Kilian and Lamy, 2012; Lamy et al.,  
41 2010; Lamy et al., 2002; Moreno et al., 2010; Romero et al., 2006; Saunders et al., 2018; Verleye and  
42 Louwye, 2010; Villa-Martínez and Moreno, 2007). Many of these studies have focussed on the Chilean  
43 margin and Patagonia, forming the only major landmass intersecting the SWW.

44 A large portion of Chilean Patagonia consists of channels and fjords, forming one of the largest  
45 estuarine systems in the world. These are in direct connection with the open ocean, resulting in a  
46 permanent inflow of subpolar as well as (deeper) subtropical water masses that currently dominate  
47 the oceanic realm in the area (Palma and Silva, 2004). The morphology of these channels is rather

48 complex, shaped through a variety of tectonic and glaciological processes (e.g., Bianchi et al., 2020;  
49 Glasser and Ghiglione, 2009; Syvitski et al., 1987). As a result, circulation patterns in the region are  
50 land-constrained, overprinting the climate-driven oceanic current dynamics. This significantly  
51 complicates the regional feedback mechanisms between bottom currents and climate, of which the  
52 evolution through time remains largely unknown. The study of sediment drifts provides a potential  
53 way forward in unravelling the missing link between these factors, as formation and evolution of such  
54 sedimentary structures is related to bottom current activity (Rebesco et al., 2014) and can result from  
55 climate changes (e.g., Amblas et al., 2006; Gilli et al., 2004; Grützner et al., 2005; Heirman et al., 2012;  
56 Van Daele et al., 2016) as well as tectonic activity (e.g., García et al., 2009; Lobo et al., 2011; Roque et  
57 al., 2012). However, the study of sediment drifts in the Patagonian fjords was, up to now, limited to  
58 identification of an ancient drift in Reloncaví Fjord (~42°S, St-Onge et al., 2012).

59 This study presents a geophysical and sedimentological analysis of the sedimentary infill of the outer  
60 part of Aysén Fjord (~45°S), which allows documenting the first sediment drifts in the region, with the  
61 aim of reconstructing the post-glacial evolution of bottom currents in the Chilean Patagonian fjords.  
62 More specifically, as Aysén Fjord is shielded from the influence of any deep-water masses by the  
63 presence of a bathymetric high resulting in < 50 m water depth (Sievers and Silva, 2008), it forms an  
64 ideal location to isolate the surficial subantarctic water mass and demonstrate its evolution in the  
65 region, which can potentially be related to climatic as well as tectonic variability.

## 66 **2 Regional setting**

### 67 2.1 Oceanographic and hydrographic setting

68 The Pacific Patagonian coastline is characterized by numerous channels and fjords (Fig. 1A), formed as  
69 a result of glacial action and tectonic processes during Cenozoic times (e.g., Glasser and Ghiglione,  
70 2009; Syvitski et al., 1987). Aysén Fjord is one of them, with a length of over 70 km, connecting the  
71 Aysén River and its catchment in the east to the Costa Channel on its western extremity (Fig. 1B). Just  
72 north of Aysén Fjord, the Costa Channel merges with the Errázuriz Channel into the Moraleda Channel.

73 The latter communicates with the Pacific Ocean through the Boca del Guafo (~43.5°S), located south  
74 of Chiloé Island, and almost 200 km north of Aysén Fjord. Due to the limited depth of these channels  
75 (~70-400 m), only the surficial water masses present in the open ocean can enter the Patagonian  
76 channels and fjords (Fig. 1C), including the Subantarctic Water (SAAW, up to 150 m depth) and the  
77 Equatorial Subsurface Water (ESSW, 150-300 m depth) (Sievers and Silva, 2008; Silva and Guzmán,  
78 2006). The SAAW is under direct influence of the ACC (Fig. 1A), currently hitting the South American  
79 continent between around 40°S and 45°S, where it splits into a southward and northward current,  
80 respectively the Cape Horn Current and Humboldt Current (e.g., Strub et al., 1998). The Cape Horn  
81 current pushes the SAAW into the Moraleda Channel through the Boca Del Guafo, where it becomes  
82 mixed with a surface layer of fresh water originating from rivers, coastal runoff, glacial melting and  
83 precipitation. This results in the formation of two separate water masses, termed the Modified  
84 Subantarctic Water (MSAAW) and a surficial, less saline layer about 50 m thick referred to as Estuarine  
85 Water (EW) (Sievers and Silva, 2008; Silva and Guzmán, 2006). The MSAAW can flow as far as ~46.5°S,  
86 thus also entering Aysén Fjord, where it is blocked by the shallowing bathymetry (Sievers and Silva,  
87 2008). The deeper ESSW, on the other hand, is blocked further north by the presence of several  
88 topographic highs on the rather irregular ocean floor, most importantly the Meninea sill, just north of  
89 Aysén Fjord (Fig. 1C).

90 Apart from controlling the surficial oceanic circulation patterns in the southern Pacific, the ACC also  
91 plays an essential role in the regional and even global climate (e.g., Barker and Thomas, 2004; Kilian  
92 and Lamy, 2012; Rintoul, 2018; Sijp and England, 2008). Its location is strongly coupled to the latitude  
93 of the SWW belt and can extend up to 10° more northward during glacial periods (e.g., Kaiser et al.,  
94 2005; Lamy et al., 2004; Verleye and Louwye, 2010). Such changes in position as well as strength of the  
95 SWW are reflected in the regional precipitation patterns west of the Andes, where the amount of  
96 precipitation is positively correlated to wind velocities (e.g., Garreaud et al., 2013; Lamy et al., 2010).  
97 Present-day annual precipitation can be as high as 10,000 mm at around 50°S but fluctuates between  
98 600 and 2,200 mm in the Aysén River watershed, and reaches up to 2,500 mm along the fjord (Fick and

99 Hijmans, 2017). This results in voluminous pluvio-nival rivers feeding the fjord, the largest of which is  
100 Aysén River (Fig. 1B) with an average discharge of 521 m<sup>3</sup>/sec (Calvete and Sobarzo, 2011). River  
101 discharge modulates the thickness of the EW (Calvete and Sobarzo, 2011) and constitutes the main  
102 source of sediment to the fjord, which is almost entirely deposited through settling from the EW  
103 (Bertrand et al., 2012). Glaciers are currently absent from the Aysén watershed. They started retreating  
104 to their present extent after the Last Glacial Maximum (LGM, ~60 to 18 cal ka BP) in response to an  
105 initial warming period that lasted up to ~15 cal ka BP, followed by a plateau and a second temperature  
106 rise around 12.5 cal ka BP (Davies et al., 2020; Haddam et al., 2018; Kilian and Lamy, 2012).

## 107 2.2 Seismic and volcanic activity

108 Southern Chile is a tectonically active region, located along the southern section of the Peru-Chile  
109 subduction trench (Fig. 1B) where the Nazca plate converges with the South American Plate (Fisher  
110 and Raitt, 1962). This results in oblique subduction, of which the trench-parallel component is  
111 accommodated by the presence of an active dextral strike-slip lineament, the Liquiñe-Ofqui Fault Zone  
112 (LOFZ), located along-strike of the subduction zone (Cembrano et al., 1996). Aysén Fjord is located on  
113 the LOFZ, about 200 kilometres from the subduction zone (Fig. 1B). The last significant earthquake ( $M_w$   
114 6.2) along the LOFZ occurred in April 2007 and had its epicentre located in the inner part of Aysén Fjord  
115 (Legrand et al., 2011), triggering multiple landslides in its catchment (Sepúlveda et al., 2010). The  
116 sedimentary infill of the fjord revealed that similar crustal earthquakes have occurred throughout the  
117 Holocene (Wils et al., 2018), and even shaking related to past megathrust earthquakes has affected  
118 the fjord (Wils et al., 2020). Another direct consequence of the subduction zone is the presence of a  
119 volcanic chain with mostly tholeiitic and high-Al basalts as well as basaltic andesites (Stern, 2004). The  
120 main volcanoes in the vicinity of Aysén Fjord are the Mentolat, Cay, Macá, and Hudson Volcano (Fig.  
121 1B). The only known eruptions for Mentolat and Macá Volcano occurred ~7,500 and ~1,400 cal yrs BP,  
122 respectively, while no records of Holocene activity exist for Cay Volcano (Naranjo and Stern, 2004).  
123 Hudson Volcano has had multiple major eruptions during the Holocene, including the H1 (~8,200 cal  
124 yrs BP) and H2 (~3,900 cal yrs BP) eruptions and several other smaller activity (Naranjo and Stern, 1998;

125 Stern and Weller, 2012). Although tephra deposits related to these eruptions have been identified  
126 within the sedimentary infill of Aysén Fjord (Wils et al., 2018), volcanic fallout in the Patagonian fjords  
127 is generally limited due to the western-blowing SWW (Fontijn et al., 2014).

### 128 **3 Material and methods**

#### 129 3.1 Bathymetry data

130 Multibeam data was recorded in March 2013 during the DETSUFA cruise on board *BIO Hespérides*  
131 (Lastras and The Shipboard Scientific Party, 2013). An EM 1002S Kongsberg multibeam echosounder  
132 was used, providing high-resolution bathymetric data in the entire fjord (Fig. 2). The system operated  
133 with an equidistant beam spacing, using a nominal sonar frequency of 95 kHz and 111 beams of 2x2°  
134 per ping. Swath width was set to 300 m, resulting in a footprint of about 3 m (Lastras and The Shipboard  
135 Scientific Party, 2013). After processing, the resulting bathymetry map was imported in IHS Kingdom  
136 Suite software for integration with the seismic reflection profiles.

137 To complement the multibeam data, which does not cover the entire outer fjord area, the GEBCO 2020  
138 bathymetry grid was used, which has a 450 m resolution (GEBCO Compilation Group, 2020).

#### 139 3.2 Seismic reflection profiles

140 Reflection-seismic profiles were also acquired during the DETSUFA cruise, resulting in a dense network  
141 of long, (mostly) fjord-parallel lines (Fig. 2A). A Kongsberg TOPAS PS18 parametric subbottom profiler  
142 was used, operating at a secondary central frequency of 1.5-4.5 kHz (Lastras and The Shipboard  
143 Scientific Party, 2013). This resulted in a theoretical vertical resolution of about 0.2 ms two-way travel  
144 time (TWT), which corresponds to ~15 cm. Real-time processing of the seismic data was performed on  
145 board of the *BIO Hespérides*, after which the data was imported in the IHS Kingdom Suite software for  
146 visualization and interpretation. Seismic-stratigraphic units were distinguished based on visible  
147 changes in acoustic reflection and spatial distribution.

### 148 3.3 Sedimentological analyses

149 Two Calypso sediment cores (MD07-3114 and MD07-3115, Fig. 2) were taken in the outer fjord during  
150 the PACHIDERME survey in February 2007 on board *RV Marion-Dufresne* (Kissel et al., 2007). Based on  
151 preliminary on-board interpretations of the seismic data, these cores were retrieved with the intention  
152 of having a partly overlapping sedimentary sequence. Both cores have a length of just under 32 metres,  
153 and were split and described immediately after retrieval, after which they were scanned at a 2 cm  
154 interval using the on-board multi-sensor core logger to reveal, among others, their gamma density and  
155 magnetic susceptibility (MS).

156 Grain-size distributions were measured on 1 cm-thick slices using a Malvern Mastersizer 3000. A  
157 sampling resolution of 20 cm was applied throughout core MD07-3114 and increased to 10 or even 5  
158 cm in specific areas of interest. Only the first 3.5 m of core MD07-3115 (i.e. the part that overlaps with  
159 core MD07-3114 based on preliminary correlation, Kissel et al. (2007)) were sampled for grain-size  
160 analysis, with samples retrieved every 4 cm (when possible). The terrigenous sediment fraction was  
161 isolated by chemical pre-treatment with H<sub>2</sub>O<sub>2</sub> (2 ml, 30%), HCl (1 ml, 10%) and NaOH (1 ml, 2 N) to  
162 remove organic matter, carbonates and biogenic silica, respectively. Grains with a diameter larger than  
163 1 mm were excluded for analysis by sieving. All samples were boiled with a sodium  
164 hexametaphosphate solution (1 ml, 2%) prior to analysis to ensure complete disaggregation. The  
165 geometric mean grain size and sorting values for the total grain-size distributions and sortable silt (SS)  
166 fractions (10–63 µm, McCave et al., 1995) were calculated using the GRADISTAT software (Blott and  
167 Pye, 2001).

168 Glass shards in two tephra layers at 15.9 m depth in core MD07-3114 and 2.24 m in core MD07-3115  
169 have been analysed for their major-element composition using a CAMECA-SX 100 Electron Microprobe  
170 (EPMA-CAMPARIS) at the Université Paris VI in France following the procedure by Carel et al. (2011).  
171 Ten elements were analysed (Na, Mg, Si, Al, Cl, K, Ca, Ti, Mn, and Fe) using an accelerating voltage of  
172 15 KV, a current of 10 nA and a beam size of 5 µm to minimize loss of alkalis such as Na. Precisions on



173 individual shards ( $1\sigma$ ) were better than 0.6% for Si, ~1% for Al, 3% for Ca and Mg, 4% for Na, 5% for Fe,  
174 6% for K, 10% for Ti, and about 30% for Cl and Mn.

175 Radiocarbon ages for selected depths in both cores have been obtained by Serno (2009) using a  
176 combination of plant remains and shells (Table 1). The occasional occurrence of shells and plant debris  
177 is most common in core MD07-3114. Here, we recalibrate all these radiocarbon ages using the more  
178 recent SHCal20 calibration curve (Hogg et al., 2020) and combine them into a single age-depth model  
179 for both cores using the Bacon R-package (Blaauw and Christen, 2011). A reservoir age of 550 years is  
180 considered for the marine samples (Serno, 2009), to which we add an uncertainty of 40 years—  
181 corresponding to the average error on the radiocarbon ages themselves.

## 182 **4 Results**

### 183 4.1 Seismic stratigraphy and bathymetric description

184 The seismic profiles allow defining six acoustic units that overly the acoustic basement—below which  
185 only locally unclear reflectors are observed—and are numbered in order of deposition ( , 4). The oldest of these units (Unit 0) consists of a succession of mostly parallel low- and high-  
186 amplitude reflections. This thick unit can only be identified in the deepest parts of the basin, where  
187 seismic penetration is limited, and hence, the base of this unit cannot be observed. Its distribution is  
188 bounded by the steep bedrock slopes associated to the mainland and/or some of the islands in the  
189 fjord (Casma, Churrecue and Elena Island, Fig. 2A). This is also true for Unit 1, a relatively thin, ponding  
190 unit with a largely acoustically transparent seismic facies near the western extent of the fjord, gradually  
191 becoming more stratified towards the inner fjord. Unit 2 consists of a succession of closely-spaced  
192 parallel high-amplitude reflections, filling up the deepest part of the basin. As a result, Unit 3 extends  
193 further towards the inner fjord but remains bounded by the subaquatic continuation of Churrecue and  
194 Elena Island. Several parallel high-amplitude reflections can be identified, albeit with a more  
195 pronounced alternation with lower-amplitude reflections compared to Unit 2. Unit 4 can be identified

197 throughout most of the outer part of the fjord, consisting of a seismic facies with parallel, low-  
198 amplitude reflections and a single strong horizon in the lower half. With the exception of slight  
199 sediment thickening south of core MD07-3115 (Fig. 4), giving the unit a subtle mounded nature,  
200 variations in sediment thickness are mostly restrained by the irregular bedrock topography, as is the  
201 case for the older sedimentation in the fjord. The most recent seismic unit (Unit 5) consists of mostly  
202 low-amplitude reflections and covers the entire outer part of Aysén Fjord, with the presence of a single,  
203 basin-wide high-amplitude reflection (Figs. 3, 4). In the centre of the basin, this unit is mostly  
204 subparallel stratified. Unit 5 has a similar seismic facies compared to Unit 4, however, several mounded  
205 morphologies are present showing a gradual thinning of the reflectors mostly towards the basin  
206 periphery (e.g.,  
  
207 , 4). In contrast to the units below, Unit 5 thus shows strong thickness variability that results in several  
208 mounded areas and local depressions with a relief of 30 m or more. This includes the mounded area  
209 in which core MD07-3114 was retrieved, which covers about 6.5 km<sup>2</sup> and is bounded in the west by an  
210 elongated, roughly north-south oriented depression of similar extent visible in the present-day  
211 bathymetry (Fig. 2B). This depression stretches from the location of core MD07-3115 to the southern  
212 limit of the fjord (into the Costa Channel), but is less pronounced in the central part of the basin (Fig.  
213 2B, Fig. 4). In this central part, the transition from the depression to the mounded area is punctuated  
214 by a subsurface fault (Fig. 2B, Fig. 4). Another major mounded sediment package of over 20 m thick is  
215 present in the central part of the outer fjord basin (~13 km<sup>2</sup>), confined between the north-south  
216 oriented depression in the east and a series of small (<0.5 km<sup>2</sup>), individual depressions with a comma-  
217 like shape towards the western extremity of the fjord, close to Casma Island (Fig. 2C). These latter  
218 depressions are much steeper and narrower compared to the central depression (Fig. 2B), and older  
219 sedimentation (Units 3 and/or 4) is occasionally eroded here (Fig. 4). A more pronounced case of  
220 erosion can be observed in the elongated east-west oriented depression of about 4 km long and 1.5  
221 km wide just northwest of the coring locations along the shoreline of Churrecue Island (Fig. 2A). This

222 forms an erosive channel-like structure that seems to prevent deposition of recent sediments and in  
223 which erosion occurred down to the base of Unit 3 (  
224 ).

#### 225 4.2 Sedimentological characterization

226 Both sediment cores have a mostly light-coloured greyish fine-grained appearance and are bioturbated  
227 throughout. Core MD07-3114 is located in an area where Unit 5 is relatively thick (mounded), whereas  
228 MD07-3115 is retrieved in one of the depressions where the recent sedimentation is more condensed  
229 (  
230 ). Therefore, core MD07-3114 allows visualizing the recent sedimentation in great detail, while MD07-  
231 3115 provides a window to the older sedimentation in this part of the fjord.

232 The lower 5 m of sediments in core MD07-3115 consists of predominantly siliciclastic sediments with  
233 a very low diatom content, including marine, brackish and freshwater species (Kissel et al., 2007).  
234 Additionally, this sediment package shows strong fluctuations in MS and density (Fig. 5) consisting of  
235 a sharp rise at their base followed by a gradual decrease. Each of these peaks corresponds to a dark,  
236 visually coarser-grained layer. These darker beds show a sharp basal contact and are generally  
237 laminated. Considering that density contrasts form the base for changes in acoustic impedance, these  
238 layers are considered as the sediment beds that are responsible for the closely-spaced high-amplitude  
239 reflections in Unit 2 (Fig. 3). Sedimentation remains siliciclastic up to about 19.5 m core depth, showing  
240 similar dark sediment beds characterized by strong fluctuation in MS and density, although they are  
241 less numerous here (Fig. 5). We therefore interpret that these high-density beds correspond to the  
242 less-closely spaced strong reflectors in Unit 3.

243 At around 19.5 m depth in MD07-3115, the lithology abruptly changes from dominantly siliciclastic to  
244 diatom-rich clays. Up to about 7 m, sediments are also lighter-coloured compared to the sediments  
245 below. MS and density reach their lowest values in this interval and start increasing from about 14 m  
246 depth (Fig. 5). At about 13 m, a tephra layer is present resulting in a sharp peak in MS and density that

247 we correlate to the strong reflector present in the lower part of Unit 4. Similar light-coloured diatom-  
248 rich clays can be identified in the lowermost part of core MD07-3114, hinting towards a first part of  
249 overlap between both sediment cores. This overlap is confirmed by a darker-coloured sediment  
250 interval that can be identified on top of the light-coloured sediments in both cores (between 7 and 4.5  
251 m in MD07-3115 and 31 to 28 m in MD07-3114), marked by two well-defined, quasi-identical increases  
252 in MS that coincide with slight mean grain-size increases visible in MD07-3114 (Fig. 5). On top of this  
253 darker-coloured sediment interval, sedimentation resumes to its previous lighter colours and shows  
254 another, considerably smaller, MS peak. Up to here, sedimentation is thus roughly consistent in both  
255 core locations and no noteworthy lateral thickness variations occur. This is no longer the case for the  
256 most recent sedimentation in both cores, as marked by their clearly deviating trends in MS (Fig. 5).  
257 Therefore, we consider this boundary to mark the base of Unit 5, where sediment thinning results in  
258 the upper ~27 m of sediment in core MD07-3114 to be condensed into the upper ~3 m of sediment in  
259 core MD07-3115 (  
260 ).

261 Sedimentation in Unit 5 is again lighter in colour and remains diatom-rich. Similar to the seismic facies,  
262 there is no visual change in sedimentary facies between Unit 4 and Unit 5 (  
263 , Fig. 4). Nevertheless, the base of Unit 5 is marked by an abrupt increase in SS mean grain size in both  
264 cores, after which SS mean grain size shows an overall gradual decrease, albeit divided in two steps  
265 (Fig. 5). The transition between the two steps (~13 m in MD07-3114, ~2 m in MD07-3115) is marked  
266 by a through in SS mean grain size. Despite these similar trends in both core locations, SS values are  
267 markedly higher in MD07-3115 (where Unit 5 is more condensed) compared to MD07-3114—a trend  
268 that is also reflected in the mean grain size. This underscores the lateral variability in sedimentation,  
269 supported by the remaining sedimentary parameters; MS, density and mean grain size roughly follow  
270 the same trend as the SS at the location of MD07-3115, while these parameters remain rather stable

271 in core MD07-3114 (Fig. 5). Only the upper 4 m of sediment in core MD07-3114 show an increasing  
272 trend in MS and density values.

273 A coarse-grained tephra layer can be identified at both core locations, resulting in a sharp MS peak as  
274 well as an increase in mean grain size (after sieving and removal of grains with a diameter > 1 mm) that  
275 is not reflected in the SS. This tephra layer is present at about 16 m depth in core MD07-3114 and at  
276 2.2 m in core MD07-3115, corresponding to the basin-wide high-amplitude reflection present in Unit  
277 5 at about two third of its total thickness (

278 ). Correlation of both tephras is verified by major-element geochemical analysis, showing the same  
279 basaltic andesite composition (Fig. 6). They are thus considered to result from the same volcanic  
280 eruption. Unfortunately, comparison to the known geochemical composition of major volcanoes and  
281 some of the many monogenetic cones in the vicinity of the fjord (D'Orazio et al., 2003; Gutiérrez et al.,  
282 2005; Haberle and Lumley, 1998; Kratzmann et al., 2009; Naranjo and Stern, 1998) is not conclusive.  
283 The Si-content excludes an origin related to the less evolved (basaltic) monogenetic cones, while the  
284 Ti content shows strong affiliations with Hudson eruptions and the K content points to either the Macá,  
285 Mentolat, or Cay Volcano.

### 286 4.3 Chronology

287 The correlation of both cores as elaborated in the previous section allows constructing a 'composite'  
288 core for which a single, continuous chronology can be established. With this composite core, the older  
289 sediments in the fjord as well as the recent sedimentation are covered in great detail (Fig. 7). The  
290 model shows two marked changes in sedimentation rate: one at the transition between siliciclastic  
291 (Unit 3) and diatom-rich (Unit 4) sedimentation at around 12.3 cal ka BP and another at the onset of  
292 Unit 5 around 3.7 cal ka BP. The sedimentation rates in the siliciclastic interval (Unit 3) were markedly  
293 higher (~1.7 cm/yr) compared to the diatom-rich interval (Unit 4; ~0.2 cm/yr). The onset of Unit 5 is  
294 marked by another abrupt increase in sedimentation rates to ~0.8 cm/yr when considering the  
295 sedimentary sequence present in core MD07-3114, while sedimentation rates in core MD07-3115—

296 located in an area of sediment thinning and characterized by a more condensed Unit 5—decrease to  
297  $\sim 0.08$  cm/yr (i.e., 3 m of sediment in  $\sim 3.7$  ka).

298 The age model additionally allows relating the different tephra layers to well-described volcanic  
299 eruptions in the southern Andes, validating the obtained model. The most prominent tephra layer in  
300 this part of the fjord is present in Unit 4 (8,130–8,670 cal yrs BP) and shows strong overlap with the  
301 age range of one of Hudson Volcano's major eruption (H1, 7,690–8,720 cal yrs BP Stern and Weller  
302 (2012)). As a tephra deposit related to this eruption has also been identified in the inner part of the  
303 fjord (Wils et al., 2018), we consider this tephra to result from the H1 eruption (Fig. 7). The tephra layer  
304 in Unit 5 (2,150–2,500 cal yrs BP) we interpret to result from Hudson's T6 eruption ( $< 2,060$ – $2,420$  cal  
305 yrs BP, Naranjo and Stern (1998)), showing the best age range overlap compared to the Holocene  
306 eruptions of volcanoes in the vicinity of the fjord (Fig. 7).

## 307 **5 Discussion**

### 308 5.1 Present-day bottom current circulation patterns

309 Seismic profiling in combination with multibeam bathymetry revealed that the modern seafloor  
310 topography in the outer part of Aysén Fjord consists of several local mounded areas and depressions  
311 as a result of large lateral thickness variability by sediment concentration during deposition of Unit 5 (  
312 , 4). This bathymetric morphology cannot be related to any features on land (Fig. 2) and the  
313 sedimentary characteristics of Unit 5 show significantly coarser SS grain sizes compared to the unit  
314 below, increasing abruptly at its base. These observations indicate that the boundary between Unit 4  
315 and 5 marks the transition from sedimentation dominated by hemipelagic settling to a current-driven  
316 sedimentary environment, resulting in the development of sediment drifts (cf., McCave et al., 1995).  
317 The areas of sediment thinning can thus be interpreted as moats, whereas the mounded areas  
318 constitute the drift body (Fig. 8A). Core MD07-3115 is thus retrieved at the location of the central  
319 moat—where currents are strong and only coarser-grained material can be deposited—while MD07-  
320 3114 is located close to the crest of the eastern drift and consists of finer-grained material

321 accumulating at a rate that is about nine times higher when considering the thickness difference of  
322 unit 5 (Fig. 5). The lack of any visible draping hemipelagic sedimentation on either of the seismic  
323 profiles in the fjord indicates that drift formation is still ongoing. Likewise, the elongated, erosive  
324 channel located just northwest of the coring locations has no recent sedimentation (  
325 ) and is thus also still active to date, likely representing a contourite channel (García et al., 2009;  
326 Rebesco et al., 2014). In this way, the identified sediment drifts and channel can provide the first direct  
327 information on the present-day bottom current circulation patterns in the southern part of the  
328 Patagonian fjords (Fig. 8).

329 According to Sievers and Silva (2008), the only deep water mass present south of the Meninea sill, and  
330 thus in the study area at the conjunction of Aysén Fjord (to the east) and Costa Channel (to the south),  
331 is the MSAAW (Fig. 1). Due to the shallowing bathymetry in the southern part of the Costa Channel  
332 and south of Casma Island, the inflow of water masses from the Moraleda Channel into Aysén Fjord  
333 must occur through the gateway north of Casma Island, where the bathymetry is slightly deeper (Fig.  
334 8). Hence, the Costa Channel is characterized by a roughly southward flow direction formed by the  
335 incoming MSAAW. The inflow of MSAAW in Aysén Fjord thus originates from the west, after which it  
336 continues southward. As such, the interpreted contourite channel along the coastline of Churrecue  
337 Island is considered to transport the incoming water mass eastwards into Aysén Fjord, consistent with  
338 the Coriolis effect—considered the driving force to cause bottom currents to stick to slopes (Rebesco  
339 et al., 2014)—that results in a current deflection to the left in the Southern Hemisphere. In this way,  
340 the contourite channel is likely the result of spilling of the MSAAW over the bathymetric high between  
341 Casma and Churrecue Island, forming a ‘plunge pool’ that would explain its erosive nature. In the east  
342 of this contourite channel, at least a part of the incoming water mass is deflected southward, towards  
343 the Costa Channel. This is likely related to the presence of a bathymetric high north of both core  
344 locations, but cannot be verified at this stage due to the lack of high-resolution bathymetry coverage  
345 in this area. In any case, this southward deflection resulted in current weakening, so that the contourite

346 channel gradually evolves into a moat and the formation of the eastern sediment drift in which both  
347 cores were retrieved (  
348 , Fig. 8A). The southern part of this drift can be identified as fault-controlled (Rebesco et al., 2014),  
349 while the northern part is likely formed in between two currents (i.e., the one flowing southward  
350 towards the Costa Channel and the other flowing eastward into Aysén Fjord). The western drift  
351 constitutes a confined drift (Rebesco et al., 2014), bounded by currents on its eastern and western  
352 margin as evidenced by the moats present there. However, the small, individual comma-shaped  
353 depressions located in the western extremity of the fjord (Fig. 2) indicate a clear northward current  
354 direction. These must thus be formed by a returning MSAAW current, strongly slope-bound in  
355 correspondence with the Coriolis effect. It is, however, unclear what controls this outgoing flow  
356 direction. This might be related to a circular flow pattern of the MSAAW, in which new MSAAW is  
357 continuously flowing into the fjord, then south in the Costa Channel and back north, while it gradually  
358 mixes with the overlying water mass at a rate that is in balance with the inflow. Alternatively, the  
359 MSAAW in the study area is a relatively stable waterbody that was refilled only sporadically, and  
360 internal tides and wind forcing (e.g., Cáceres et al., 2002) combined with the Coriolis effect result in  
361 south- and northward bottom currents that stick to the eastern and western slopes, respectively. A  
362 third, and most likely, option is a combination of these two processes, in which tides control the inflow  
363 of MSAAW and through internal tides most of the currents, but a net clockwise current and gradual  
364 mixing with the EW is active as well.

365 The above interpretation shows that a detailed investigation of the drift-related structures allows  
366 linking them with the present-day circulation pattern. However, unravelling how these patterns were  
367 established is related to the mechanism behind drift formation, thus requiring thorough insight into  
368 the sedimentation and bottom water evolution of the fjord through time.



## 369 5.2 Sedimentation history and climate variability

### 370 5.2.1 Late Glacial (Units 0–3, ~18–12.3 cal ka BP)

371 The oldest sedimentation in the outer part of Aysén Fjord (Unit 0, over 30 m thick) is considered to be  
372 deposited since ~18 ka, corresponding to the onset of deglaciation in the area (e.g., Davies et al., 2020;  
373 Haddam et al., 2018; Kilian and Lamy, 2012). Up until stratigraphic Unit 3, sedimentation is confined  
374 by the irregular bedrock topography and restricted to the deepest parts of the outer fjord basin (

375 ). Sediments are dominantly siliciclastic with high MS values (Fig. 5), in agreement with the high  
376 terrigenous input generally observed during deglaciation (e.g., Hebbeln et al., 2007; Siani et al., 2010).  
377 Nevertheless, the mixture of marine, brackish and freshwater diatoms present in Units 2 and 3 (Kissel  
378 et al., 2007) suggests that a marine influence was being imposed towards the end of this stratigraphic  
379 sequence, consistent with continuously retreating glaciers and a relative sea-level rise. The deepest  
380 part of the fjord thus consists of rapid late-glacial sedimentation, reaching values of ~1.7 cm/yr in Unit  
381 3 (Fig. 7) that are likely even higher during deposition of Unit 2 considering the more rapid succession  
382 of high-amplitude reflectors (

383 ). These are the result of the numerous coarse-grained beds with sharp basal contacts that can thus be  
384 interpreted as turbidites. They could be related to glacial lake outburst floods, currently frequently  
385 occurring near the Northern Patagonian Icefield (Dussailant et al., 2009; Vandekerkhove et al., 2020)  
386 to which the pre-Holocene setting of Aysén Fjord may have been similar (Davies et al., 2020; García et  
387 al., 2018). Continuous recession of these large and thick glaciers in the area thus explains their  
388 decreasing frequency from Unit 2 to Unit 3. Alternatively, turbidites could also have a seismic trigger,  
389 as postglacial rebound could result in increased activity of one of the numerous LOFZ-related faults in  
390 the area or even the megathrust (cf., Beck et al., 1996; Brooks and Adams, 2020; Stewart et al., 2000).

### 391 5.2.2 Holocene (Units 4–5, ~12.3 cal ka BP – present)

392 Since roughly the beginning of the Holocene, sedimentation is less strongly influenced by the input of  
393 high-MS glacial sediments but is dominated by fine-grained mud rich in diatoms (Fig. 5). Units 4 and 5

394 thus represent hemipelagic sedimentation and are post-glacial in age. The entire Aysén Fjord and most  
395 likely a large part of its watershed are considered to be deglaciated by 12.3 cal ka BP, consistent with  
396 the minimum age of glacial outwash deposits ~15 km inland (Vargas et al., 2013), and with the glacial  
397 history of the nearby Cisnes Valley (García et al., 2018). Unit 4 stretches out over the entire outer fjord  
398 basin, indicating a relative higher sea level compared to the more confined Unit 3. This is in agreement  
399 with the rapid sea-level rise following deglaciation up to ~7 cal ka BP (Garrett et al., 2020; Lambeck et  
400 al., 2014; Siddall et al., 2003).

401 The main difference between Units 4 and 5 resides in their morphology and sedimentation rate rather  
402 than their composition. Unit 5 shows the presence of thick sediment drifts whereas Unit 4 is draping  
403 and displays significantly lower accumulation rates, at least compared to the mounded areas of Unit 5  
404 (~0.2 compared to 0.8 cm/yr; Fig. 7). Subtle sediment drifts already occur in Unit 4 (Fig. 4), reflecting  
405 the increasingly strong marine influence resulting from the (sporadic) inflow of the MSAAW into the  
406 fjord basin that only becomes dominant in Unit 5.

407 Independently of the presence of drifts, the Aysén sediment record displays several intervals with high  
408 MS values at 5.5–4.0 cal ka BP and during the last 500 years, superimposed on generally higher MS  
409 values after 9.5 cal ka BP (Fig. 9). Since MS is not significantly correlated to sediment grain size ( $R = -$   
410  $0.19$ ,  $p > 0.01$ ), increases in MS along the core likely reflect intervals richer in detrital particles, i.e.,  
411 increases in terrigenous input by rivers. The progressive increase in MS between 9.5–8.0 cal ka BP is  
412 coeval with the main increase in precipitation and westerly wind speed derived from the sediments of  
413 Lago Castor (Fiers et al., 2019; Van Daele et al., 2016) (Fig. 9), located in the upper Aysén watershed  
414 (Fig. 1A), suggesting that it reflects a gradual increase in precipitation of westerly origin. The latter is  
415 coeval with a decrease in SWW speed at 52–53°S and therefore likely reflects an expansion (Lamy et  
416 al., 2010) or northward shift (Lamy et al., 2001) of the entire SWW belt. In addition to this long-term  
417 shift, the two intervals with the highest MS values (5.5–4.0 cal ka BP and last 500 years) correspond  
418 remarkably well to two periods of low Holocene sea surface temperature (SST) in the Concepción

419 Channel (Fig. 1A), characterized by a strong marine influence (Caniupán et al., 2014) (Fig. 9). Such a  
420 relation between decreasing SST and higher detrital input has been observed for the last millennium  
421 in two fjords immediately north (Jacaf Fjord; Sepúlveda et al., 2009) and south (Quitralco Fjord;  
422 Bertrand et al., 2014) of Aysén Fjord (Fig. 1A, Fig. 9), where it was explained as the response of the  
423 SWW belt to changes in temperature. Decreases in temperature result in a stronger polar cell, which  
424 in turn displaces the SWW equatorward, resulting in increased year-round precipitation and therefore  
425 increased terrestrial input in the Aysén Fjord area (Bertrand et al., 2014). This relation between SST  
426 and precipitation is likely also responsible for the increase in detrital input at 5.5–4.0 cal ka BP, which  
427 is supported by the 1.5°C decrease in SST at 50°S (Fig. 9) and the darker sediment colour in Aysén Fjord  
428 (Fig. 5), likely reflecting increased terrestrial organic matter content. This time interval additionally  
429 corresponds to the interval of highest Holocene precipitation (Fiers et al., 2019; Markgraf et al., 2007)  
430 and westerly wind speed (Van Daele et al., 2016) in the upper Aysén watershed (Fig. 9). Finally, the  
431 more moderate MS increases throughout the Holocene, especially at 8.0–7.5 and 6.2–6.0 cal ka BP,  
432 may also represent increased precipitation as they correspond to minor drops in SST (Fig. 9), although  
433 these may not be relevant at regional scale. This interpretation of the MS values suggests that the  
434 sediments of Aysén Fjord register regional changes in precipitation, like most other fluvially-fed fjords  
435 in northern Chilean Patagonia (e.g., Bertrand et al., 2014; Sepúlveda et al., 2009). Based on available  
436 datasets, the presence of sediment drifts during the Late Holocene seems to be a unique feature of  
437 Aysén Fjord, and can therefore not solely be related to climate variability.

### 438 5.2.3 Late Holocene sediment drifts

439 After an initial increase in SS grain size at the onset of drift formation, variations in SS grain size within  
440 the sediment drift suggests that bottom currents gradually weakened during the last 3.7 cal ka BP (Unit  
441 5; Fig. 5 and 9). This weakening trend matches the decrease in westerly wind speed observed in the  
442 sediments of Lago Castor (Fiers et al., 2019; Van Daele et al., 2016) (Fig. 9), suggesting a SWW control  
443 on MSAAW inflow and bottom current strength in Aysén Fjord. In addition, the decrease in SWW speed  
444 during the last 3.7 ka resulted in a direct decrease in precipitation in the Aysén watershed (Fiers et al.,

445 2019), which in turn led to a thinner EW layer in Aysén Fjord (cf. Silva and Guzmán, 2006) and thus a  
446 lower salinity gradient. This generally decreases estuarine circulation patterns (Geyer, 2010), and could  
447 thus also contribute to the weakening of bottom water flows. These two mechanisms thus reinforce  
448 each other and resulted in a marked decrease in bottom current strength that is reflected in the SS  
449 grain size (Fig. 9).

450 In addition to the general decrease in SS grain size in the sampled sediment drift, the SS grain size also  
451 shows a minimum strength around 1.9 cal ka BP (Fig. 5 and 9). This coincides with a period of regionally  
452 lower SWW strength and precipitation around 2 cal ka BP as observed in Mallín Pollux (Markgraf et al.,  
453 2007) and Lago Shaman (de Porrás et al., 2012). This implies that even centennial-scale changes in  
454 westerly wind speed modulate the strength of bottom currents in the Patagonian fjords, most likely  
455 through decreased MSAAW inflow combined with EW thinning.

### 456 5.3 Climatic and tectonic controls on the onset of drift formation

457 The two major sediment drifts in Aysén Fjord were formed simultaneously around 3.7 cal ka BP.  
458 Although bottom currents were likely already present earlier, the drift structures in Unit 4 are much  
459 less pronounced (Fig. 4) and sedimentation (rate) was still relatively uniform at both core sites (Fig. 7).  
460 This points to a sudden change in bottom current patterns and/or local increase in strength at the  
461 transition from Unit 4 to Unit 5, rather than a first appearance of the MSAAW. This is supported by the  
462 contourite channel, which must have been formed after or at the end of deposition of Unit 4 to explain  
463 the lack of sediment thinning in Unit 4 in that area (  
464 ). Moreover, this is consistent with the ACC already extending further north than Chiloé Island since  
465 ~9.8 cal ka BP, after which it did not shift any further south (Verleye and Louwye, 2010), so that the  
466 SAAW likely already flowed through the Moraleda Channel since the early Holocene.

467 The onset of drift sedimentation in Aysén Fjord at 3.7 cal ka BP does not seem to reflect a marked  
468 change in regional climate conditions. Records around 45°S show the presence of relatively strong  
469 SWW during the late Holocene (de Porrás et al., 2012; Fiers et al., 2019; Markgraf et al., 2007; Moreno

470 et al., 2019; Van Daele et al., 2016), but without any specific increase near 3.7 cal ka BP that could  
471 explain the onset of sediment drifts. If anything, most regional records concur to indicate that by 3.7  
472 cal ka BP the SWW were well-established in the region (de Porras et al., 2012; Moreno et al., 2019;  
473 Quade and Kaplan, 2017) with some records even suggesting a gradual decrease in westerly wind  
474 speeds from 4–5 cal ka BP onwards (Fiers et al., 2019; Van Daele et al., 2016). Hence, the abrupt  
475 increase in erosion and drift formation does not seem to be related to a change in SWW speed near  
476 45°S. Records located further north (41°S), however, suggest an equatorward expansion of the SWW  
477 (Lamy et al., 2010; Lamy et al., 2002) and ACC (Verleye and Louwye, 2010) during the late Holocene,  
478 which may have strengthened the Cape Horn current and ultimately resulted in an increase in the  
479 inflow of SAAW to the Chilean Fjords through Boca del Guafo (43–44S°), as observed during modern  
480 winters (Strub et al., 2019). However, it is hard to imagine that climatic forcing alone would cause such  
481 an abrupt start in drift sedimentation and results in the formation of a strongly-erosive contourite  
482 channel. Although there is clear relation between climate variability and sedimentation within the drift  
483 due to the strong control of SWW speed on ACC inflow, an alternative hypothesis is required to explain  
484 the sudden onset of sediment drift formation in Aysén Fjord.

485 Compilation of paleoseismological research in several locations along the Chilean subduction zone  
486 revealed that a major megathrust earthquake occurred roughly along the northern Patagonian fjords  
487 (~42–46°S) around 3.9 cal ka BP (Wils et al., 2020). This earthquake triggered a tsunami, of which the  
488 imprint in two coastal lakes (Lake Huelde and Lake Cucao) on Chiloé Island was exceptionally large  
489 compared to other tsunami deposits (Kempf et al., 2020; Kempf et al., 2017). Moreover, sedimentation  
490 in Lake Cucao was significantly altered after this earthquake, which has been attributed to coseismic  
491 coastal subsidence (Kempf et al., 2020). The same area subsided about 1 m during the 1960 Great  
492 Chilean Earthquake (Plafker and Savage, 1970). Therefore, subsidence during the 3.9 cal ka BP  
493 earthquake could have resulted in an abrupt absolute lowering of the Meninea sill or any other  
494 blockage, enforcing the precipitation-driven effects on the EW. Tectonic activity could thus have  
495 resulted in a sudden increase in MSAAW presence in Aysén Fjord, thus contributing to the present-day

496 erosion and deposition pattern in the fjord. Nevertheless, further research on the role of the Meninea  
497 sill and other topographic heights in the Patagonian channels is required to fully understand the  
498 processes that govern bottom current circulation patterns in the area.

## 499 **6 Conclusions**

500 Thorough analysis of multibeam bathymetry, seismic-reflection data and two radiocarbon-dated ~30  
501 m long piston cores retrieved in the outer part of Aysén Fjord in Chilean Patagonia allowed  
502 reconstruction of the late glacial and Holocene sedimentation history of the fjord. This revealed for  
503 the first time the presence of active sediment drifts in the Patagonian fjords and channels, providing  
504 new insights into the bottom current patterns of northern Patagonia. The oldest sediments in the fjord  
505 are siliciclastic and deposited during the late glacial, following the onset of glaciation in the area after  
506 ~18 cal ka BP. The subsequent sea-level rise was associated to a period of dominantly marine  
507 hemipelagic sedimentation, which is maintained throughout the duration of the Holocene. Most of the  
508 sedimentary variability in the fjord can be related to variability in the SWW belt, controlling the amount  
509 of terrigenous input into the fjord. Since 3.7 cal ka BP, several large sediment drifts are being formed  
510 in the fjord. Their location allowed fine-tuning the previously-established flow patterns in northern  
511 Patagonia, showing an incoming as well as a returning flow of the MSAAW. The onset of drift formation  
512 is tentatively attributed to the occurrence of a megathrust earthquake, further enhancing the influence  
513 of climate on drift formation. This highlights the importance of climatic and tectonic factors on current  
514 patterns in the Patagonian fjords and channels and underscores the need for multi-disciplinary studies  
515 to understand the evolution of bottom currents through time, especially in complex land-constrained  
516 settings.

## 517 **Data availability**

518 All data necessary to understand the conclusions of this study are presented in the manuscript and/or  
519 supplementary info. The studied sediment core halves are stored in the core repository of the Alfred  
520 Wegener Institute in Bremerhaven, Germany. Core data (grain size, magnetic susceptibility and

521 radiocarbon) are publicly available through the PANGAEA data repository:  
522 <https://doi.org/10.1594/PANGAEA.932303>. Seismic profiles are available upon request to the authors.

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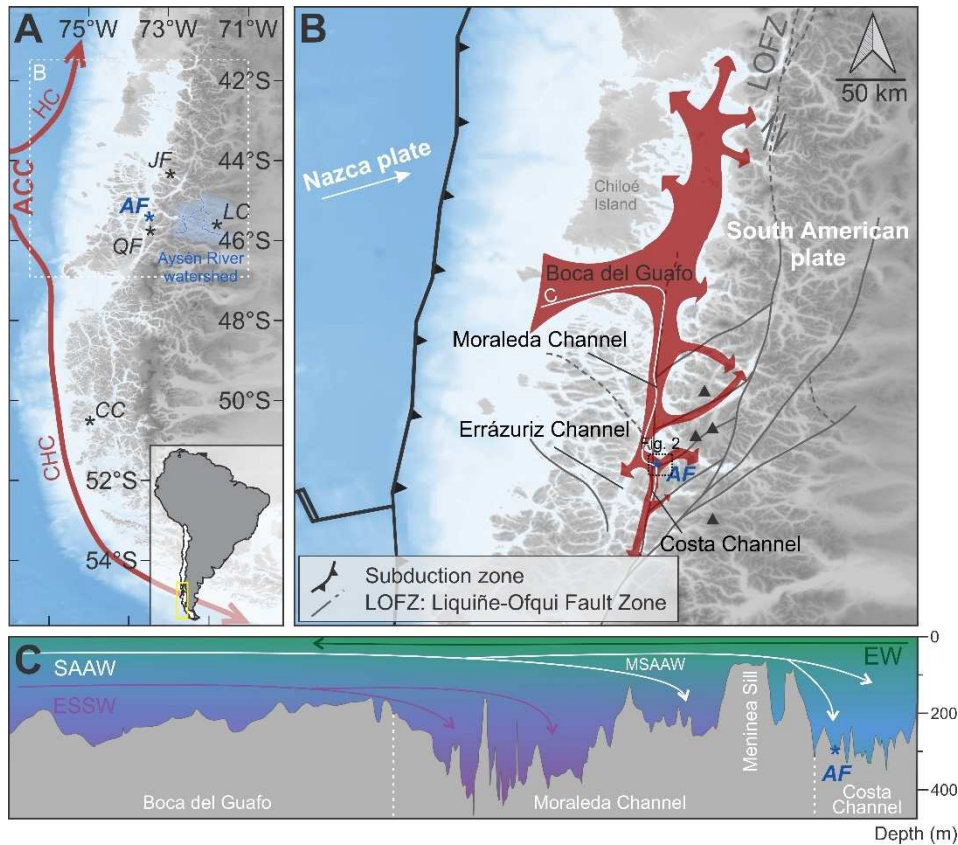
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## 808 **Figure captions**

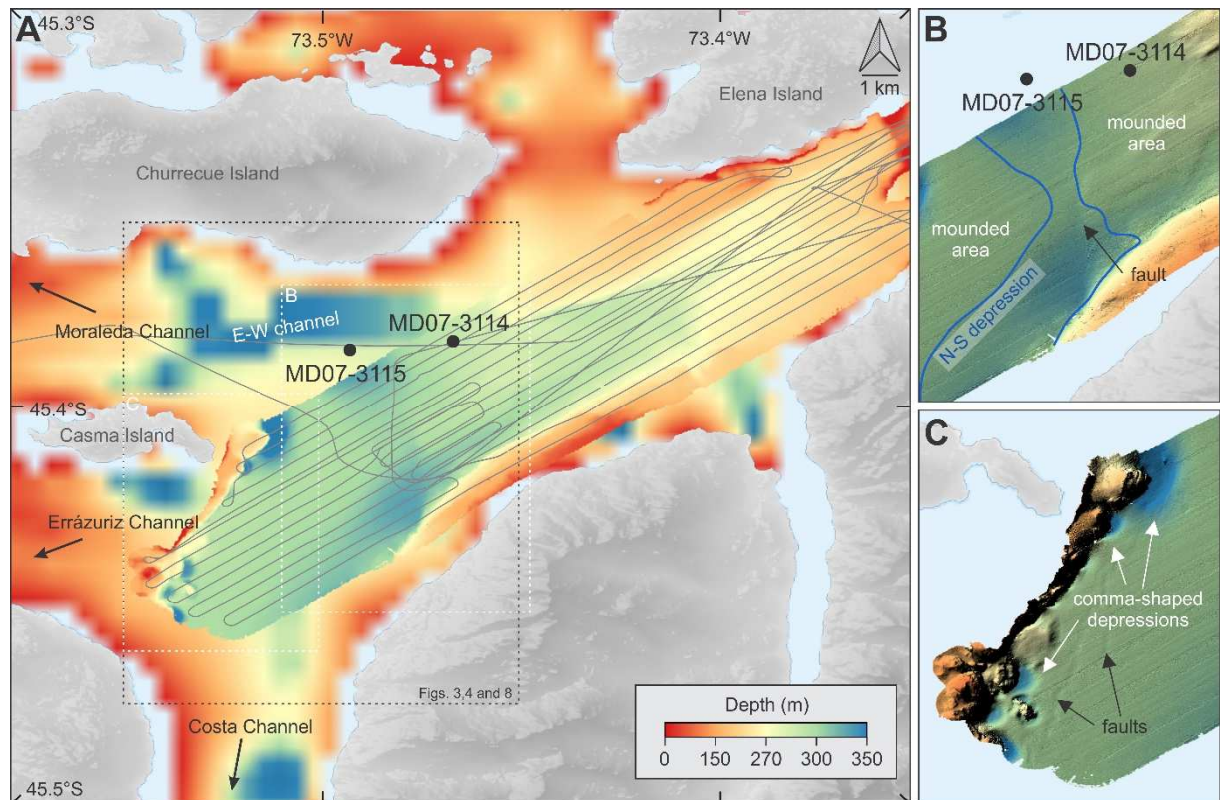




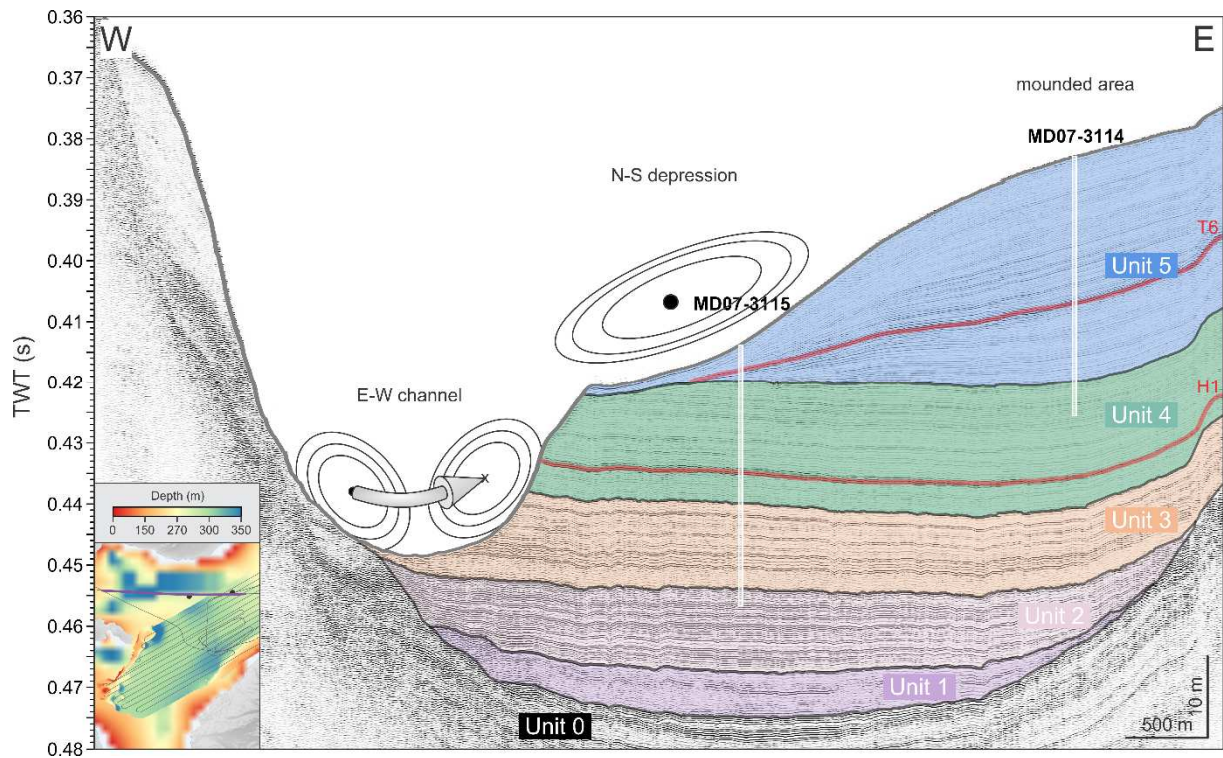
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810 Fig. 1: Tectonic and oceanographic setting of Aysén Fjord (AF), located in southern Chile. A) The  
 811 regional oceanic circulation pattern is dominated by the Antarctic Circumpolar Current (ACC). The  
 812 latter hits the continent between about 40-45°S, where it splits into the northward Humboldt Current  
 813 (HC) and the southward Cape Horn Current (CHC). Additional sites used in this study are indicated: JF  
 814 = Jacaf Fjord, LC = Lago Castor, QF = Quitrusco Fjord and CC = Concepción Channel. B) Close-up on the  
 815 study area, with indication of the approximate trace of the Chilean subduction zone, where the Nazca  
 816 plate subducts beneath the South American plate. The trench-parallel component of oblique  
 817 subduction is accommodated by the presence of the Liqueñe-Ofqui Fault Zone (LOFZ), of which several  
 818 fault branches intersect Aysén Fjord. The four main volcanoes in the vicinity of Aysén Fjord are  
 819 indicated by black triangles (from north to south: Mentolat, Cay, Macá, and Hudson Volcano). The  
 820 circulation pattern in the Patagonian fjords is indicated in red. The latter enters the fjords through the  
 821 Boca del Guafo where it is split into two branches. The southern branch enters the Moraleda channel,  
 822 continuing southwards through the Errázuriz and Costa Channel, the latter passing by Aysén Fjord  
 823 (Sievers and Silva, 2008). C) Cross-section showing the bathymetry (GEBCO Compilation Group, 2020)

824 and the different water masses in the region following the white line in panel B (Sievers and Silva,  
 825 2008): ESSW = Equatorial Subsurface Water (purple), (M)SAAW = (Modified) Subantarctic Water (blue)  
 826 and EW = estuarine water (green). Continued southward flow of the ESSW is blocked by the presence  
 827 of the Meninea sill.



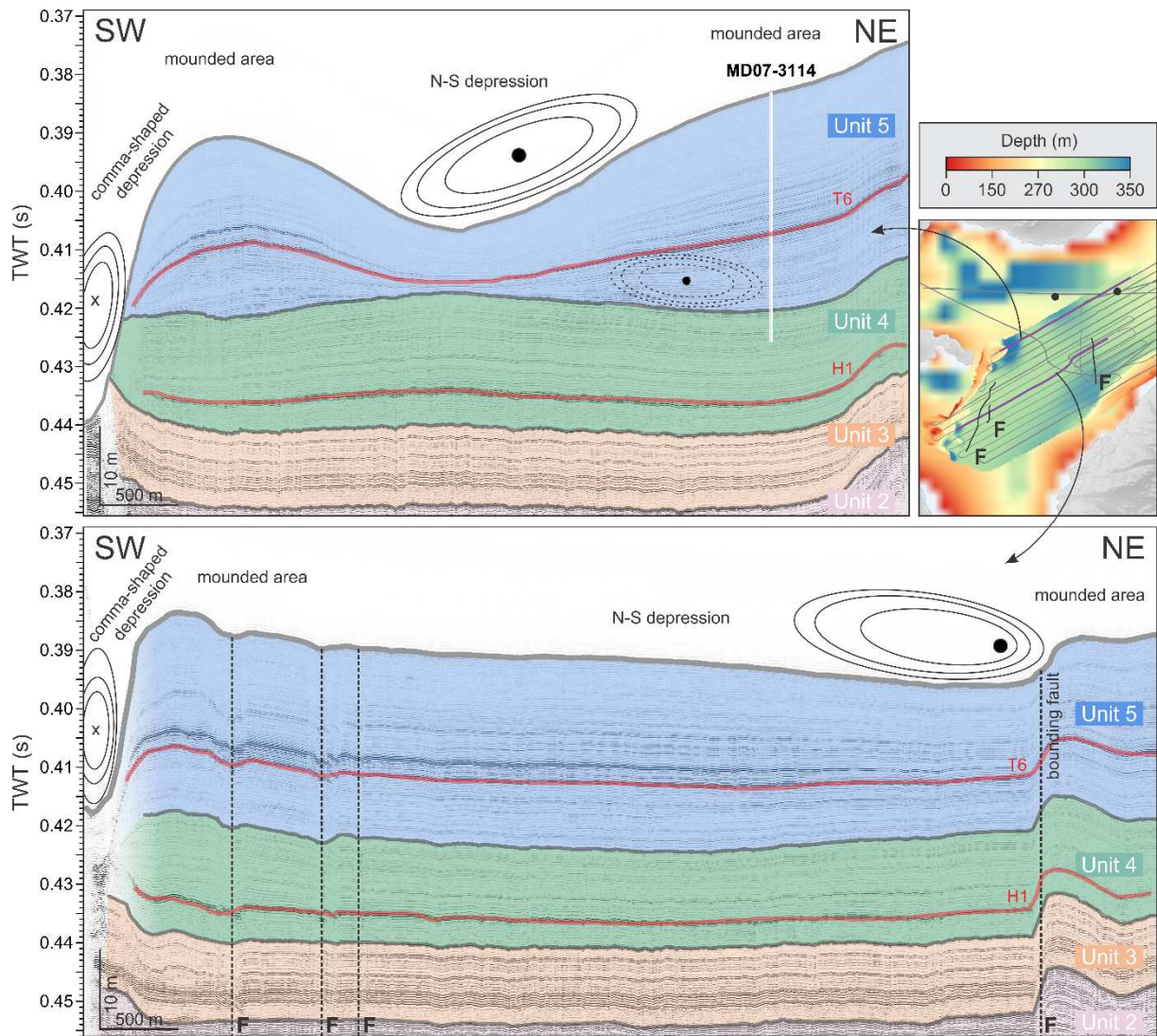
828  
 829 Fig. 2: Bathymetric setting of the study area. A) Multibeam bathymetric map of the outer part of Aysén  
 830 Fjord plotted on a GEBCO bathymetry map (GEBCO Compilation Group, 2020), showing the TOPAS  
 831 seismic lines (grey) and the two core locations (black dots, MD07-3114 and MD07-3115). The apparent  
 832 offset in water depth between both bathymetric maps is mainly the result of a resolution difference.  
 833 A major east-west oriented channel is present just north of both core locations. Detailed visualization  
 834 of other relevant features visible on the multibeam bathymetry are presented in panels B and C, with  
 835 hill shading. B) A north-south oriented elongated depression is located at the longitude of core location  
 836 MD07-3115, in between two mounded areas. A fault trace can be identified on the eastern side of this  
 837 depression. C) Several small, individual depressions with a comma-like shape can be observed in the  
 838 western extremity of the fjord, near Casma Island. Surficial fault traces can also be identified.



839

840 Fig. 3: Interpreted TOPAS seismic profile 209\_002 (vertical exaggeration 50 times), expressed in two-  
 841 way travel time (TWT), with indication of the interpreted current direction at multiple locations. Note  
 842 that these all represent the same current, flowing quasi-parallel to the profile orientation in the erosive  
 843 E-W channel (illustrated by the arrow) and (partially) deflecting southward at the location of MD07-  
 844 3115. A projected length for both sediment cores assuming an acoustic velocity of 1,500 m/s (Chen  
 845 and Millero, 1977) is indicated. The location of the profile (purple line) and cores (black dots) with  
 846 respect to the fjord is given in the lower left corner (see Fig. 2). Six seismic-stratigraphic units can be  
 847 defined (Unit 0-5) and are indicated by colour-coding. The two tephra layers (T6 and H1) are indicated  
 848 in red. An uninterpreted version of this profile can be found in supplementary info (Fig. S1).

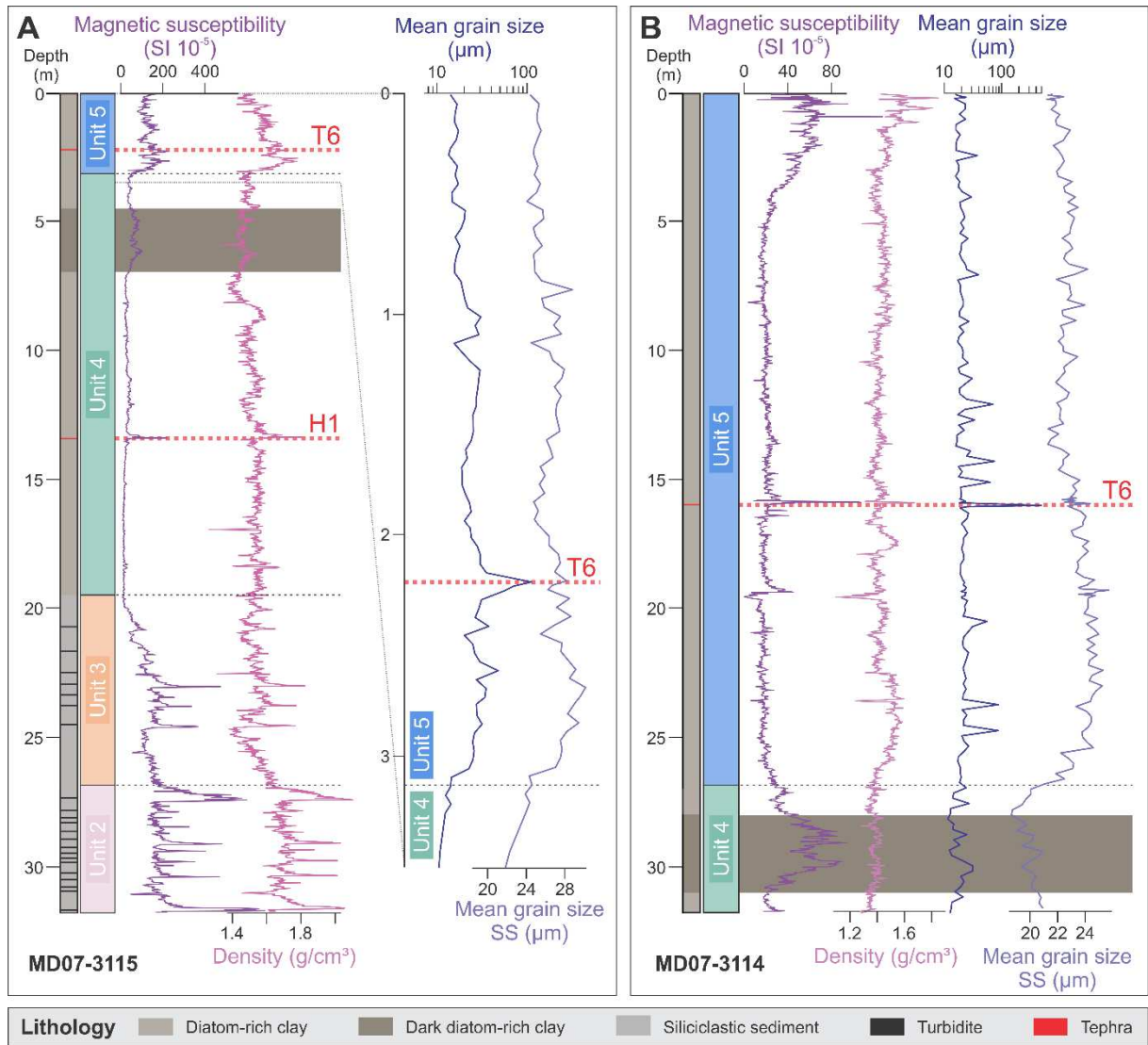




849

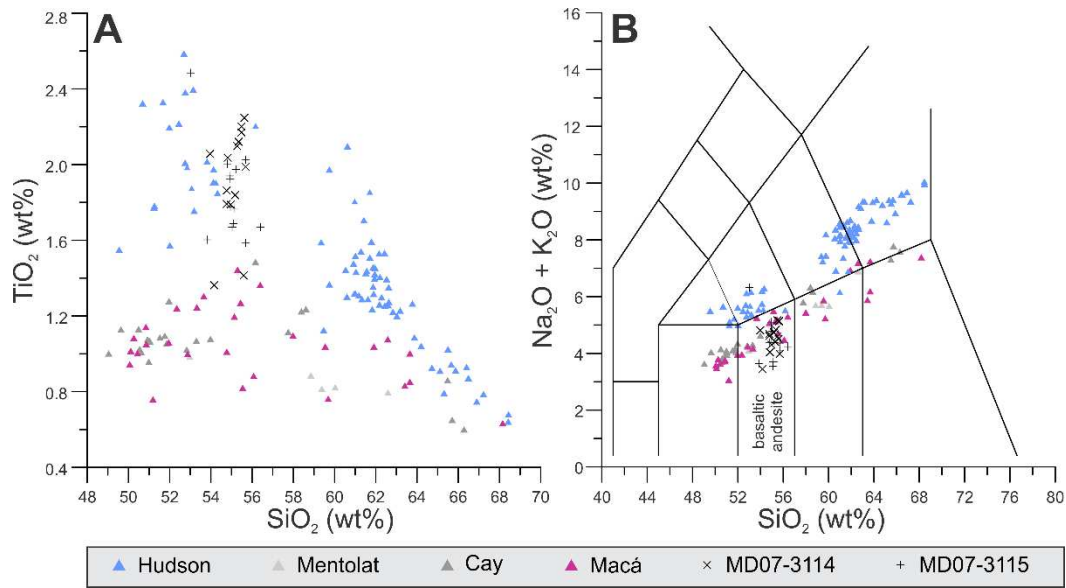
850 Fig. 4: Interpreted TOPAS seismic profiles 187 (upper panel) and 173 (lower panel) (vertical  
 851 exaggeration 50 times), expressed in two-way travel time (TWT), with indication of the inferred current  
 852 directions (dashed when less certain). A projected length for core MD07-3114 assuming an acoustic  
 853 velocity of 1,500 m/s (Chen and Millero, 1977) is indicated in the upper profile. The location of both  
 854 profiles (purple lines) and cores (black dots) with respect to the fjord is given in the upper right corner  
 855 (see Fig. 2). Seismic-stratigraphic Units 2-5 are indicated by colour-coding, similar to those used in  
 856 . Faults (F) are indicated in black on the profiles (dashed lines) and on the bathymetry (full lines). The  
 857 two tephra layers (T6 and H1) are indicated in red. An uninterpreted version of these profiles can be  
 858 found in supplementary info (Fig. S2).





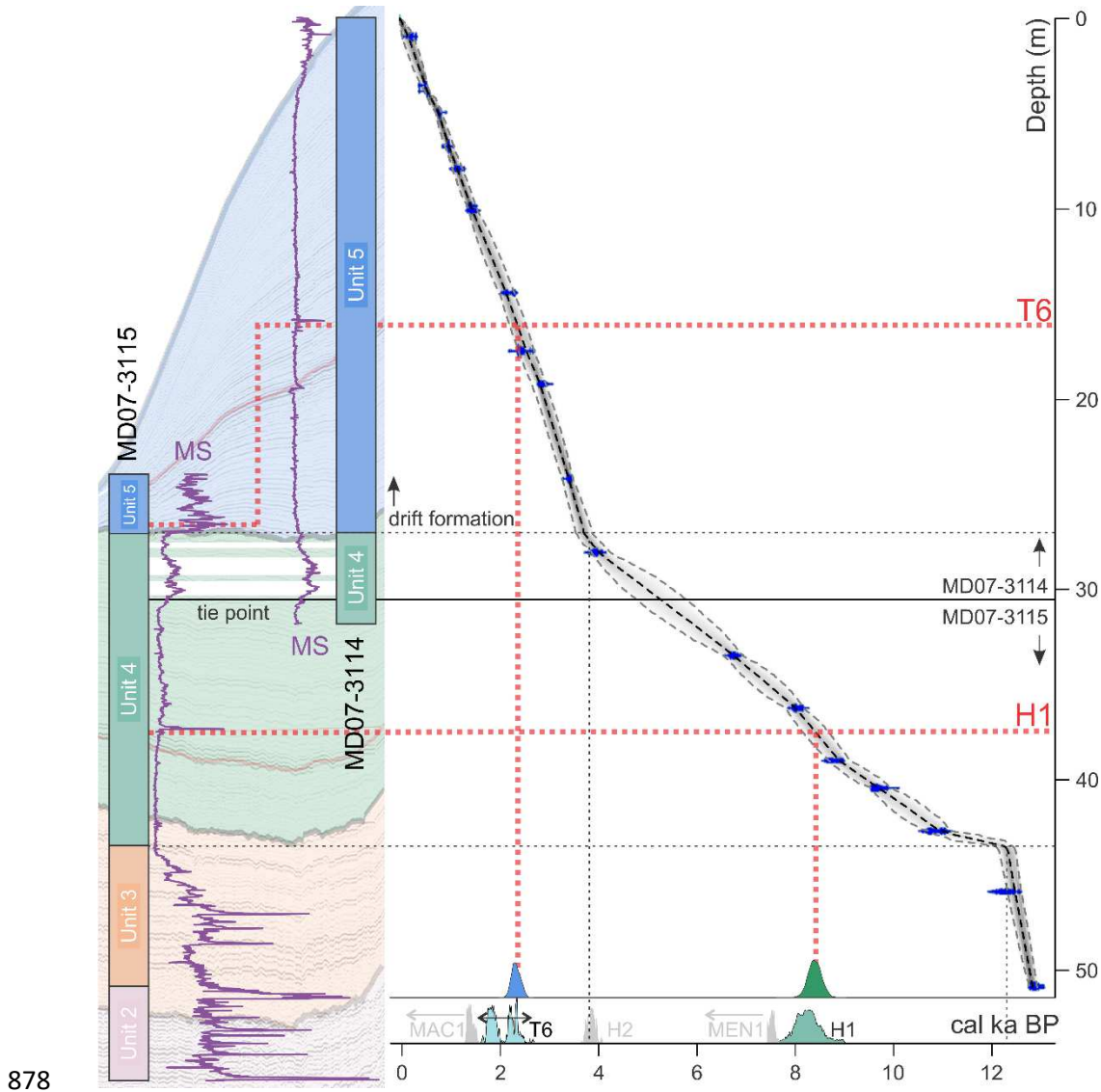
859

860 Fig. 5: Sedimentological characteristics of core MD07-3115 (A) and MD07-3114 (B). The sedimentary  
 861 characteristics are indicated by a schematic lithology representation (diatom-rich clays in brown,  
 862 siliciclastic sediment in grey, turbidites in black, tephra in red), presented along-side the magnetic  
 863 susceptibility (purple), density (pink), mean grain size (dark blue), and sortable silts (SS, light blue)  
 864 values. For core MD07-3115, only the grain-size results for the upper 3.5 m of sediment are presented.  
 865 The darker-coloured sediment interval within Unit 4, marked by two peaks in magnetic susceptibility  
 866 and present in both cores, is indicated in dark grey. The sedimentological variability in both cores can  
 867 be related to seismic-stratigraphic Units 2 to 5, indicated by colour-coding according to  
 868 . The two tephra layers (T6 and H1) are indicated in red.



869

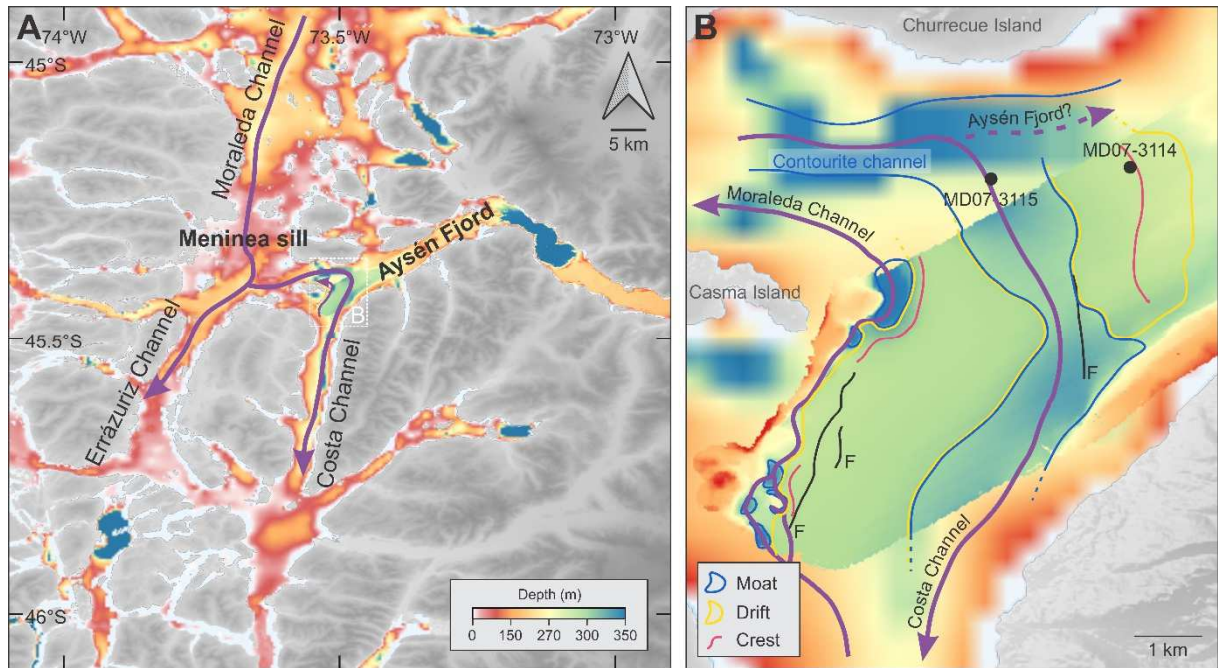
870 Fig. 6: Major-element concentrations of glass shards in a tephra layer (T6) present in both cores. Both  
 871 show a very similar composition, and are considered to result from the same eruption. Comparison to  
 872 whole-rock composition of Mentolat, Cay, Macá, and Hudson volcanoes and volcanic glass of Hudson  
 873 Volcano (D'Orazio et al., 2003; Gutiérrez et al., 2005; Haberle and Lumley, 1998; Kratzmann et al., 2009;  
 874 Naranjo and Stern, 1998) shows that the Ti-content of this tephra layer (A) points to a Hudson Volcano  
 875 origin, while the total alkali content (B) advocates for an eruption of any of the other volcanoes. The  
 876 Si-content excludes an origin related to any of the less evolved monogenetic cones in the vicinity of  
 877 the fjord.



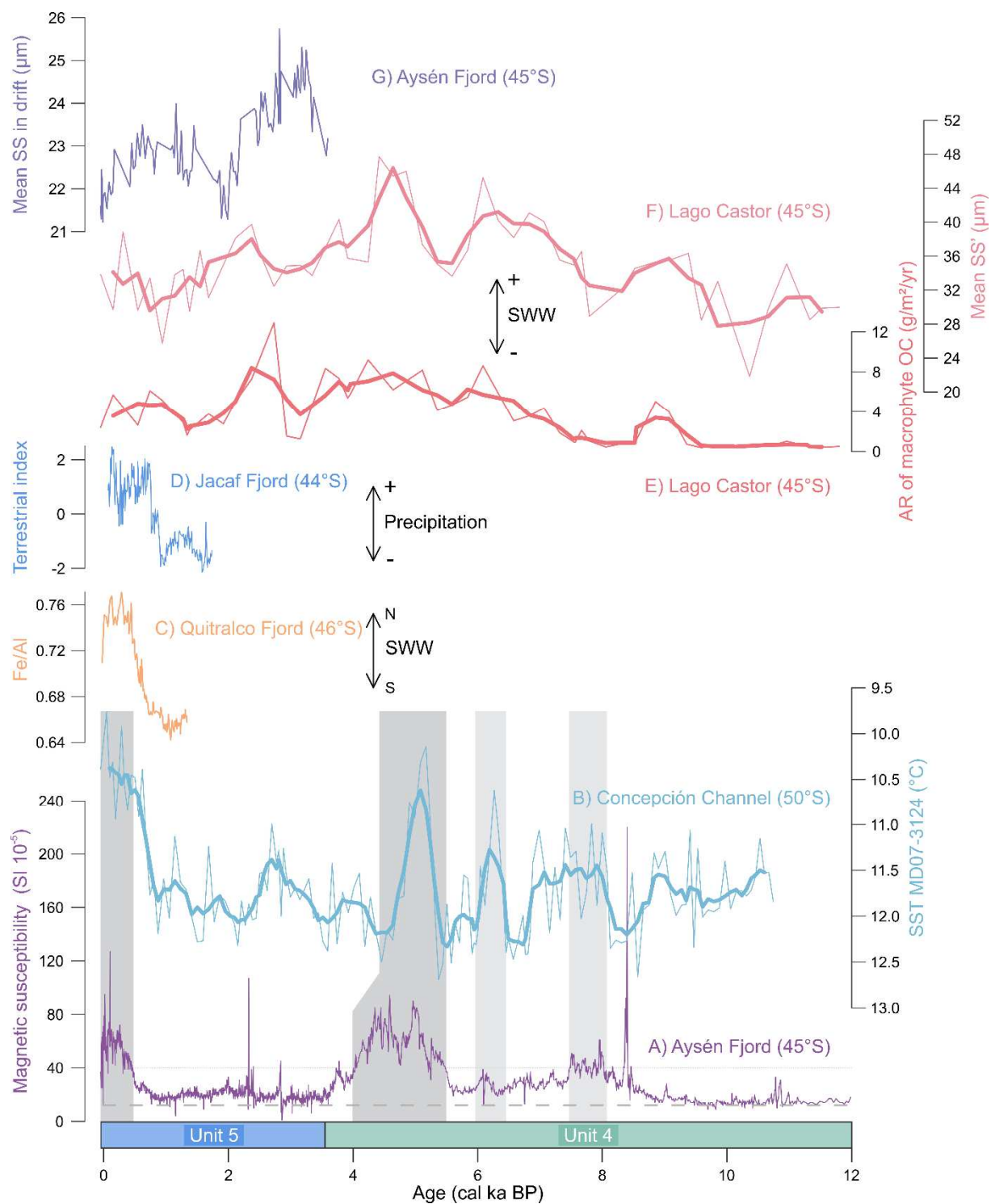
878

879 Fig. 7: Age-depth model obtained for the 'composite' core, covering the sediment drift as well as the  
 880 older sedimentation in the fjord. Left: core-to-core correlation, pinpointed by three simultaneous  
 881 increases (white beams) in magnetic susceptibility (MS) values (purple). The tie point in Unit 4 used to  
 882 construct the composite core is indicated. Both cores are schematically represented by their  
 883 sedimentary units, which are correlated to the seismic units as shown by seismic line 209\_002 in the  
 884 background (colour-coded according to  
 885 ). Right: age model (95% confidence interval in grey) for both sediment cores constructed using  
 886 calibrated radiocarbon ages (blue). Two changes in sedimentation rate can be observed (dashed lines):  
 887 at the boundary between Unit 3 and 4 and between Unit 4 and 5, the latter corresponding to the onset

888 of drift formation. The age range of two tephra layers (T6 and H1, red) in the cores (blue and green)  
 889 are indicated and compared to the ages of volcanic eruptions (bottom axis) described in literature  
 890 (Haberle and Lumley, 1998; Naranjo and Stern, 1998, 2004; Stern and Weller, 2012).



891  
 892 Fig. 8: Present-day current patterns of the (M)SAAW in the Patagonian fjords (purple). A) Visualization  
 893 of the current patterns in the surroundings of Aysén Fjord, derived from our seismic and bathymetrical  
 894 data in combination with the flow pattern suggested by Sievers and Silva (2008) indicated on a GEBCO  
 895 bathymetric map (GEBCO Compilation Group, 2020). B) Bathymetric map of the outer part of Aysén  
 896 Fjord (see Fig. 2) with indication of sediment drifts (moats in blue, drifts in white, drift crest in yellow)  
 897 and the contourite channel, showing a detailed pattern of inflowing and outflowing MSAAW. Faults  
 898 are indicated in black (F).



899

900 Fig. 9: Comparison of sedimentological data for the composite core in Aysén Fjord with regional  
 901 records of sea surface temperature (SST), Southern Westerly Wind (SWW) strength and/or latitudinal  
 902 position and precipitation. From bottom to top: A) magnetic susceptibility in Aysén Fjord (composite  
 903 core, this study), B) alkenone-based SST in Canal Concepción (Caniupán et al., 2014), C) ICP-OES-based  
 904 Fe/Al counts in Quitralco Fjord (Bertrand et al., 2014), D) terrestrial index (PCA score) based on bulk

905 organic geochemistry in Jacaf Fjord (Sepúlveda et al., 2009), E) accumulation rate (AR) of macrophyte  
906 organic carbon (OC) in Lago Castor (Fiers et al., 2019), F) mean modified (10-125  $\mu\text{m}$ ) sortable silt (SS')  
907 in Lago Castor (Van Daele et al., 2016), and G) mean sortable silt (SS) grain size in Aysén Fjord  
908 (composite core, this study, see Fig. 1A for locations). The extent of Units 4 and 5 are indicated, of  
909 which the transition is marked by the onset of sediment drift formation.



910 **Table captions**

911 Table 1: Overview of all radiocarbon ages with their depth in the MD07-3114 and MD07-3115 cores as well as the composite depth used for age-depth  
 912 modelling derived from correlation of both cores. All ages were calibrated using the SHCal20 calibration curve (Hogg et al., 2020). The reservoir age of the  
 913 marine samples is estimated at 550 years (Serno, 2009), to which we add an uncertainty of 40 years.

Core ID	Sample Number	Depth (cm)	Composite depth (cm)	Material	14C age (years BP)	2σ calibrated age (years BP )	Relative probability (%)	Reservoir age (years)
MD07-3114	1	95	95	leaf fragments	215 ± 30	0-24	4.4	
						72-83	2	
						101-113	2.2	
						139-229	62.9	
						242-298	23.4	
	2	351	351	leaf fragments	420 ± 30	327-380	31.3	
						385-403	4.3	
						439-501	59.4	
	3	382	382	scaphopods	1040 ± 25	804-867	41.3	550 ± 40
						900-936	39.7	
940-957						13.9		
4	493	493	leaf fragments	895 ± 30	682-705	7.7		
					718-799	84.2		
					871-879	1.9		
5	671	671	leaf fragments	1070 ± 35	891-896	1		
					805-866	17.3		
					901-980	73.8		
						1029-1047	3.9	

	6	789	789	bivalve	1810 ± 35	1586-1747 1773-1784 1797-1810	91.1 1.8 2	550 ± 40
	7	984	984	leaf fragments	1550 ± 35	1312-1431 1437-1480 1498-1511	82.4 10.1 2.5	
	8	1008	1008	scaphopods	2120 ± 45	1928-1971 1986-2137 2272-2291	11 81.3 2.7	550 ± 40
	9	1440	1440	bivalve	2760 ± 30	2755-2881 2905-2919	92.6 2.3	550 ± 40
	10	1745	1745	leaf fragments	2410 ± 35	2181-2195 2209-2223 2320-2503 2505-2513 2593-2613 2639-2695	1.3 0.9 81.3 0.5 3.4 7.7	
	11	1919	1919	leaf fragments	2790 ± 55	2754-2968 2979-2995	93 1.9	
	12	2416	2416	leaf fragments	3220 ± 35	3267-3287 3335-3456 3473-3480	3.5 90.3 1.2	
	13	2808	2808	bivalve	4210 ± 30	4579-4603 4605-4605 4610-4834	5.3 0.2 89.5	550 ± 40
MD07-3115	14	940	3349	bivalve	6510 ± 35	7280-7299 7306-7431 7449-7467	2.6 89.4 2.8	550 ± 40



15	1216	3625	bivalve	7810 ± 55	8412-8649 8672-8696	93.2 1.8	550 ± 40
16	1492	3901	bivalve	8520 ± 40	9437-9539	95	550 ± 40
17	1636	4045	bivalve	9300 ± 40	10282-10571	95	550 ± 40
18	1862	4271	bivalve	10105 ± 50	11322-11599 11601-11779 11800-11834	49.2 42 3.8	550 ± 40
19	2180	4589	scaphopods	11000 ± 50	12765-18978 12986-13013 13038-13065	87.6 3.6 3.7	550 ± 40
20	2672	5081	wood	11000 ± 40	12766-12975 12991-13000 13042-13052	93 0.9 1	
21	2683	5092	wood	11008 ± 50	12768-12979 12984-73017 13038-13067	84.7 5.4 4.9	

914