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# 1 Post–glacial tephrochronology record off the Chilean 2 continental margin (~41° S)

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## 12 Abstract

13 The Southern Volcanic Zone of the Andes (~33–46° S) is a very active volcanic zone with several  
14 volcanic centers recording recurrent historical activity (e.g. Llaima, Villarrica, Puyehue-Cordón  
15 Caulle, Osorno, Calbuco and Hudson). Tephrochronology is a valuable tool to help better  
16 understand the eruptive history of volcanic centers, essential for producing volcanic hazard maps.  
17 Additionally, tephrochronology can also be very useful to synchronize stratigraphic records of  
18 different nature such as paleoclimatological, paleoceanographical and archaeological records on  
19 land, lakes, ice and the ocean. Here we present a (crypto) tephrochronological record from two  
20 marine sediment cores retrieved in the Chilean continental margin at ~41° S and ~41.6° S. The

21 records display continuous sedimentation since the late glacial, as robustly constrained by  
22 planktonic foraminifera  $\delta^{18}\text{O}$  and  $^{14}\text{C}$  dates. During this period, in total twenty three cryptotephra  
23 and two thepras were identified as glass shard peaks together with two ~25–30 cm-thick visible  
24 tephra (one in each core). The source of the (crypto) tephra was mainly constrained by major and  
25 trace element geochemistry of individual glass shards together with their stratigraphic position,  
26 since physical characteristics such as color and grain size are not possible to observe when analyzing  
27 cryptotephra. From these, one cryptotephra was robustly correlated with the HW7 eruption from  
28 the Hudson volcano occurring in the Late Holocene at ~1.5 cal ka BP; and the two visible tephra  
29 layers were identified as distant correlatives of the Lepu  tephra originating from Michinmahuida  
30 volcano and occurring in the Deglaciation/Holocene transition at around 11 cal ka BP. Additionally,  
31 nine cryptotephra occurring at ~3.6, 6.2, 7.0, 8.5, 9.6, 14.2, 15.9 and 18.2 cal ka BP were robustly  
32 identified as sourced from Michinmahuida volcano but where otherwise not correlated, providing  
33 novel evidence of pre Holocene activity of this volcanic center.

34 Keywords: Post–glacial, Quaternary, South America, Southern Volcanic Zone, Sedimentology-marine  
35 cores, Tephrochronology, Radiocarbon, Major and trace elements.

## 36 **1. Introduction**

37 Tephrochronology is a powerful tool not only to unveil the eruptive history of a territory,  
38 but to provide a robust chronological framework for its stratigraphic record. By identifying  
39 the pyroclastic remains of specific eruptions (tephra) in different sites, the eruptive history  
40 of a determined volcanic center can be reconstructed: its recurrence in time, the dispersion  
41 of its products and the magnitude of the different eruptive events. At the same time, the  
42 remains of the eruptions can be viewed as stratigraphic time markers, which provide

43 regional chronological tie points for different records. When an explosive eruption occurs,  
44 the ejected pyroclastic material is deposited in different environments (land, lakes, ice  
45 caps, and the ocean) and by identifying its remains, the chronologies of paleoceanographic,  
46 paleoclimatological and archaeological records can be aligned, which is crucial when  
47 interpreting the evolution of complex systems, such as the climate.

48 The Southern Volcanic Zones of the Andes (SVZ) is a very active volcanic zone, composed of  
49 at least 60 active volcanic centers between 33° S and 46° S (Figure 1), many of which have  
50 had recurrent explosive volcanic activity in post-glacial times (~20 cal ka BP), such as  
51 Mocho-Choshuenco (Rawson et al., 2015), Chaitén (Alloway et al., 2017b; Amigo et al.,  
52 2013; Watt et al., 2013), Hudson (Haberle and Lumley 1998; Naranjo et al., 1993; Naranjo  
53 and Stern, 1998; Kratzmann et al., 2008; Carel et al., 2011; Weller et al., 2014; Weller et al.,  
54 2015); and/or in historical times, such as Llaima (Naranjo and Moreno, 1991, 2005; Reubi et  
55 al., 2011; Schindlbeck et al., 2014), Puyehue-Cordón Caulle (Alloway et al., 2015; Bertrand  
56 et al., 2014; Gerlach et al., 1988; Lara et al., 2006; Naranjo et al., 2017; Singer et al., 2008;  
57 Villarosa et al., 2006) and Calbuco (Morgado et al., 2019; Romero et al., 2016; Sellés and  
58 Moreno, 2011; Watt et al., 2011b). Since many of these volcanic centers are located nearby  
59 populated areas, volcanology together with tephrochronology become key in  
60 reconstructing their eruptive history, from which volcanic hazard maps are produced (e.g.  
61 Bertin et al., 2018; Moreno and Naranjo, 2002). At the same time, this recurrent explosive  
62 volcanic activity, together with the close proximity of the volcanic centers (Figure 1), have  
63 produced a continuous and intricate tephrochronological record in the area (Fontijn et al.,  
64 2016), which is entangled within stratigraphic records on land, lakes, the ocean, peat bogs,

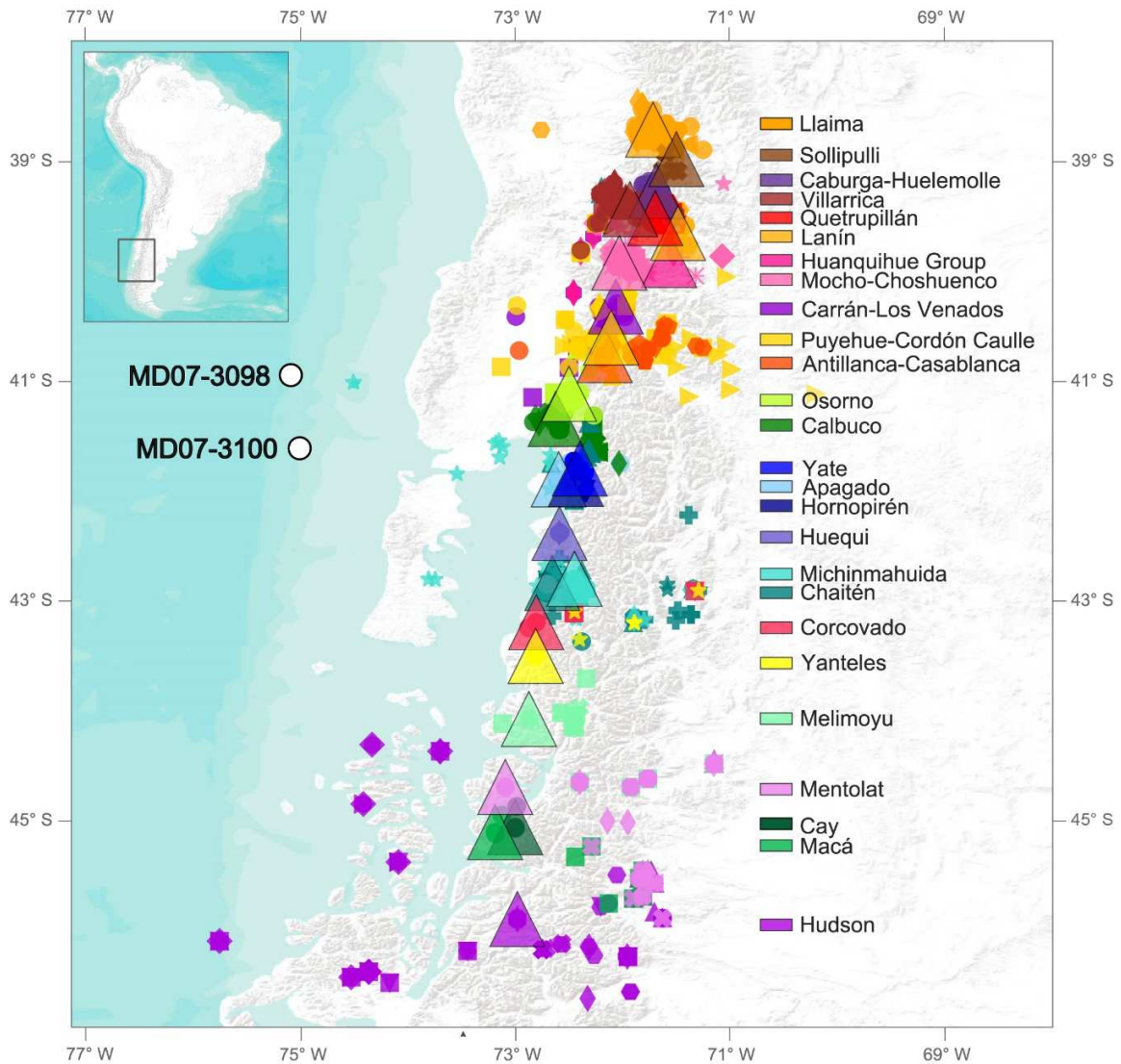
65 etc. The latter has the potential of providing robust chronologies for paleo environmental  
66 records in Patagonia, crucial for understanding the behavior of global-scale climatic  
67 features thought to be important factors controlling glacial–interglacial transitions, such as  
68 the Southerly Westerly Winds (e.g. Menviel et al., 2018; Toggweiler et al., 2006) and  
69 upwelling in the Southern Ocean (e.g. Anderson et al., 2009; Siani et al., 2013).

70 In the last 40 years, the study of volcanology together with tephrochronology have  
71 increased importantly our understanding of the eruptive history of the SVZ since the late  
72 glacial period, revealing high recurrence and explosivity of previously thought to be much  
73 less active volcanic centers, such as Chaitén (Alloway et al., 2017b; Amigo et al., 2013;  
74 Iglesias et al., 2011; Moreno et al., 2014; Watt et al., 2013). However, it has also become  
75 apparent that there are still many volcanic centers in the area, such as Yate, Apagado,  
76 Hornopirén (Mella, 2008; Watt et al., 2011b), Yanteles and Corcovado (López-Escobar et  
77 al., 1993; Naranjo and Stern, 2004) for which the eruptive history is largely unknown.  
78 Efforts towards increasing our knowledge of the volcanic activity in the SVZ include  
79 undertaking field work in previously relatively less researched areas (Watt et al., 2011a, b)  
80 and increasing the research around specific particularly active volcanic centers such as  
81 Puyehue-Cordón Caulle (Alloway et al., 2015; Bertrand et al., 2014; Gerlach et al., 1988;  
82 Lara et al., 2006; Naranjo et al., 2017; Singer et al., 2008) and Hudson (Kratzmann et al.,  
83 2008; Naranjo et al., 1993; Naranjo and Stern, 1998; Weller et al., 2014). Understanding  
84 the eruptive history of this area has additionally benefited from the study of records such  
85 as peat (e.g. Stern, 2008; Weller et al., 2019), lake (Bertrand et al., 2008; Fontijn et al.,  
86 2014; Moreno et al., 2014; Watt et al., 2011b) and marine sediment cores (Carel et al.,

87 2011). These records provide particularly relevant information in areas where the on land  
88 record of the eruptions is not available, either because it is of difficult access, it is in a highly  
89 vegetated area (such as volcanoes south of  $\sim 42^\circ$  S), or because it has been eroded, for  
90 example by the presence of the Patagonian Ice Sheet (PIS), which covered most of the SVZ  
91 south of  $\sim 38^\circ$  S during the late glacial period (Davies et al., 2020). Additionally, if  
92 continuous sedimentation occurs, paleoenvironmental signals such as palynological or  
93 marine isotopic curves can be obtained from lake or sediment cores, which together with  
94  $^{14}\text{C}$  ages provide a more comprehensive context in which the tephras were deposited.

95 Because of the latter, marine sediment cores retrieved in the Chilean continental margin  
96 have great potential of providing relevant information to help complete the  
97 tephrochronological record in the area. However, as a result of the predominant wind  
98 pattern, it is expected that most eruptions leave rather thin ash deposits within marine  
99 sediments as opposed to potent tephra layers, which are routinely studied on land.  
100 Cryptotephrochronology identifies the remains of past volcanic eruptions which are  
101 invisible to the naked eye, and thus can be used to identify the remains of SVZ eruptions in  
102 the Chilean continental margin. Here we use both a tephrochronological and  
103 cryptotephrochronological approach to find evidence of past volcanic eruptions in two  
104 marine sediment cores retrieved in the Chilean continental margin (Figure 1): core MD07-  
105 3098 ( $\sim 40.93^\circ$  S,  $\sim 75.03^\circ$  W, 3055 m) and core MD07-3100 ( $\sim 41.60^\circ$  S,  $\sim 74.95^\circ$  W, 1609 m).  
106 Since physical characteristics of tephras such as grain size, thickness and color, cannot be  
107 used in the identification of the volcanic source of the cryptotephras, their characterization  
108 is based on the geochemical composition (major and trace elements) of individual glass

109 shards together with the stratigraphic position of the cryptotephra, given by planktonic  
 110 foraminifera  $\delta^{18}\text{O}$  and  $^{14}\text{C}$  ages.



111

112 Figure 1. Position of marine sediment cores MD07-3098 (~40.93° S, ~75.03° W, 3055 m) and MD07-3100 (~41.60° S,  
 113 ~74.95° W, 1609 m), here studied. Additionally, volcanic centers of the Southern Volcanic Zone of the Andes which have  
 114 been active since at least the late glacial are shown. Only volcanic centers between Llaima and Hudson volcanoes are  
 115 considered, given the position of the cores. Also shown are the position of tephrochronological samples which have been  
 116 analyzed for geochemistry and/or dated. Each volcanic center and associated analyzed sample is assigned a particular  
 117 color and each eruptive event a particular symbol (detailed legend and references in Figure S16).

## 118 2. Volcanological context

119 The SVZ is one of the four volcanic zones that have been defined in the Andes (Northern  
120 (3–6° S), Central (15–27° S), Southern (33–46° S) and Austral (49–55° S); Stern, 2004),  
121 which, together with the northern and central zones, derive from the subduction of the  
122 Nazca plate under the South American plate. In particular, the SVZ corresponds to the  
123 volcanic activity between Tupungatito (~33° S) and Hudson (~46° S) volcanoes, where so far  
124 ~60 active volcanic centers have been identified, many of which have recorded recurrent  
125 explosive activity in post-glacial times (Global Volcanism Program, Smithsonian Institution.  
126 The volcanic products in the SVZ are characterized by a geochemical composition typical of  
127 subduction zones: calc-alkaline trends with medium to high-K, relatively high mobile  
128 incompatible element concentrations (Cs, Rb, K, Sr, Ba, Pb, Th, U), low REE and HFSE (Zr, Hf,  
129 Ta, Nb).

130 Among the more active and explosive volcanic centers in the area are Puyehue-Cordón  
131 Caulle, Calbuco, Michinmahuida and Hudson. Puyehue-Cordón Caulle, situated at ~40.6° S  
132 (Figure 1), consists of the Puyehue volcano, which has been active since at least ~300 ka,  
133 and the Cordón Caulle fissure, active at least since ~170 ka (Singer et al., 2008). The  
134 products from Puyehue-Cordón Caulle range from basaltic to rhyolitic with medium/low-K  
135 to medium/high-K affinities (Alloway et al., 2015; Bertrand et al., 2014; Gerlach et al.,  
136 1988; Lara et al., 2006; Naranjo et al., 2017; Singer et al., 2008). The most recent explosive  
137 events of Puyehue-Cordón Caulle occurred in 1921-1922, 1960 and 2011, the last of which  
138 had a Volcanic Explosivity Index (VEI) of 5 (Naranjo et al., 2017), and all of which erupted  
139 mainly rhyolitic products. Other eruptive events of similar magnitude have been described



140 in the literature during the Holocene, such as Mil Hojas, Puyehue 2/PCC2 and Puyehue  
141 1/PCC1 (Fontijn et al., 2016; Lara et al., 2006; Naranjo et al., 2017; Singer et al., 2008); and  
142 many effusive products corresponding to deglacial times have been dated (Singer et al.,  
143 2008) although no explosive correlatives have been identified yet.

144 The Calbuco volcano, situated at  $\sim 41.3^\circ$  S, has also recorded continuous volcanic activity  
145 since around 300 ka (Sellés and Moreno, 2011) and extensive historical activity (e.g. 1792,  
146 1893, 1917, 1929, 1932, 1945, 1961, 1972, 2015) the last of which had a VEI of 4 (Romero  
147 et al., 2016). As with Puyehue-Cordón Caulle, Holocene tephras have been more  
148 thoroughly described (Watt et al., 2011b), but evidence of explosive activity during the  
149 Deglaciation has also been found and dated (Sellés and Moreno, 2011). Most products  
150 from Calbuco are andesites or basaltic andesites that have medium to low-K geochemical  
151 affinities (López-Escobar et al., 1995, 1992; Sellés and Moreno, 2011).

152 The Michinmahuida volcano is located at  $\sim 42.8^\circ$  S in a little populated area and has no  
153 record of historical activity, however recurrent Holocene activity has been proposed in the  
154 literature (Amigo et al., 2013; Moreno et al., 2014). One of the most widespread tephra in  
155 the area is associated with Michinmahuida: the Amarillo Ignimbrite, also called the Lepué  
156 tephra (Alloway et al., 2017a; Amigo et al., 2013). The eruption that produced this tephra  
157 occurred at the beginning of the Holocene ( $\sim 11$  cal ka BP) and erupted bimodal products  
158 ranging from basaltic andesites to rhyolites with medium to high-K affinities and has been  
159 estimated to have had a VEI of 6 (Amigo et al., 2013).

160 The Hudson volcano is located at  $\sim 46^\circ$  S in a little populated area difficult to access, as  
161 Michinmahuida volcano. Contrastingly, more information can be found about the Hudson  
162 volcano than for Michinmahuida, probably because of its highly explosive post-glacial and  
163 historical activity, notably, its 1991 eruption with a VEI 4 (Naranjo et al., 1993). Products of  
164 Hudson range from basaltic andesites to rhyolites with medium to high-K affinities (Carel et  
165 al., 2011; Haberle and Lumley, 1998; Kratzmann et al., 2008; López-Escobar et al., 1993;  
166 Naranjo et al., 1993; Naranjo and Stern, 1998; Weller et al., 2014, 2015) and at least 8  
167 tephras have been described and dated in the literature (Ho, HW1, HW2, HW3, H1, H2,  
168 HW6 and HW7), the oldest one: Ho, occurred around  $\sim 17$  cal ka BP with an estimated VEI  
169 of 6 and has been identified as far as  $\sim 900$  km SE of the volcano (Weller et al., 2014).

170 The presence of the cryptotephra layers in marine cores, retrieved west of the volcanic  
171 centers can be somewhat counterintuitive because the dispersion of volcanic ash in the SVZ  
172 is largely controlled by the prevailing Southerly Westerly Winds in the area. From what is  
173 observed in the literature, but also in modern eruptions such as Chaitén, 2008 (Watt et al.,  
174 2009) and Puyehue Cordón Caulle, 2011 (Bertrand et al., 2014), most of the ash produced  
175 by explosive eruptions in the central ( $37\text{--}42^\circ$  S) and southern ( $42\text{--}46^\circ$  S) zones of the SVZ  
176 are transported to the east by the Southerly Westerly Winds, the Subtropical jet and/or the  
177 Polar Front Jet (Gallego et al., 2005; Rahn and Garreaud, 2014). A lesser amount of the  
178 eruptions record a predominantly northward dispersion, which is associated with the  
179 southeast Pacific subtropical anticyclone transporting ashes to the north (Rahn and  
180 Garreaud, 2014). Among these, Cha1 from Chaitén (Watt et al., 2013; Fontijn et al., 2016),  
181 Neltume from Mocho-Choshuenco (Fontijn et al., 2016; Rawson et al., 2015) and the

182 Hudson eruption in 1991 (Naranjo et al., 1993). Less frequently, winds to the south and to  
183 the west also occur as part of smaller scale phenomena, such as low pressure systems and  
184 changes associated to the Southern Hemisphere's baroclinic annular mode (Pérez-Santos et  
185 al., 2019), for example during the 2015 eruption of Calbuco (Romero et al., 2016). The  
186 latter might explain the less common occurrence of tephra deposits to the west and  
187 northwest of volcanoes, such as HW1–7 (Haberle and Lumley, 1998) and Lepué Tephra  
188 (Alloway et al., 2017a). Thus, even though ash transported by these short-lived wind  
189 directions might not be enough to deposit potent tephra in the ocean in most cases, they  
190 are enough to deposit thin ash layers and thus leave traces of past volcanic eruptions  
191 identifiable via cryptotephrochronology.

## 192 **3. Methods**

### 193 **3.1 Core retrieving**

194 Marine sediment cores MD07-3098 (~40.93° S, ~75.03° W, 3055 m) and MD07-3100  
195 (~41.60° S, ~74.95° W, 1609 m) are two CALYPSO cores retrieved by the French R/V Marion  
196 Dufresne during the PACHIDERME expedition in February 2007 (Kissel and The Shipboard  
197 Scientific Party, 2007). MD07-3098 is a 20.74 m long core composed mainly of silty clay  
198 with silty and sandy layers horizons, which also contains a visible tephra layer of ~30 cm  
199 between ~7.6–8 m labeled T6/98. MD07-3100 is a 29.8 m long core composed mainly of  
200 homogenous silty clays, which displays a visible tephra layer of ~25 cm between 7.25–7.5 m  
201 labeled T8/100. For this study, the first 13.5 m and 15 m from core MD07-3098 and MD07-  
202 3100 (respectively) have been analyzed, which correspond to the post-glacial portion of

203 these sedimentary archives. Because of the difference in lithology given by the presence of  
204 the tephra, both cores present a ~25 cm gap in the upper part of the visible tephra (Figure  
205 2), which are corrected for in the age models.

### 206 **3.2 Stable isotope stratigraphy**

207 The stratigraphic constraint of the cores is provided by stable oxygen isotope data ( $\delta^{18}\text{O}$ )  
208 measured in planktonic foraminifera at the Laboratoire des Sciences du Climat et  
209 l'Environnement (LSCE) on Optima VG, GV Isoprime and Finnigan Delta + mass  
210 spectrometers. Isotopic measurements were made every 10 cm in one to thirty specimens  
211 of planktonic foraminifera *G. bulloides* (250–315  $\mu\text{m}$  fraction) in both cores. This  
212 information allowed for a first order identification of the timeframe the cores span (Figure  
213 2). The measurements stopped where a stable late glacial isotopic signal was reached, at  
214 ~13.5 m in core MD07-3098 and ~15 m in core MD07-3100. Results from core MD07-3100  
215 have already been published by Haddam et al. (2018) and those for core MD07-3098 are  
216 displayed in Table S1.

### 217 **3.3 Tephra and Cryptotephra identification**

218 Cryptotephra layers were identified as peaks in the relative amount of glass shards and  
219 micro-pumice. The cores were sampled every 10 cm in 4 cc box samples, washed, sieved  
220 for the fraction  $>40 \mu\text{m}$  and rinsed in a 10% HCl solution to dilute the carbonated fraction  
221 and allow for the identification only of the detrital fraction. The percentage of glass shards  
222 and micro-pumice was obtained by counting the relative amount of glass shards, micro-  
223 pumice, crystals, lithics and other detrital material in the detrital fraction  $>40 \mu\text{m}$  in at least

224 300 particles per sample. Primary glass shards and micro-pumice were identified as having  
225 pristine angular morphologies with fragile tips and no remineralization coatings (Figure S1).  
226 In most depths glass shards color range from a brown-honey color to white, with diverse  
227 degrees of vesicularity. The glass shard and micro-pumice peaks were considered where  
228 the percentage exceeded the background value at least three times. For core MD07-3100,  
229 115 samples were counted each 10 cm between 60 and 1220 cm, the background was  
230 around 5% and sixteen glass shard and micro-pumice peaks were identified as  
231 cryptotephra and analyzed for geochemistry with peak values between 15 and 45%. For  
232 core MD07-3098, 86 samples were counted with a resolution between 10 and 20 cm, from  
233 1 to 1321 cm depth, the background was 6% and seven glass shard peaks were identified as  
234 cryptotephra and analyzed for geochemistry with peak values between 18 and 27%.

235 In both cores, the visible tephra are recognized by an important change in lithology easily  
236 recognizable as a change from olive brown silty clay to black sand (Figure 2). Both tephra  
237 present a sharp erosive contact at the base and graded at the top. Because of the graded  
238 upper contact, the upper limit of the tephra were identified as an important drop in the  
239 glass shard and micro-pumice count in both cores. Additionally, as mentioned in section  
240 3.1, both cores present a gap in the upper part of the tephra deposits, thus the thickness of  
241 each tephra is estimated as the sum of the thickness above and below each gap.

### 242 **3.4 Geochemistry**

243 In order to characterize the geochemistry of the cryptotephra and visible tephra,  
244 individual glass shards from the identified layers were hand-picked and mounted in epoxy  
245 resin beads and polished on an automated polish wheel to avoid compositional variations

246 due to surficial alteration processes. Major element were determined for each individual  
247 glass shard with a CAMECA-SX 100 Electron Microprobe (EPMA-CAMPARIS) at the  
248 University Paris VI (France). Ten elements were analyzed (Na, Mg, Si, Al, Cl, K, Ca, Ti, Mn  
249 and Fe) using an accelerating voltage of 15 KV, a current of 10 nA and a beam size of 5  $\mu\text{m}$   
250 to minimize loss of alkalis such as Na and the results are presented in Table S2. The  
251 instrument calibration was done using natural mineral standards (e.g. Na is calibrated from albite, K  
252 from Orthose, Si, Ca and Mg from diopside, Fe from hematite). Precisions on individual shards ( $1\sigma$ )  
253 were better than 0.6% for Si,  $\sim 1\%$  for Al, 3% for Ca and Mg, 4% for Na, 5% for Fe, 6% for K, 10% for  
254 Ti and about 30% for Cl and Mn. The in-house glass standard Lipari was also analyzed as an  
255 unknown sample during the analytical sessions (Table S4) and display comparable values to  
256 published data (Jochum et al., 2007). As seen on Figure S1, glass shards from different cryptotephra  
257 layers and even in a single layer (T8/100) might have very different aspects, such as highly  
258 vesiculated and microlite-rich glass shards coexist with non-vesiculated microlite-poor glass  
259 shards. Attention was paid to analyze glass shards with the less microlites possible and between  
260  $\sim 10$  and 30 analyses were acquired for each cryptotephra. Major element data obtained  
261 here, as well as those from the literature, were normalized to 100% before plotting in order  
262 to being able to compare the analyses.

263 Additionally, trace element concentrations were measured in ten of the sixteen depths in  
264 core MD07-3100 with a Laser Ablation High Resolution ICP-MS (LA-HR-ICPMS) at GEOPS  
265 laboratory (Paris-Saclay University). Results are presented in Table S2. For the ablation, a  
266 CETAC-Teledyne Excimer 193 nm laser system was connected to an Element XR HR-ICP-MS  
267 and the sample was transported to the torch using He gas. A frequency of 5Hz and a

268 fluence of 3.65 J/cm<sup>2</sup>, for a spot size of 40 microns were used during analyses. The  
269 instrument was calibrated using the international glass standard NIST 612 as an external  
270 standard. In order to cover the large range of silica contents determined by EMPA for the  
271 glass shards, international glass standards of rhyolitic (ATHO-G; 75.6 wt.% SiO<sub>2</sub>) or basaltic  
272 (KL2-G; 50.3 wt.% SiO<sub>2</sub>) compositions were also run as unknown samples to control the  
273 quality of the measurements (Table S4). The reported values in Table S4 compare well with  
274 the preferred values from GEOREM (Jochum et al., 2007). For the purpose of the study,  
275 internal standards KL2-G and ATHO-G gave a precision (1σ) on La between 2.4 and 3.7%,  
276 between 2.7 and 5.2% for Yb, between 3.2 and 3.7% for Zr and between 2 and 2.5% for Nb  
277 (Table S4).

### 278 **3.5 Radiocarbon dating**

279 After identifying the glass shard and micro-pumice peaks, planktonic foraminifera were  
280 picked either at the base or in the middle of the peak, according to foraminifera peak  
281 abundances. Radiocarbon dates for core MD07-3100 are published in Haddam et al. (2018)  
282 whereas new radiocarbon dates for core MD07-3098 are presented in Table S5. Either  
283 mono specific *G. bulloides* or mixed planktonic foraminifera were picked from the >150 μm  
284 fraction, samples >3 mg were measured at UMS-ARTEMIS (Pelletron 3 MV) AMS facilities  
285 (CNRS-CEA Saclay, France, Dumoulin et al., 2017; Moreau et al., 2013), whereas samples <3  
286 mg were analyzed on the gas ion source ECHOMICADAS at the LSCE (Gif-Sur-Yvette, France)  
287 using coupling of cracker system to the gas handling system (Wacker et al., 2013; Tisnérat-  
288 Laborde et al., 2015). <sup>14</sup>C results are reported in conventional age BP according to the  
289 convention of Stuiver and Polach (1977), normalized to the base of δ<sup>13</sup>C of −25.0‰ relative

290 to the Pee Dee Belemnite (PDB) international standard, and corrected by the age of the  
291 background subtraction. The age model for the cores was then obtained with the software  
292 Undatable (Lougheed and Obrochta, 2019) with a depth uncertainty of 1 cm,  
293 corresponding to the sampling width, and calibrating according to SHCal20 curve (Hogg et  
294 al., 2020).

## 295 **4. Results**

### 296 **4.1 Chronological framework**

297 The obtained stable isotope stratigraphy (Table S1, Figure 2) reveals that both studied  
298 cores cover at least the time interval between the late glacial and the Late Holocene and  
299 register similar variations and sedimentation rates since. The MD07-3100  $\delta^{18}\text{O}$  record  
300 displays steady values around 2.7 ‰ between ~12.5–15 m, which characterizes the late  
301 glacial in the core (Figure 2). A deglacial decreasing trend is observed between ~12.5–7.5  
302 m, when values decrease from ~2.7 to ~0.9‰ interrupted by two plateaus: around ~11 m  
303 and ~8–10 m. Minimum values of ~0.9‰ PDB are reached at ~7.5 m, coincident with the  
304 occurrence of the ~25 cm visible tephra (T8/100). Values characteristic of the Holocene are  
305 observed between ~7.5 m and the top. In particular: around 7 m  $\delta^{18}\text{O}$  values reach its  
306 mean global minimum of around 0.9‰, followed by a steady increase until 2‰ between 6  
307 and 5 m and finally, between ~5–0 m somewhat steady values of around 1.6‰ occur, with  
308 two intervals of lower values between ~3.6–4 and ~0.7–1 m.

309 The late glacial is observed in core MD07-3098 between ~13.5–11.7 m with steady values  
310 around 3‰ (Figure 2). A deglacial decrease is observed between ~11.7–8.1 m. Between



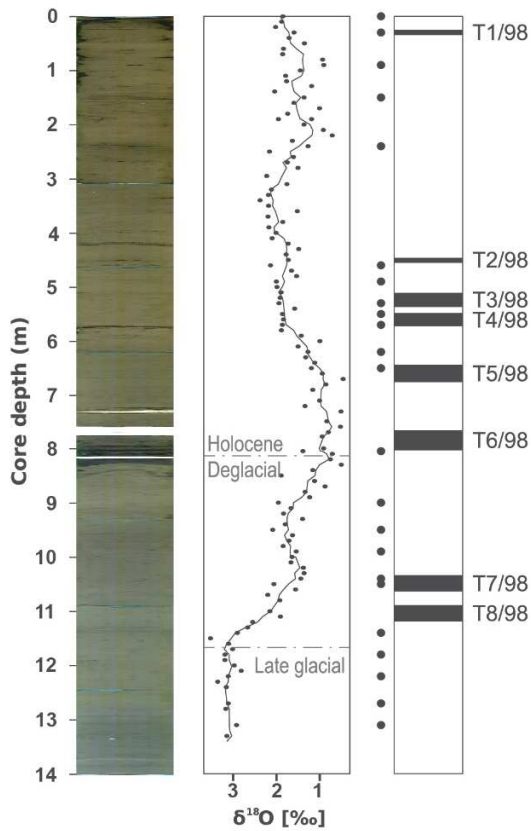
311 ~11.7–10.2 m values decrease until 1.5‰, a slight increase until 1.6‰ is observed between  
312 ~10.2–9 m, followed by a steady decrease until reaching minimum values of ~0.9‰ at 8.1  
313 m, coincident with the presence of a ~30 cm thick tephra layer (T6/98). Minimum values of  
314 ~0.9‰ maintain until 6.6 m. Between ~6.6–5.6 m, values steadily increase until ~1.8‰ and  
315 are maintained at this value until ~3.2 m. From ~3.2 m to the core top values are relatively  
316 lighter and more variable.

317 The age model of core MD07-3098 was obtained from 27 planktonic  $^{14}\text{C}$  ages (Table S5,  
318 Figure S3), corrected by a mobile marine surface reservoir age ( $R_s$ ), following the method  
319 by Haddam et al. (2018) for core MD07-3100. The similarity of the  $\delta^{18}\text{O}$  curves of both  
320 cores versus their planktonic  $^{14}\text{C}$  age (Figure S2) suggests that they record variations of the  
321 same water mass (Subantarctic Water) in the studied time interval. Thus, we apply the  
322 same  $R_s$  correction used for MD07-3100 by Haddam et al. (2018) to the planktonic  $^{14}\text{C}$  ages  
323 from core MD07-3098 (Table S5). The conventional radiocarbon ages were subsequently  
324 calibrated according to the SHCal20 curve (Hogg et al., 2020) and the age model was  
325 obtained with the software Undatable (Lougheed and Obrochta, 2019).

326 From the resulting age model (Table S1, Figure S3, S4), core MD07-3098 has relatively high  
327 and nearly constant sedimentation rates during the Holocene and Deglaciation of ~70  
328 cm/ky. Core MD07-3100 has nearly constant sedimentation rates of ~65 cm/ky during the  
329 whole studied interval (Haddam et al., 2018). The age model together with the stable  
330 isotope stratigraphy indicate that in the studied cores, the late glacial ends between 17–  
331 16.5 cal ka BP (Figure S4), the Deglaciation occurs between ~17–16.5 and ~10.5 cal ka BP  
332 and the Holocene between ~10.5 cal ka BP and the present.

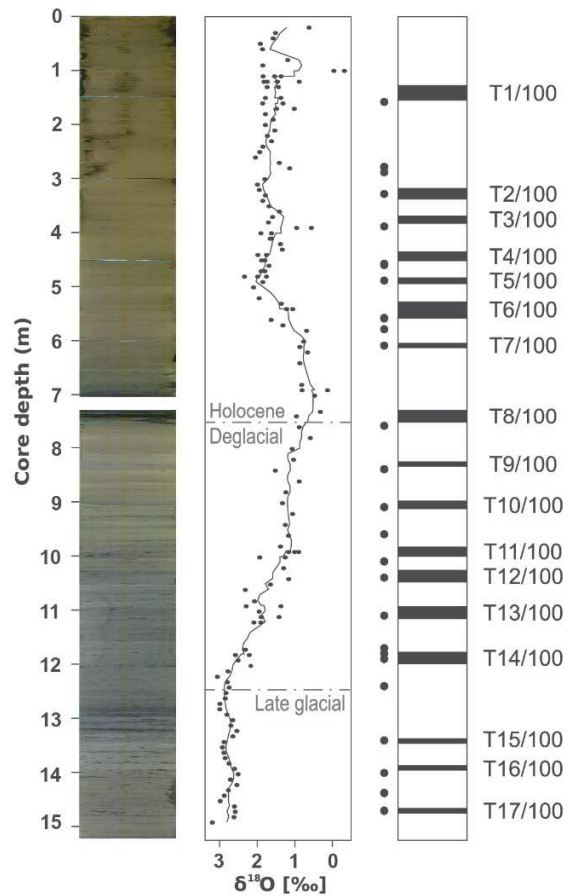
## MD07-3098

40.93° S, 75.03° W, 3055 m



## MD07-3100

41.60° S, 74.95° W, 1609 m



333

334

Figure 2. Stratigraphy from cores MD07-3098 and MD07-3100. Left: Composite photograph of each core, showing the position of the visible tephra as a dark layer at around 8 m (T6/98 in MD07-3098) and 7.5 m (T8/100 in MD07-3100). Middle:  $\delta^{18}\text{O}$  from planktonic foraminifera *G. bulloides*; dots indicate measurements every 10 cm, the curve corresponds to the three point moving average. Also indicated are climate intervals inferred from the observed changes in  $\delta^{18}\text{O}$ . Right: Depths considered as potential cryptotephra layers (relative glass shard peaks) labeled T1/98 to T8/98 (MD07-3098) and T1/100 to T17/100 (MD07-3100), together with the position where planktonic foraminifera  $^{14}\text{C}$  ages were obtained as dark grey dots. Details for  $\delta^{18}\text{O}$  and  $^{14}\text{C}$  measurements in core MD07-3098 in Tables S1 and S5 and for core MD07-3100 in Haddam et al. (2018).

341

342

### 4.2 Cryptotephras

343

In core MD07-3100, sixteen depths where relative glass shard and micro-pumice peaks

344

occurred were recognized as cryptotephra layers labeled T1/100 to T17/100, in addition

345

to T8/100 (Figure 2, Table 1). From the stable isotope stratigraphy, cryptotephra T1/100

346

through T7/100 occur during the Holocene, T8/100 in the Holocene/Deglaciation transition,

347 T9/100 to T14/100 during the Deglaciation and T15/100, T16/100 and T17/100 during the  
348 late glacial.

349 In core M07-3098, seven depths where relative glass shard and micro-pumice peaks occur  
350 were identified as cryptotephra layers labeled T1/98 to T8/98, in addition to the visible  
351 tephra layer T6/8 (Figure 2, Table 1). T1/98 through T5/98 occur in the Holocene portion of  
352 the core, T6/98 in the Holocene/Deglaciation transition and T7/98 and T8/98 during the  
353 Deglaciation.

#### 354 **4.3 Geochemistry of the marine ash layers**

355 All analyzed samples display major element composition belonging to the calc-alkaline  
356 series and most can be classified as medium-K to high-K, with rare glass shards of low-K  
357 affinity (Figure 3, 4, 5, Table S2, 3). Major elements in all samples follow a trend, either  
358 bimodal or continuous, ranging from basaltic andesite (minimum value of 49.8 wt.% SiO<sub>2</sub> in  
359 cryptotephra T8/100) to rhyodacite or rhyolite (maximum 77.8 wt.% SiO<sub>2</sub> in cryptotephra  
360 T9/100).

361 Three groups can be defined based on the relative amount of K<sub>2</sub>O versus SiO<sub>2</sub>:

- 362 • Group 1 (Figure 3, S5): samples with relatively high-K contents (0.7–3.9 wt.% K<sub>2</sub>O),  
363 with basaltic andesitic (52–56 wt.% SiO<sub>2</sub>), andesitic (56–63 wt.% SiO<sub>2</sub>) and dacitic  
364 composition (63–70 wt.% SiO<sub>2</sub>) plot along the boundary between medium and high-  
365 K fields (Peccerillo and Taylor, 1976) and points with rhyolitic composition in the  
366 high-K field (>70 wt.% SiO<sub>2</sub>). The two visible tephtras T6/98 and T8/100 and eleven  
367 cryptotephtras among the twenty three analyzed, fall in this group, notably T1/100  
368 through T7/100, T12/100, T14/100, T15/100 and T17/100;

369 Group 2 (Figure 4, S6): samples with relatively lower K contents than Group 1 (0.4–  
370 4.21 wt.% SiO<sub>2</sub>). Basaltic andesitic and andesitic compositions plot in the medium–K  
371 field and sample points with dacitic and rhyolitic compositions at the boundary  
372 between medium–K and high–K. Eleven cryptotephra belong to this group: T1/98  
373 through T5/98, T7/98, T9/100, T10/100, T11/100, T13/100, T16/100;

- 374 • Group 3 (Figure 5, S7): restricted to cryptotephra T8/98, which has consistently  
375 lower K contents (0.7–1.3 wt.% K<sub>2</sub>O; with the exception of one outlier with high–K  
376 around 4.2 wt.% K<sub>2</sub>O), plotting near the low/medium–K field boundary, especially at  
377 dacitic and rhyolitic composition.

378 Additionally, nine cryptotephra from core MD07-3100 were analyzed for trace  
379 elements: T1/100, T2/100, T4/100, T5/100, T6/100, T7/100, T9/100, T12/100, T14/100  
380 and T15/100 together with tephra T8/100. Overall these samples are relatively enriched  
381 in the most incompatible elements (Rb, Ba, Th, U), have lower HFSE (negative anomaly  
382 of Nb, Ta) and lower REE and Sr contents and display a positive anomaly in Pb. Three  
383 different behaviors are observed when analyzing the La/Yb versus Zr/Nb ratios of the  
384 samples (Figure 3b, 4b): cryptotephra layers T2/100, T4/100, T5/100, T6/100, T7/100,  
385 T8/100, T12/100, T14/100 and T15/100 form a cluster with La/Yb values between ~4–7  
386 and Zr/Nb between ~17–31; T1/100 has relatively higher La/Yb values between ~8–12  
387 and Zr/Nb values between ~20–25; and T9/100 displays more dispersed values, with  
388 La/Yb between ~5–11 and Zr/Nb between ~18–30.

389

| Cryptotephra | Core      | Depth range (cm) | Calendar age median (years BP) | 1 $\sigma$ range | 2 $\sigma$ range | Group   | Source        | Potential correlative |
|--------------|-----------|------------------|--------------------------------|------------------|------------------|---------|---------------|-----------------------|
| T1/98        | MD07-3098 | 25–35            | 294                            | 445–132          | 747–62           | Group 2 |               | YA2?                  |
| T2/98        | MD07-3098 | 445–455          | 6233                           | 6410–5999        | 7053–5856        | Group 1 |               |                       |
| T3/98        | MD07-3098 | 505–535          | 7633                           | 7853–7320        | 8019–7027        | Group 2 |               | PCC2?                 |
| T4/98        | MD07-3098 | 545–570          | 7928                           | 8138–7738        | 8380–7639        | Group 2 |               |                       |
| T5/98        | MD07-3098 | 645–675          | 9356                           | 9581–9124        | 9916–8939        | Group 2 |               | PCC1?                 |
| T6/98        | MD07-3098 | 764–801          | 10512                          | 10702–10296      | 11265–10050      | Group 1 | Michinmahuida | Lepué                 |
| T7/98        | MD07-3098 | 1030–1060        | 13965                          | 14387–13681      | 14919–13546      | Group 2 |               |                       |
| T8/98        | MD07-3098 | 1090–1120        | 15472                          | 15778–15163      | 15991–14596      | Group 3 |               |                       |
| T1/100       | MD07-3100 | 130–160          | 1557                           | 1410–1720        | 1315–1855        | Group 1 | Hudson        | HW7                   |
| T2/100       | MD07-3100 | 320–340          | 3635                           | 3895–3382        | 4118–3141        | Group 1 | Michinmahuida | LTT-10?               |
| T3/100       | MD07-3100 | 370–385          | 4117                           | 4342–3939        | 4424–3745        | Group 1 |               |                       |
| T4/100       | MD07-3100 | 440–460          | 6170                           | 6694–5716        | 6992–5122        | Group 1 | Michinmahuida | LTT14? LTT-15? MIC1?  |
| T5/100       | MD07-3100 | 485–495          | 7044                           | 7214–6838        | 7520–6686        | Group 1 | Michinmahuida | LTT15? MIC1?          |
| T6/100       | MD07-3100 | 530–560          | 8514                           | 8636–8364        | 8812–8228        | Group 1 | Michinmahuida | LTT20?                |
| T7/100       | MD07-3100 | 605–615          | 9603                           | 9703–9478        | 9892–9318        | Group 1 | Michinmahuida |                       |
| T8/100       | MD07-3100 | 728–755          | 10616                          | 10706–10440      | 10938–10348      | Group 1 | Michinmahuida | Lepué                 |
| T9/100       | MD07-3100 | 825–835          | 12163                          | 12464–11941      | 12604–11760      | Group 2 |               |                       |
| T10/100      | MD07-3100 | 895–915          | 12839                          | 12966–12671      | 13151–12575      | Group 2 |               |                       |
| T11/100      | MD07-3100 | 980–1005         | 13812                          | 13906–13672      | 14047–13596      | Group 2 |               |                       |
| T12/100      | MD07-3100 | 1025–1050        | 14154                          | 14296–13950      | 14533–13808      | Group 1 | Michinmahuida |                       |
| T13/100      | MD07-3100 | 1125–1145        | 15253                          | 15493–15048      | 15701.5–14777    | Group 2 |               |                       |
| T14/100      | MD07-3100 | 1170–1190        | 15930                          | 16063–15775      | 16228–15649      | Group 1 | Michinmahuida |                       |
| T15/100      | MD07-3100 | 1340 cm          | 18226                          | 18422–18030      | 18595–17870      | Group 1 | Michinmahuida |                       |
| T16/100      | MD07-3100 | 1390 cm          | 19386                          | 19709–19193      | 19900–18758      | Group 2 |               |                       |
| T17/100      | MD07-3100 | 1470 cm          | 20448                          | 20638–20266      | 20818–20080      | Group 1 |               |                       |

390 Table 1. Summary of identified tephras and cryptotephras in cores MD07-3098 and MD07-3100.

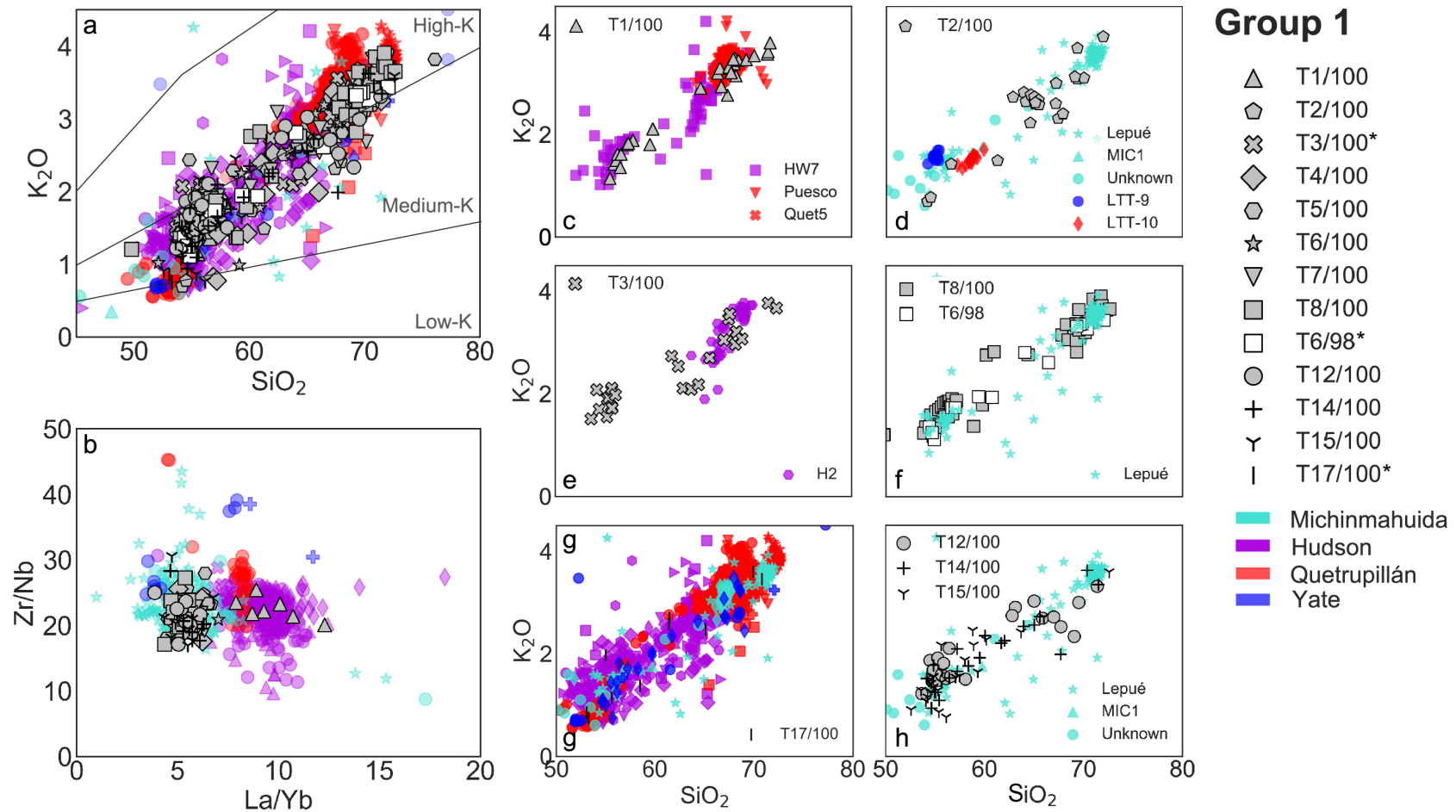
## 391 5. Correlations

392 For the identification of the potential source of the tephtras and cryptotephtras, we  
393 compared the geochemistry of the glass shards here analyzed with the geochemistry of  
394 both explosive and effusive products that were analyzed by both micro analytical and bulk  
395 geochemical methods from volcanic centers between Llaima and Hudson. The latter is  
396 achieved taking into account that for some of the volcanic centers in this area no micro  
397 analytical data were found in the literature (e.g. Yanteles and Corcovado by Naranjo and  
398 Stern (2004)) and that, especially for older eruptions, only effusive products were analyzed  
399 (e.g. Puyehue-Cordón Caulle by (Singer et al., 2008)). Additionally, previous work in the  
400 region has previously shown that a similar geochemical trend is followed by effusive and  
401 explosive products from the Hudson volcano (Carel et al., 2011), which is also observed in  
402 the literature here considered (Figure S8, S9). Most of this literature corresponds to major  
403 element compositions, from which it can be observed that the major element signature of  
404 many volcanoes overlap (Figure S10). Thus major element compositions cannot be used  
405 alone to pinpoint the volcanic source of the cryptotephtras. Nevertheless, the potential  
406 volcanic sources can still be narrowed down based on their  $K_2O$  versus  $SiO_2$  contents. With  
407 this in mind, in the following section, we organize the discussion regarding the potential  
408 source of the tephtras and cryptotephtras according to the three groups previously defined  
409 based on the relative amount of  $K_2O$  versus  $SiO_2$  together with some key trace element  
410 ratios of the analyzed samples, when available (Figure 3, 4, 5). The potential eruptive event  
411 is further constrained based on the stratigraphy and estimated age of the tephtras and  
412 cryptotephtras.

## 5.1 Group 1

413  
414 The thirteen tephras and cryptotephras belonging to Group 1, with relatively higher K (0.7–  
415 3.9 wt.% K<sub>2</sub>O), especially at silica levels higher than ~68 wt.% SiO<sub>2</sub> (Figure 3a, S4), display a  
416 geochemical trend similar to volcanoes Michinmahuida (Alloway et al., 2017a; Amigo et al.,  
417 2013; Naranjo and Stern, 2004), Hudson (Carel et al., 2011; Del Carlo et al., 2018; Haberle  
418 and Lumley, 1998; Kratzmann et al., 2008; López-Escobar et al., 1993; Naranjo et al., 1993;  
419 Naranjo and Stern, 1998; Smith et al., 2019; Stern, 2008; Weller et al., 2015, 2014),  
420 Quetrupillán (Brahm et al., 2018; Fontijn et al., 2016; Hickey-Vargas et al., 1989; Rawson et  
421 al., 2016; Simmons et al., 2020) and Yate (López-Escobar et al., 1993; Mella, 2008; Watt et  
422 al., 2011b). These volcanic centers can be further discriminated by means of their trace  
423 element geochemistry, in particular, their Zr/Nb versus La/Yb ratios (Figure 3b). From the  
424 thirteen tephras and cryptotephras in this group, one tephra and nine cryptotephras have  
425 been analyzed for trace elements: T1/100, T2/100, T4/ 100, T5/100, T6/100, T7/100,  
426 T8/100, T12/100, T14/100 and T15/100. When comparing their Zr/Nb versus La/Yb ratios,  
427 with the available information from Michinmahuida, Hudson, Quetrupillán and Yate, the  
428 tephra (T8/100) and eight of the nine cryptotephra layers (T2/100, T4/100, T5/100, T6/100,  
429 T7/100, T12/100, T14/100 and T15/100) display compositions similar to Michinmahuida  
430 and one cryptotephra (T1/100) displays compositions similar to Hudson. Similar behaviors  
431 are observed when looking at other trace elements (Figure S11), such as Zr versus Y, La  
432 versus Tm and Th versus Nb, although the differences among the different volcanic centers  
433 are not as clear as in the case of La/Yb versus Zr/Nb.

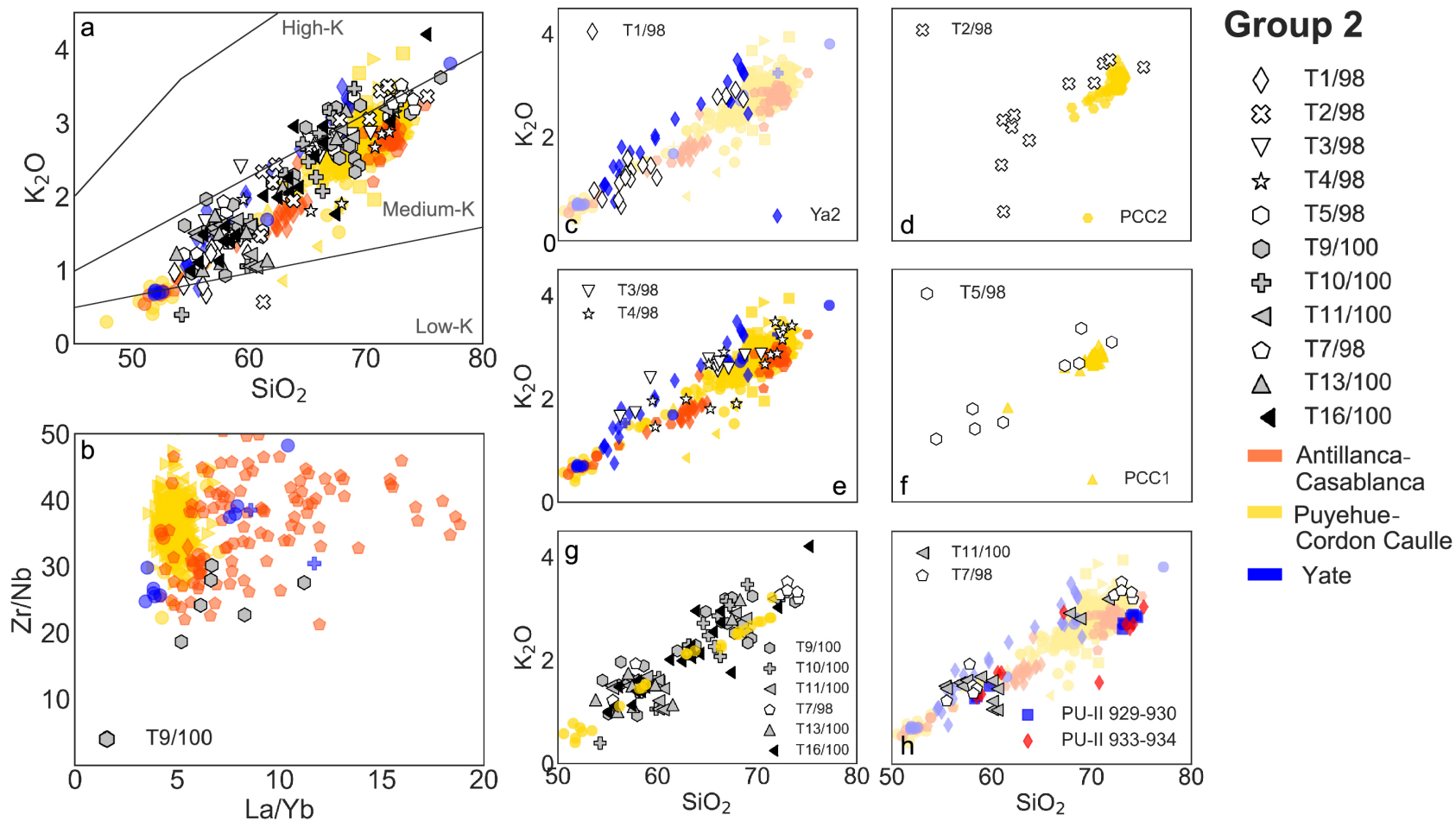
434 In addition, one potential correlation among the marine sediment cores can be established



435

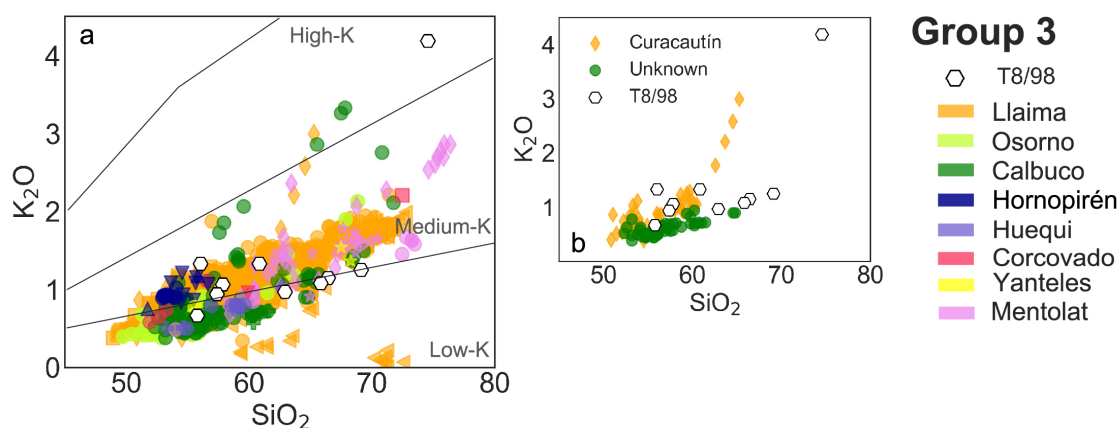
436 Figure 3. Individual glass shard composition from the cryptotephra layers here studied corresponding to Group 1 (legend to the right). The latter are compared with available  
 437 analyses of post-glacial deposits from volcanic centers displaying a similar behavior (color legend to the right): Michinmahuida (Alloway et al., 2017a; Amigo et al., 2013; López-  
 438 Escobar et al., 1993; Naranjo and Stern, 2004), Hudson (Carel et al., 2011; Del Carlo et al., 2018; Haberle and Lumley, 1998; Kilian et al., 2003; Kratzmann et al., 2008; López-  
 439 Escobar et al., 1993; Naranjo et al., 1993; Naranjo and Stern, 1998; Smith et al., 2019; Stern, 2008; Weller et al., 2014, 2015), Quetrupillán (Brahm et al., 2018; Fontijn et al., 2016;  
 440 Hickey-Vargas et al., 1989; Rawson et al., 2016) and Yate (López-Escobar et al., 1993; Mella, 2008; Watt et al., 2011b). “Unknown” corresponds to products not assigned to any  
 441 particular eruption. a: All cryptotephra in Group 1 are plotted, lines separating areas with low-K, medium-K and high-K according to Peccerillo and Taylor (1976) are shown. b:  
 442 Only tephra for which trace elements were analyzed are shown (cryptotephra without an asterisk in the legend to the right). c, d, e, f, g, h: Comparison of specific cryptotephra  
 443 where a potential relative could be identified on land with available geochemical analyses. c: HW7 (Haberle and Lumley, 1998), Puesco and Quet5 (Fontijn et al., 2016). d: LTT-9  
 444 and LTT-10 (Moreno et al., 2014) together with available post-glacial data from Michinmahuida (Lepué, MIC1 and Unknown). e: H2 (Haberle and Lumley, 1998; Naranjo and Stern,  
 445 1998; Smith et al., 2019; Weller et al., 2015). f: Lepué (Alloway et al., 2017a; Amigo et al., 2013). g: same as a. h: available post-glacial data from Michinmahuida (Lepué, MIC1 and  
 446 Unknown). Major element data correspond to normalized volatile-free compositions.





447

448 Figure 4. Individual glass shard composition from the cryptotephra layers here studied corresponding to Group 2 (legend to the right). The latter are compared with available  
 449 analyses of post-glacial deposits from volcanic centers (color legend to the right) displaying a similar behavior: Puyehue-Cordón Caulle (Alloway et al., 2015; Bertrand et al., 2008;  
 450 Fontijn et al., 2016; Gerlach et al., 1988; Lara et al., 2006; Naranjo et al., 2017; Rawson et al., 2016; Singer et al., 2008), Antillanca-Casablanca (Fontijn et al., 2016; Geoffroy et al.,  
 451 2018; Jacques et al., 2014; Naranjo et al., 2017; Rawson et al., 2016; Villarosa et al., 2006) and Yate (López-Escobar et al., 1993; Mella, 2008; Watt et al., 2011b). a: All  
 452 cryptotephtras in Group 2 are plotted, lines separating areas with low-K, medium-K and high-K according to Peccerillo and Taylor (1976) are shown. b: Trace element comparison  
 453 between T9/100 and available information from volcanic centers in Group 2. c, d, e, f, g, h: Comparison of specific cryptotephtras were a potential correlative could be identified on  
 454 land with available geochemical analyses. c: Glass shards analyses from Ya2 (Watt et al., 2011b) are highlighted. d and f: Comparison with glass shard analyses of PCC2 and PPC1 by  
 455 Fontijn et al. (2016) and bulk tephra by Singer et al. (2008), Rawson et al. (2016) and Naranjo et al. (2017). g: Comparison of deglacial and late glacial cryptotephtras with deglacial  
 456 products from Puyehue-Cordón Caulle, whole-rock analyses by Singer et al. (2008) dated by  $^{40}\text{Ar}/^{39}\text{Ar}$  in  $13.2 \pm 2$  and  $18.7 \pm 2.1$  cal ka BP; post-glacial whole-rock analyses by  
 457 Gerlach et al. (1988). h: comparison of T7/98 and T11/100 with tephra PU-II 929-930 and 933-934 cm (Bertrand et al., 2008). Major elements correspond to normalized volatile-free  
 458 compositions.



459

460 Figure 5. Individual glass shard composition from the cryptotephra here studied corresponding to Group 3 (T8/98, legend  
 461 to the right). T8/98 is compared with post-glacial analyses from volcanic centers (color legend to the right) displaying a  
 462 similar behavior: Llaima (Bouvet de Maisonneuve et al., 2012; Naranjo and Moreno, 1991, 2005; Rawson et al., 2016;  
 463 Reubi et al., 2011; Schindlbeck et al., 2014; Lohmar, 2008), Osorno (Bertrand et al., 2008; Jacques et al., 2014; López-  
 464 Escobar et al., 1993; López-Escobar et al., 1992; Moreno et al., 2010; Tagiri et al., 1993), Calbuco (López-Escobar et al.,  
 465 1995, 1992; Morgado et al., 2019; Watt et al., 2011b; Sellés and Moreno, 2011), Huequi, Hornopirén (Watt et al., 2011a,  
 466 b), Mentolat (López-Escobar et al., 1993; Naranjo and Stern, 2004; Stern et al., 2015; Weller et al., 2017, 2015) Yanteles  
 467 and Corcovado (López-Escobar et al., 1993; Naranjo and Stern, 2004). b: comparison with Curacautín tephra products:  
 468 glass shards (Fontijn et al., 2016), matrix glass (Lohmar, 2008) and bulk tephra (Naranjo and Moreno, 2005; Lohmar, 2008;  
 469 Schindlbeck et al., 2014) and eruptive products not assigned to any particular eruption from Calbuco, marked as  
 470 "Unknown" (Sellés and Moreno, 2011). Major elements correspond to normalized volatile-free compositions.

471 for tephtras T6/98 and T8/100 since they display similar major element geochemistry and

472 occur in a similar stratigraphic position (Figure 2, 3f, 6).

473 In the following section we discuss the potential correlative of each cryptotephra based on  
 474 its stratigraphic position, estimated calendar age and available geochemical information on  
 475 land from the potential sources. Because many of the cryptotephtras display a  
 476 Michinmahuida signature, we discuss them together.

#### 477 5.1.1. T1/100

478 The stratigraphic position of this cryptotephra corresponds to the Late Holocene, with an  
 479 estimated mean calendar age at its base of  $\sim 1.6$  ( $1\sigma$ : 1.7–1.4) cal ka BP (Table 1). Around  
 480 this time, two of the volcanic centers associated with Group 1: Hudson and Quetrupillán,  
 481 had eruptions recorded in the literature (Figure 6): HW7 from Hudson ( $\sim 1.5$  cal ka BP;  $1\sigma$ :

482 1.7–1.4; Haberle and Lumley, 1998), Quet5 (~1.7 cal ka BP; 1 $\sigma$ : 1.8–1.6) and Puesco (~1.9  
483 cal ka BP; 1 $\sigma$ : 2–1.8; Fontijn et al., 2016) from Quetrupillán. Individual glass shard major  
484 element geochemistry has been obtained for all of these eruptions and therefore we can  
485 compare their geochemical fingerprints with those of the marine cryptotephra (Figure 3c).  
486 From these, T1/100 displays values closer to HW7 (Carel et al., 2011; Haberle and Lumley,  
487 1998), both of which follow a bimodal trend. In contrast Quet5 and Puesco only display  
488 more evolved products, between 65–68 wt.% SiO<sub>2</sub>, thus the less evolved part of the trend,  
489 observed in T1/100, is absent. Major and trace element data thus support the Hudson  
490 volcano as the source of the T1/100 cryptotephra (Figure 3a, b, c, S7). We propose T1/100  
491 as a distant correlative of the HW7 eruption of the Hudson volcano. The latter is consistent  
492 with prior information indicating a northwest dispersion of some Hudson volcano products  
493 (Figure 6) such as eruptions HW2, HW3, HW6, HW1 and HW7, found in lake cores in Taitao  
494 peninsula and Chonos archipelago (Haberle and Lumley, 1998) as well as in marine  
495 sediment core MD07-3088 (Carel et al., 2011).

#### 496 5.1.2 T3/100

497 Cryptotephra T3/100 occurs in the Late Holocene at an estimated calendar age of ~4.1 (1 $\sigma$ :  
498 4.3–3.9) cal ka BP. During this time period, from the volcanoes in Group 1, only Hudson has  
499 a recorded eruption (Figure 6): H2 at ~4 cal ka BP (Naranjo and Stern, 1998). From the  
500 comparison of the major element geochemistry, it is plausible that T3/100 corresponds to  
501 the eruption H2 (Figure 3e), however H2 corresponds to fairly evolved products ranging  
502 from 64–70 wt.% SiO<sub>2</sub> whereas T3/100 also presents a less evolved cluster between 54–  
503 56% SiO<sub>2</sub> not present in H2, as documented in the literature. As mentioned before, the

504 major element geochemistry of T3/100 is also consistent with Yate, Michinmahuida and  
505 Quetrupillán (Figure S8). The source of this cryptotephra could be further constrained by  
506 trace element analyses and future research around these volcanoes which might reveal  
507 other potential correlatives.

### 508 5.1.3 T17/100

509 T17/100, occurs in the Late glacial at an estimated age of  $\sim 20.5$  ( $1\sigma$ : 20.6–20.3) cal ka BP.  
510 From the major elements alone is not possible to further narrow down among the volcanic  
511 centers in Group 1 as the possible source (Figure 3g, S9). Additionally, little evidence for  
512 tephra deposits at this time exists on land because of the predominant presence of the PIS  
513 in the area until  $\sim 18$  cal ka BP (Davies et al., 2020). From the volcanoes in Group 1, so far  
514 only volcanic products from Yate have been dated so far, by  $^{40}\text{Ar}/^{39}\text{Ar}$  in  $22 \pm 7$  cal ka BP  
515 (Mella, 2008), indicating volcanic activity during this time. No evidence for Michinmahuida,  
516 Hudson or Quetrupillán at this time exists so far, however because of the position of  
517 Quetrupillán north of the sediment core, which is opposite to the predominant winds in the  
518 area, it is less likely that T17/100 was emitted from this volcano.

### 519 5.1.4 Cryptotephtras with Michinmahuida signature

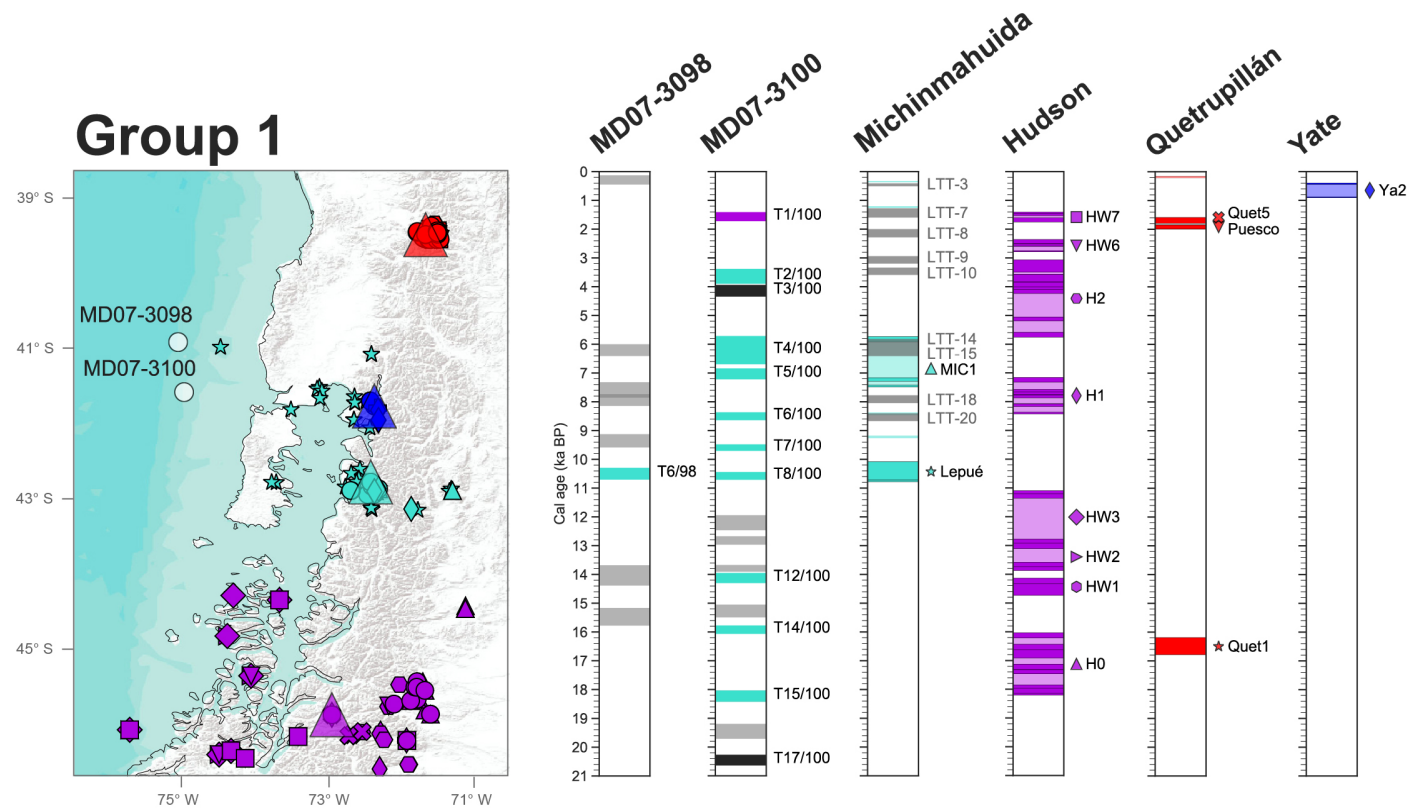
520 We now discuss the potential correlations of the tephra and cryptotephra layers which  
521 have geochemical compositions similar to the Michinmahuida volcano, as discussed above  
522 (T2/100, T4/100, T5/100, T6/100, T7/100, T8/100, T6/98, T12/100, T14/100 and T15/100).

#### 523 5.1.4.1 T2/100

524 The younger cryptotephra with a geochemistry similar to Michinmahuida volcano is T2/100  
525 (Figure 3a, b, d) with an estimated age of  $\sim 3.6$  cal ka BP ( $1\sigma$ : 3.9–3.4). In the literature,

526 evidence for eruptive activity of this volcano at that time has been found in a lacustrine

527 core



528

529 Figure 6. Chronology for cores MD07-3098 and MD07-3100 highlighting the cryptotephra associated to Group 1, as described in the text, compared with the chronologies of  
 530 explosive products from each volcanic center displaying a similar geochemical behavior (Figure 3). Only known events dated by  $^{14}\text{C}$  are plotted. To the left: position of the marine  
 531 sediment cores here studied together with the position of each volcanic center associated to Group 1 indicated as a triangle, from north to south: Quetrupillán, Yate,  
 532 Michinmahuida and Hudson. Also shown are the position of tephrochronological samples which have been analyzed for geochemistry and/or dated as smaller size symbols. Each  
 533 symbol corresponds to a different eruptive event indicated in the respective volcanic center chronology (to the right). Pyroclastic deposits which have been dated but no  
 534 correlation has been yet established are plotted in the chronology as colored rectangles of the corresponding volcanic center color, but they are unlabeled (no symbol or name is  
 535 indicated to their right) and their position is indicated as a dot of the corresponding volcanic center color in the map. These samples are labeled “Unknown” in Table S6. Details on  
 536 the chronologies of each volcanic center in Table S6.

537 ~24 km southwest of Michinmahuida at Lago Teo (~42.9° S, ~72.7° W; Moreno et al., 2014)  
538 and in the Mallines section ~13 km southeast of it (~42.9° S, ~72.3° W; Amigo et al., 2013).  
539 In the Mallines section, Amigo et al. (2013) report a scoria deposit identified in the upper  
540 part of the section (*3<sup>rd</sup> scoria*), bracketed between ~5 and 1.3 cal ka BP, which they  
541 attribute to the Michinmahuida volcano. For this deposit, major and trace elements were  
542 analyzed in the bulk tephra, however the sample has been reported as altered (Amigo et  
543 al., 2013) so further robust correlation is not possible. At Lago Teo, on the other hand, four  
544 tephtras with a Michinmahuida-like major element geochemistry have been identified  
545 between ~5–1.3 cal ka BP by Moreno et al. (2014). From these, the more likely correlatives  
546 to T2/100 correspond to LTT-9 and LTT-10 (Figure 6), which have estimated calendar ages  
547 of ~3 (1 $\sigma$ : 3.2–2.9) and ~3.5 (1 $\sigma$ : 3.6–3.3) ka BP. However, it is not possible to establish a  
548 robust correlation because the majority of the analyzed glass shards in T2/100 correspond  
549 to the more evolved part of the trend with dacitic and rhyolitic compositions, which are not  
550 present in LTT-9 and LTT-10.

#### 551 5.1.4.2 T4/100 and T5/100

552 During the Middle Holocene, two cryptotephtras with a Michinmahuida signature occur in  
553 the marine sediment cores: T4/100 and T5/100. T4/100 has a calibrated age of ~6.2 (1 $\sigma$ :  
554 6.7–5.7) cal ka BP, whereas T5/100 has an age of ~7 (1 $\sigma$ : 7.2–6.8) cal ka BP.

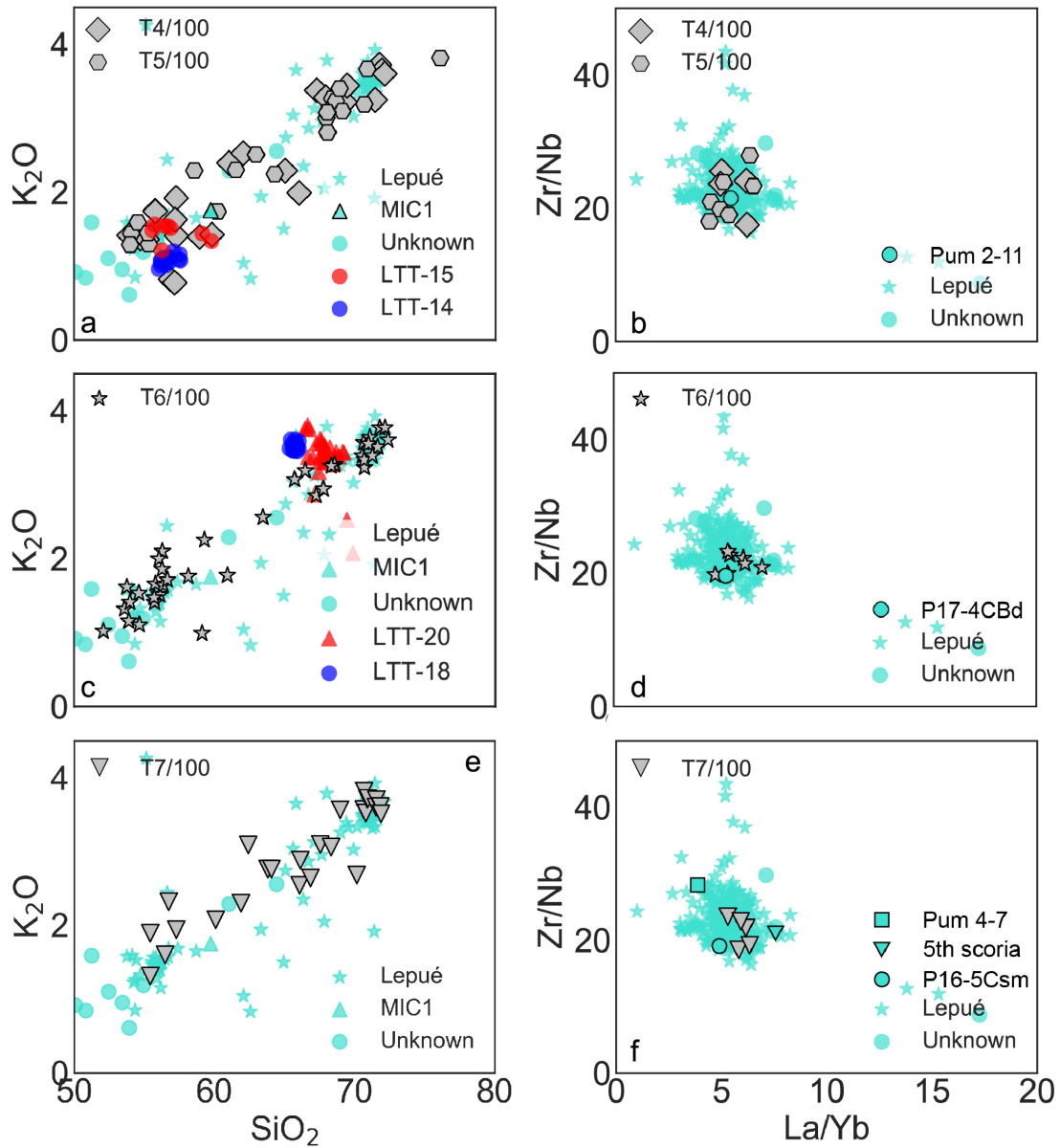
555 At Lago Teo, two tephtras occur in the Middle Holocene (LTT-14 and LTT-15), which have  
556 respectively estimated calendar ages of ~6.1 (1 $\sigma$ : 6.3–5.8) and ~6.2 (1 $\sigma$ : 6.4–6) cal ka BP  
557 (Moreno et al., 2014), and have been suggested to follow a geochemical trend similar to  
558 Michinmahuida. From these, LTT-14 displays values with lower K (Figure 7a) than

559 commonly reported for Michinmahuida, meanwhile the composition of LTT-15 is closer to  
560 Michinmahuida and to the less evolved part of the trend in both T4/100 and T5/100 (Figure  
561 7a). The estimated calendar ages of LTT-15 and LTT-14 overlap (Figure 6) and they are both  
562 within error synchronous to T4/100, hence it is not possible to further correlate these  
563 deposits.

564 Additionally, both T4/100 and T5/100 could potentially be correlated to the MIC1 tephra  
565 from Michinmahuida volcano (Amigo et al., 2013; Naranjo and Stern, 2004). The MIC1  
566 tephra was first identified as an andesitic pumice deposit ~90 km east of Michinmahuida  
567 volcano by Naranjo and Stern (2004). They obtained two <sup>14</sup>C ages from peat bracketing the  
568 MIC1 deposit, which indicate an age equal to or younger than ~7.2 and older than ~5.8 cal  
569 ka BP (Figure 6, Table S6). A correlative of this tephra was later identified by Amigo et al.  
570 (2013) in the Mallines section, who dated the underlying paleosol at ~7.5 cal ka BP.  
571 However, a younger age was obtained in the same paleosol a few centimeter below,  
572 indicating this age altered by Amigo et al. (2013), thus a robust correlation based on  
573 geochemistry cannot be done. However, from the sole available analysis, major elements  
574 display similar values to T4/100 and T5/100 (Figure 7a). Additional evidence for eruptive  
575 activity of Michinmahuida during this time is also provided by Alloway et al. (2017b), who  
576 describe the occurrence of a tephra layer in various sections (Pumalín-3, -4, -7, -2) in the  
577 area nearby Michinmahuida volcano occurring directly under Pumalín/Cha2 tephra,  
578 coincident with the position of MIC1 in the Mallines section. One trace element analysis in  
579 Pumalín-2 section (Pum 2-11) is provided by Alloway et al. (2017a) in bulk tephra, which  
580 displays a Michinmahuida-like geochemistry (Figure 7b).



581 From a chronological point of view, it is possible that T4/100 is correlated to LTT-15 (and/or



582

583 Figure 7. Comparison of specific cryptotephra belonging to Group 1 displaying a Michinmahuida signature with potential  
 584 correlatives on land. In all cases all the available information from Michinmahuida is plotted as cyan symbols in the  
 585 background (Alloway et al., 2017a; Amigo et al., 2013; López-Escobar et al., 1993; Naranjo and Stern, 2004) and specific  
 586 correlatives are indicated in the legend. LTT-14, LTT-15, LTT-18 and LTT-20 (Moreno et al., 2014), MIC1 ( Amigo et al.,  
 587 2013; Naranjo and Stern, 2004), Pum 2-11 and Pum 4-7 (Alloway et al., 2017b), P17-4CBd and P16-5Csm (Casati et al.,  
 588 2019), 5th scoria (Amigo et al., 2013), Lepué (Alloway et al., 2017a; Amigo et al., 2013), Unknown correspond to products  
 589 not assigned to any particular eruption (Amigo et al., 2013; López-Escobar et al., 1993). Major elements correspond to  
 590 normalized volatile-free compositions.

591 estimate might not be accurate. Unfortunately, only three geochemical data points  
592 identified as MIC1 are provided in the literature, all in bulk samples, two of which are  
593 thought to be  
594 LTT-14) and that T5/100 is correlated to MIC1. Additionally, because of the large  
595 uncertainty associated with the age of MIC1, it is also possible that LTT-15 and T4/100 are  
596 correlatives of MIC1 (Figure 6) and that no correlative of LTT-15 has been identified.

#### 597 5.1.4.3 T6/100

598 T6/100 occurs in the Early Holocene and has an estimated calendar age of  $\sim 8.5$  cal ka BP  
599 ( $1\sigma$ : 8.6–8.3). Evidence of Michinmahuida activity around this time is again found in Lago  
600 Teo and in one section nearby the village of Chaitén (Casati et al., 2019). Casati et al. (2019)  
601 identified a Michinmahuida–sourced deposit dated in  $\sim 8.4$  cal ka BP ( $1\sigma$ : 8.4–8.3) for which  
602 one geochemical analysis is available displaying La/Yb versus Zr/Nb ratios coincident with  
603 T6/100 (Sample P17-4CBd, Figure 7d), thus we propose this deposit is a likely correlative of  
604 T6/100. In Lago Teo, two tephra identified as Michinmahuida occur during this period:  
605 LTT-18 and LTT-20 with calendar age estimates for tephra LTT-18 of  $\sim 7.9$  cal ka BP ( $2\sigma$ : 8–  
606 7.8) and of  $\sim 8.5$  cal ka BP ( $2\sigma$ : 8.6–8.4) for LTT-20. Thus, LTT-20 could correspond to  
607 T6/100. When comparing the major element geochemistry, LTT-20 represents a cluster  
608 around 68 wt.% SiO<sub>2</sub>, whereas T6/100 is rather bimodal with a wider range of  
609 compositions, almost identical to Lepué tephra (Figure 7c). However, they might still  
610 correspond to the same tephra, and further information might reveal in the future a wider  
611 compositional range for the eruption that formed LTT-20.

#### 612 5.1.4.4 T7/100

613 T7/100 is the oldest cryptotephra that occurs in the Early Holocene, which has an  
614 estimated age of  $\sim 9.6$  cal ka BP ( $1\sigma$ : 9.7–9.5). Evidence on land for Michinmahuida activity  
615 around this time is presented by Casati et al. (2019), who dated a deposit attributed to  
616 Michinmahuida at  $\sim 9.2$  cal ka BP ( $1\sigma$ : 9.3–9.2), very similar to the estimated calendar age  
617 for T7/100 (Figure 6). Amigo et al. (2013) identify one scoria deposit (*5<sup>th</sup> scoria* in the  
618 Mallines section) from Michinmahuida and bracket this event between  $\sim 7.6$  and 10.3 cal ka  
619 BP. This deposit is positioned above Cha1 tephra, for which a calibrated age estimate  
620 ranging from  $\sim 9.5$  to 10.5 cal ka BP has been proposed (Alloway et al., 2017b; Amigo et al.,  
621 2013; Fontijn et al., 2016; Naranjo and Stern, 2004; Watt et al., 2011b). A potential  
622 correlative of this deposit is also identified by Alloway et al. (2017b) in section Pumalín-4,  
623 which occurs above Cha1 as the *5<sup>th</sup> scoria* from the Mallines section and is proposed to  
624 have a geochemical composition corresponding to Michinmahuida. A comparison of the  
625 element measurement for Pumalín-4 (Pum 4-7), the deposit (P16-5Csm) identified by  
626 Casati et al. (2019) and the *5<sup>th</sup> scoria*, display La/Yb versus Zr/Nb values close to those know  
627 for Michinmahuida and to T7/100 (Figure 7f).

#### 628 5.1.4.5 T6/98 and T8/100

629 T6/98 and T8/100 correspond to the only visible tephras in cores MD07-3098 and MD07-  
630 3100, respectively. Both tephras occur in the same stratigraphic position in each core at the  
631 Holocene-Deglaciation transition, and at synchronous calendar ages, within error, of  $\sim 10.6$   
632 ( $1\sigma$ : 10.7–10.4) for T8/100 and  $\sim 10.5$  ( $1\sigma$ : 10.7–10.3) cal ka BP for T6/98 (Figure 6).  
633 Additionally, they both have a thickness of  $\sim 25$ –30 cm and similar major element

634 geochemistry (Figure 3f), thus we propose that both tephras are correlatives. The major  
635 element compositions tend to be bimodal with two clusters at 54–57 and 68–73 wt.% SiO<sub>2</sub>  
636 and both major and trace elements are consistent with volcanic deposits attributed to  
637 Michinmahuida, in particular the Lepué tephra (Figure 3f).

638 So far, only one deposit from Michinmahuida volcano has been thoroughly described in the  
639 literature: the Amarillo Ignimbrite/Lepué Tephra. The Amarillo Ignimbrite was first  
640 described and dated by Amigo et al. (2013) in the vicinities of Michinmahuida volcano.  
641 However because of the lack of geochemical information they were not able to identify the  
642 fall deposit associated with the eruption. Later, Alloway et al. (2017a) identified and  
643 described the associated regional tephra and named it Lepué because of its occurrence in  
644 and around Lago Lepué on Chiloé island, northwest of the Michinmahuida volcano. This is a  
645 tephra with a bimodal composition with values ranging from ~54–72 wt.% SiO<sub>2</sub> (Figure 3f),  
646 with an estimated calendar age around 10.9 cal ka BP (Alloway et al., 2017a). The Lepué  
647 tephra has been identified in dozens of different sites, as far away as ~250 km northwest  
648 from the Michinmahuida volcano (Figure 6), with a predominantly northern and  
649 northwestern dispersion (Alloway et al., 2017a). In particular, it has been identified on the  
650 Chilean continental margin at ODP-202 site 1233, located ~20 km southeast of marine core  
651 MD07-3098 and 50 km northeast of core MD07-3100. At ODP-202 site 1233, the Lepué  
652 tephra is ~30 cm thick and its age is bracketed between ~10 and 12.3 cal ka BP (Alloway et  
653 al., 2017a). The latter is consistent with the visible tephra layers of ~25–30 cm thick found  
654 in cores MD07-3100 and MD07-3098. Furthermore, when comparing their trace and major  
655 element geochemistry, both tephras in the studied cores are consistent with the Lepué

656 Tephra products (Figure 3b, f). Thus we identify the only visible tephra in the cores as the  
657 regional marker Lepu  tephra.

#### 658 5.1.4.6 T12/100, T14/100, T15/100

659 Cryptotephras T12/100, T14/100 and T15/100 have estimated calendar ages of  $\sim 14.2$  ( $1\sigma$ :  
660 14.3–14),  $\sim 15.9$  ( $1\sigma$ : 16.1–15.8) and  $\sim 18.2$  ( $1\sigma$ : 18.4–18) cal ka BP. No tephra with a  
661 geochemistry similar to Michinmahuida has been thoroughly described during the  
662 Deglaciation and late glacial and more particularly, none older than Lepu  tephra. The lack  
663 of Michinmahuida tephras could be related to the presence of the PIS in the area proximal  
664 to the volcano in the last glacial period. However, a recent study in the La Zeta sequence  
665 located in the Argentina sector of the Andean complex has allowed to recover some  
666 tephras whose origin has been tentatively attributed to the Michinmahuida, however no  
667 geochemistry is available for further comparison (Alloway et al., 2017b). In particular, the  
668 JT1 tephra dated at  $\sim 18$  cal ka BP ( $1\sigma$ : 18.2–17.8) has an age similar to that of the T15/100  
669 event ( $\sim 18.2$  cal ka BP). Additionally, Alloway et al. (2017b) identify two other deposits, JT4  
670 and JT5 with age estimates of 17.8 ( $1\sigma$ : 17.9–17.6) and 17.5 ( $1\sigma$ : 17.6–17.3) cal ka BP.  
671 These ages are not coincident with any of the marine cryptotephras here identified,  
672 however they represent an evidence of volcanic activity of the Michinmahuida volcano  
673 during deglacial period.

674 As a summary, it was possible to robustly identify two eruptions from the Group 1 tephras  
675 and cryptotephras: the Lepu  tephra from Michinmahuida (T6/98 and T8/100), occurring at  
676  $\sim 10.9$  cal ka BP; and the HW7 tephra from the Hudson volcano (T1/100), occurring at  $\sim 1.5$   
677 cal ka BP. Both of which have been shown to record a northwestern dispersion (Figure 6;

678 Alloway et al., 2017a; Carel et al., 2011; Haberle and Lumley, 1998). Additionally, eight  
679 cryptotephra layers were identified as Michinmahuida-sourced (T2/100, T4/100, T5/100,  
680 T6/100, T7/100, T12/100, T14/100 and T15/100). These results are consistent with  
681 previous research around Michinmahuida volcano which had suggested recurrent volcanic  
682 activity during the Holocene (Amigo et al., 2013; Moreno et al., 2014). However, it is still  
683 hard to provide robust correlations among deposits, which in part is due to the paucity and  
684 accuracy of age/stratigraphic constraints, major and/or trace elements geochemistry.  
685 Additionally, the apparent high recurrence of Michinmahuida volcano can also be a source  
686 of uncertainty in the correlation since age estimates might not be precise enough to  
687 distinguish among eruptions, as in the case of LTT-14 and LTT-15 (Figure 6). Here we  
688 combine age/stratigraphic constraints, major and/or trace elements geochemistry to  
689 provide robust correlatives for Michinmahuida-sourced tephras, in the ocean. The  
690 occurrence of tephras older than T8/100 (T2/100, T14/100 and T15/100, Figure 6) is  
691 particularly interesting because it provides robust new evidence for recurrent activity of  
692 the Michinmahuida volcano since the late glacial period. This new information might be  
693 missing on land because of the highly vegetated area, less research around Michinmahuida  
694 because of its inactivity in historical times, or because of glacial erosion associated to the  
695 presence of the PIS. In many sites the Lepu  tephra was found lying above glacial deposits  
696 (Alloway et al., 2017a), however in some other sites additional tephra deposits have been  
697 found between Lepu  and glacial till and more so, organic matter directly above glacial  
698 deposits has been dated as old as ~17 cal ka BP. In fact, estimates of the PIS extent since  
699 the Last Glacial Maximum indicate that a rapid retreat would have begun in this area

700 around 18 cal ka BP and that by 15 cal ka BP the ice was confined to the vicinities of the  
701 volcanoes (Davies et al., 2020), allowing for the deposition of deglacial tephras on land.  
702 Thus, it might be possible to find on land correlatives for the Michinmahuida cryptotephras  
703 discussed here. In particular, deposits tentatively associated with Michinmahuida have  
704 already been mentioned in the literature (Alloway et al., 2017a), which might be correlated  
705 if geochemical information becomes available.

## 706 **5.2 Group 2**

707 Cryptotephra layers T1/98, T2/98, T3/98, T4/98, T5/98, T9/100, T10/100, T11/100, T7/98,  
708 T13/100, T16/100 display a major element trend similar to that of volcanic centers  
709 Antillanca-Casablanca (Fontijn et al., 2016; Naranjo et al., 2017; Geoffroy et al., 2018;  
710 Villarosa et al., 2006), Puyehue-Cordón Caulle (Figure 4, S5; Alloway et al., 2015; Bertrand  
711 et al., 2014; Fontijn et al., 2016; Gerlach et al., 1988; Lara et al., 2006; Naranjo et al., 2017;  
712 Singer et al., 2008; Villarosa et al., 2006) and Yate (López-Escobar et al., 1993; Mella, 2008;  
713 Watt et al., 2011b). T1/98 occurred in the Late Holocene; T2/98, T3/98, and T4/98 occurred  
714 in the Middle Holocene; T5/98 in the Early Holocene; T9/100, T10/100, T11/100, T7/98 and  
715 T13/100 during the Deglaciation; and T16/100 during the late glacial (Figure 2, 8).

716 Three eruptions sourced from Antillanca-Casablanca have been recognized so far in the  
717 literature: Nahuel Huapi (Fontijn et al., 2016; Naranjo et al., 2017), Rayhuen (Fontijn et al.,  
718 2016) and Playas Blanca-Negra (Fontijn et al., 2016; Geoffroy et al., 2018; Naranjo et al.,  
719 2017; Villarosa et al., 2006). All of them have been dated and occurred between ~1.3–3.5  
720 cal ka BP (Fontijn et al., 2016; Naranjo et al., 2017; Villarosa et al., 2006) and cannot be

721 correlated with any of the cryptotephtras studied here (Figure 8). In the case of Puyehue-  
722 Cordón Caulle, five Holocene tephtras deposits are described in the literature: Mil Hojas,  
723 North Cordón Caulle, Ranco, Puyehue2/PCC2 and Puyehue1/PCC1 (Figure 8; Fontijn et al.,  
724 2016; Lara et al., 2006; Naranjo et al., 2017; Singer et al., 2008), and volcanic activity during  
725 the Deglaciation has been identified by  $^{40}\text{Ar}/^{39}\text{Ar}$  dating of some effusive products between  
726  $12 \pm 2.8$  and  $18.7 \pm 1.1$  cal ka BP (Singer et al., 2008). Finally, two tephtras of Yate volcano  
727 have been recognized in the literature: Ya1 and Ya2. Ya2 has been dated bracketed  
728 between  $\sim 0.4$  and  $0.9$  cal ka BP, whereas Ya1 is thought to represent a Holocene eruption  
729 but its age is not constrained (Watt et al., 2011b). Additionally,  $^{40}\text{Ar}/^{38}\text{Ar}$  dating together  
730 with the observation of volcanic products associated or not with glacial erosion have been  
731 identified and ascribed to the Yate volcano, suggesting a continuous activity since  $122 \pm 19$   
732 cal ka BP (Mella, 2008). Geochemically, some differences can be observed between these  
733 volcanic centers. When comparing glass shard major elements (Figure 4, Figure S8),  
734 Puyehue-Cordón Caulle products are mainly dacitic or rhyolitic (66–74 wt.%  $\text{SiO}_2$ ) with  
735 relatively intermediate K (2.2–3.5 wt.%  $\text{K}_2\text{O}$ ; (Alloway et al., 2015; Fontijn et al., 2016;  
736 Naranjo et al., 2017)) compared to the other volcanoes with similar silica concentration  
737 (Yate: 2.6–3.5 wt.%  $\text{K}_2\text{O}$ ; Antillanca-Casablanca: 2.2–3.2 wt.%  $\text{K}_2\text{O}$ ). Antillanca-Casablanca  
738 glass shards display a trend with relatively less evolved products (56–75 wt.%  $\text{SiO}_2$ ) than  
739 Puyehue-Cordón Caulle, from andesitic to rhyolitic compositions at relatively lower K than  
740 Puyehue-Cordón Caulle (Fontijn et al., 2016; Geoffroy et al., 2018; Naranjo et al., 2017;  
741 Villarosa et al., 2006). In the case of Yate, individual glass shards ranging from andesites to  
742 rhyolites have been erupted since the last glaciation (Mella, 2008; Watt et al., 2011b). The

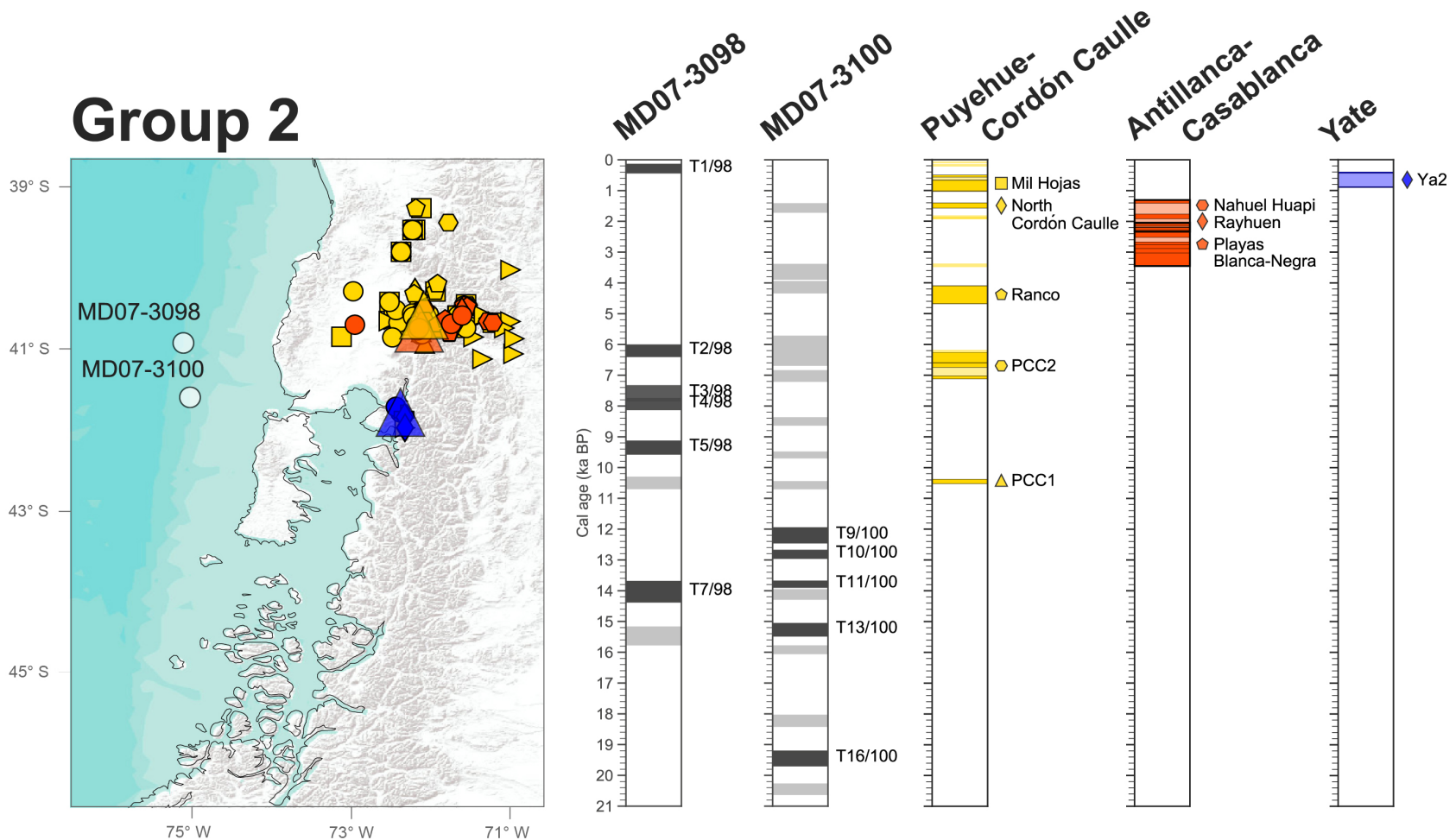


743 eruption with the most information corresponds to Ya2, which displays a continuous trend  
744 from basaltic andesite to rhyolite with relatively higher K than the two other volcanoes  
745 (Figure 4a, c, e), particularly at andesitic compositions (Figure 4c, e). For the three volcanic  
746 centers, whole-rock compositions (Gerlach et al., 1988; Jacques et al., 2014; Lara et al.,  
747 2006; López-Escobar et al., 1993; Mella, 2008; Naranjo et al., 2017; Rawson et al., 2016;  
748 Singer et al., 2008) record a more complete trend, from basalts and basaltic andesite to  
749 rhyolite, however andesites are rare in Puyehue-Cordón Caulle and Antillanca-Casablanca.

750 Additionally, these volcanic centers can be distinguished based on their trace elements.  
751 Notably, La/Yb values of Puyehue-Cordón Caulle are lower, plotting between 4–7 (Figure  
752 4b), with Zr/Nb values ranging between 22 to 50; meanwhile Antillanca-Casablanca displays  
753 very heterogeneous La/Yb values ranging between 4–21 and Zr/Nb values from 21 to 62;  
754 and Yate displays La/Yb values between 3–12 and Zr/Nb from 24 to 39. The latter helps  
755 discriminate between Puyehue-Cordón Caulle and the other two volcanic centers, however  
756 Antillanca-Casablanca and Yate cannot be distinguished based on these ratios. When  
757 looking at other trace elements, for example Th versus Zr, Th versus Nb or Ce versus Ta  
758 (Figure S10) a similar behavior continues to be observed between Antillanca-Casablanca  
759 and Yate, thus at present a robust geochemical discrimination between them is not  
760 possible. In the following section, we discuss the potential source and correlative of each  
761 cryptotephra based on the available information, which for the moment is mainly  
762 associated with eruptions from the Puyehue-Cordón Caulle, however always keeping in  
763 mind that future work might reveal new tephras associated with Antillanca-Casablanca  
764 and/or Yate.

765 5.2.1 T1/98

766 The most recent cryptotephra layer from Group 2, T1/98, occurs in the Late Holocene and  
767 has an estimated age of  $\sim 0.3$  cal ka BP ( $1\sigma$ : 0.4–0.1). T1/98 displays bimodal compositions  
768 with two clusters between 54–60 and 66–69 wt.% SiO<sub>2</sub>, and its major element  
769 geochemistry



770

771 Figure 8. Chronology for cores MD07-3098 and MD07-3100 highlighting the cryptotephra associated to Group 2, as described in the text, compared with the chronologies of  
 772 explosive products from each volcanic center displaying a similar geochemical behavior (Figure 4). Only events dated by  $^{14}\text{C}$  are plotted. To the left: position of the marine  
 773 sediment cores here studied together with the position of each volcanic center associated to Group 2 indicated as a triangle, from north to south: Puyehue-Cordón Caulle,  
 774 Antillanca-Casablanca and Yate. Also shown are the position of tephrochronological samples associated with each volcanic center which have been analyzed for geochemistry  
 775 and/or dated as smaller size symbols. Each symbol corresponds to a different eruptive event indicated to the right in the respective volcanic center chronology (to the right).  
 776 Pyroclastic deposits which have been dated but no correlation has been established are plotted in the chronology but unlabeled and the position is indicated as a dot of the  
 777 corresponding volcanic center color. Details on the chronologies of each volcanic center in Table S6.

778 is more consistent with the available information from Yate than that of Puyehue-Cordón  
779 Caulle or Antillanca-Casablanca (Figure 4c). Notably, andesitic compositions are frequent in  
780 the most primitive glass shards of T1/98, but are virtually absent or very rare in Puyehue-  
781 Cordón Caulle and Antillanca-Casablanca volcanic products. Therefore, tephra T1/98 could  
782 be a correlative of the most documented eruption from Yate volcano Ya2 (Figure 8),  
783 bracketed between  $\sim 0.4$  and  $\sim 0.9$  cal ka BP. Additionally, although less likely, it could  
784 correspond to a Puyehue-Cordón Caulle eruption, for which many Late Holocene tephras  
785 have been described. In particular, the most recent described tephra from Puyehue-Cordón  
786 Caulle corresponds to the Mil Hojas eruption, for which an age estimate has been obtained  
787 by several authors between  $\sim 0.5$  and  $\sim 1$  cal ka BP (Fontijn et al., 2016; Lara et al., 2006;  
788 Naranjo et al., 2017; Singer et al., 2008), slightly older than T1/98. Additionally, two  
789 pyroclastic density current deposits associated with the Puyehue-Cordón Caulle and  
790 younger than Mil Hojas have age estimates of  $\sim 0.1$  ( $1\sigma$  range: 0.1–0.06) and  $\sim 0.2$  ( $1\sigma$  range:  
791 0.2–0.1) cal ka BP (Table S6, Lara et al., 2006). At the same time, Naranjo et al. (2017) also  
792 recognize two tephra deposits associated to Puyehue-Cordón Caulle which are older than  
793 the historic eruption of AD 1921–1922 and younger than the Late Holocene Mil Hojas  
794 eruption. It is possible that these deposits are correlated with the pyroclastic density  
795 current dated by Lara et al. (2006) and T1/98. Unfortunately, there is no geochemistry  
796 available for either of these deposits and thus, a further correlation cannot be done. The  
797 most likely correlative of T1/98 would thus be the eruption Ya2 based on available  
798 information.

### 799 5.2.2 T2/98

800 T2/98 occurred during the Middle Holocene at an estimated calendar age of 6.2 cal ka BP  
801 ( $1\sigma$ : 6.4-6). Around this time, one eruption associated with Puyehue-Cordón Caulle has  
802 been described in the literature: Puyehue 2/PCC2, with estimated calendar ages ranging  
803 from 6.4 to 7.1 cal ka BP (Fontijn et al., 2016; Lara et al. 2006; Naranjo et al., 2017; Singer  
804 et al., 2008). The deposits corresponding to Puyehue 2/PCC2 were first  $^{14}\text{C}$  dated by Lara et  
805 al. (2006) at  $\sim 6.7$  ( $1\sigma$ : 6.7–6.6) cal ka BP (Figure 8, Table S6). The same deposit is also  
806 described by Singer et al. (2008) who obtained bulk tephra geochemistry for the pyroclastic  
807 deposit and whole-rock geochemistry for its effusive correlative, a rhyolite lava dome  
808 named Pr2 and dated by  $^{40}\text{Ar}/^{39}\text{Ar}$  in  $6.4 \pm 0.9$  and  $6.9 \pm 1.6$  cal ka BP. Additional  
809 information is provided by Fontijn et al. (2016), who obtained individual glass shard major  
810 element geochemistry for a likely correlative of this tephra (which they name PCC2), and  
811 provide a modelled age of  $\sim 6.4$  cal ka BP. Finally, Naranjo et al. (2017) date the underlying  
812 sediment of this tephra in 7.1 cal ka BP ( $1\sigma$ : 7.2–7). From a chronological point of view it is  
813 possible that cryptotephra T2/98 is correlated to Puyehue2/PCC2. From a geochemical  
814 point of view, the most evolved cluster from T2/98, with values between  $\sim 70$ –75 wt.%  $\text{SiO}_2$   
815 is consistent with analyses of PCC2 (Figure 4d) but the marine tephra follows a more  
816 complete trend with  $\text{SiO}_2$  values ranging from  $\sim 60$ –75 wt.%  $\text{SiO}_2$ , whereas on land volcanic  
817 products are restricted to 70–73 wt.%  $\text{SiO}_2$ . However, most of the available analyses were  
818 obtained by Fontijn et al. (2016), who mention that less evolved products were sampled  
819 but not analyzed because of the highly crystalline ground mass hampering major element

820 measurements. Thus, the lack of less evolved products from Puyehue 2/PCC2 might be an  
821 analytical bias rather than a geochemical fingerprint of this eruption.

### 822 5.2.3 T3/98 and T4/98

823 T3/98 occurred during the Middle-Early Holocene with calendar age estimates of  $\sim 7.6$  cal  
824 ka BP ( $1\sigma$ : 7.9–7.3). It displays bimodal compositions with two clusters between 56–59 and  
825 65–72 wt.% SiO<sub>2</sub> (Figure 4e). T4/98 occurred during the Middle-Early Holocene at  $\sim 7.9$  cal  
826 ka BP ( $1\sigma$ : 8.1–7.7). Glass shards in T4/98 display a continuous trend from 59–73 wt.% SiO<sub>2</sub>.  
827 During this time interval none of the volcanic centers in Group 2 has a recorded activity  
828 (Figure 8), except for effusive products of Puyehue-Cordón Caulle dated at  $6.9 \pm 1.6$  ka  
829 which have been correlated to the Puyehue2/PCC2 eruption (Singer et al., 2008). From the  
830 major elements alone it is not possible to further constrain the volcanic source of either of  
831 these two cryptotephra, thus future work might reveal potential correlatives for them.

### 832 5.2.4 T5/98

833 T5/98 occurs in the Early Holocene at  $\sim 9.4$  cal ka BP ( $1\sigma$ : 9.6–9.1). It presents a bimodal  
834 geochemistry with values between 54–61 and 67–72 wt.% SiO<sub>2</sub>. The only described  
835 eruption from volcanoes in Group 2 during the Early Holocene corresponds to Puyehue  
836 1/PCC1. Puyehue 1/PCC1 is a tephra found overlying moraine deposits or directly on rocks  
837 presenting glacial erosion (Naranjo et al., 2017), which has been dated by Singer et al.  
838 (2008), as part of unit Pr1, at  $\sim 10.5$  cal ka BP ( $1\sigma$ : 10.5–10.4) and its whole-rock analysis  
839 corresponds to a dacitic composition. The estimated calendar ages of T5/98 and PCC1 do  
840 not coincide, notably the estimated calendar age of Puyehue 1/PCC1 is  $\sim 1000$  years older  
841 than that of T5/98. When comparing the major element geochemistry of T5/98 and that of

842 PCC1/Puyehue 1 (Figure 4f), the most evolved part of the bimodal trend in the marine  
843 tephra is coincident with glass shard analyses by Fontijn et al. (2016) and whole-rock  
844 analyses by Singer et al. (2008). Additionally, one bulk tephra analysis (Rawson et al., 2016)  
845 has values similar to the less evolved part of the trend observed in T5/98, around 61 wt.%  
846 SiO<sub>2</sub> (Figure 4f).

847 With the current age assessment for Puyehue 1/PCC1, it seems unlikely that T5/98 is the  
848 correlative of this tephra. However because so far only one age constraint is available, the  
849 estimated age of PCC1 might greatly vary in the future. As mentioned, for example, in the  
850 case of MIC1 in Section 5.1.4.2, for which the estimated age ranges between ~5.8 and 7.2  
851 cal ka BP, age constraints for a tephra can greatly vary at different sites depending on the  
852 measured material and stratigraphic position of it, and thus the assessment of more dates  
853 could help further constraint the correlation between Puyehue 1/PCC1 and T5/98.

#### 854 5.2.5 T9/100

855 Cryptotephra T9/100, occurring at ~12.2 cal ka BP (1 $\sigma$ : 12.5–11.9), is the only cryptotephra  
856 from Group 2 for which trace elements have been analyzed. As mentioned before,  
857 volcanoes in Group 2 can be further distinguished based in their La/Yb ratios. In particular,  
858 Puyehue-Cordón Caulle has very uniform and relatively low La/Yb values, around 5. In  
859 contrast, Antillanca-Casablanca and Yate volcanic centers display more heterogeneous  
860 La/Yb values ranging between ~3–21, similar to what is observed in T9/100 which presents  
861 values between ~5–12 (Figure 4b). Surprisingly, when looking at other trace element  
862 concentrations, such as Th versus Zr or Nb, cryptotephra T9/100 seems to follow a

863 Puyehue-Cordón Caulle trend (Figure S14). From the latter, the geochemical signature from  
864 T9/100 is so far puzzling and a potential source is not yet identifiable.

#### 865 5.2.6 Deglacial and late glacial cryptotephra

866 According to our results, five cryptotephra from Group 2 occur during the Deglaciation  
867 and one during the late glacial period in addition to T9/100. Cryptotephra T10/100 occurs  
868 at ~12.8 cal ka BP ( $1\sigma$ : 13–12.7), T11/100 at ~13.8 cal ka BP ( $1\sigma$ : 13.9–13.7), T7/98 at ~14  
869 cal ka BP ( $1\sigma$ : 14.4–13.7), T13/100 at ~15.3 cal ka BP ( $1\sigma$ : 15.5–15) and finally T16/100 at  
870 ~19.4 cal ka BP ( $1\sigma$ : 19.7–19.2). From these, it is possible that T7/98 and T11/100 are  
871 correlated since they are within error synchronous and display similar major element  
872 trends (Figure 4h, S15).

873 No tephra deposits have been robustly identified on land for the volcanic centers in Group  
874 2 during the Deglaciation. As previously mentioned, Puyehue1/PCC1 (~10.5 cal ka BP ( $1\sigma$ :  
875 10.5–10.4) is found deposited over moraine deposits or glaciated rocks (Naranjo et al.,  
876 2017), hence no older tephra deposit is known. However, Singer et al. (2008) dated effusive  
877 products of the Puyehue-Cordón Caulle by  $^{40}\text{Ar}/^{39}\text{Ar}$  at  $12 \pm 2.8$ ,  $12.3 \pm 1.9$ ,  $14.9 \pm 2.9$ ,  $18.7$   
878  $\pm 1.1$  cal ka BP. Unfortunately, the higher error associated with the  $^{40}\text{Ar}/^{39}\text{Ar}$  method  
879 relative to  $^{14}\text{C}$  dating does not allow a direct correlation with these products. However, this  
880 information demonstrates that Puyehue-Cordón Caulle was active during the Deglaciation  
881 and late glacial and thus it is a plausible source for the marine cryptotephra. Additionally,  
882 the effusive products from Puyehue-Cordón Caulle follow a wide range from basalts to  
883 rhyolites, coherent with the compositional variations found in cryptotephra from Group 2  
884 (Figure 4g).



885 Further information on the volcanic activity of volcanic centers in Group 2 during the  
886 Deglaciation might be found in a lake core record from Lago Puyehue (Pu-II; Bertrand et al.,  
887 2008), where two tephra layers with similar geochemistry (PU-II 929–930 and 933–934 cm)  
888 have an estimated age slightly older than 13.8 cal ka BP. Bertrand et al. (2008) propose a  
889 Puyehue-Cordón Caulle source for PU-II 933–934 cm and suggest a more distant source for  
890 PU-II 929–930, however no exact volcanic center is indicated. When observing their major  
891 element geochemistry, they both present a bimodal trend with the more evolved part of  
892 the trend plotting very closely to the more evolved products from Antillanca-Casablanca  
893 and Puyehue-Cordón Caulle (Figure 4h). On the other hand, the less evolved part of the  
894 trend could correspond to either of the volcanic centers in Group 2. When comparing these  
895 tephtras with T11/100 and T7/98, dated at ~13.8 and ~14 cal ka BP respectively, both  
896 cryptotephtras and tephtras on land represent bimodal trends with geochemical similarities  
897 around the less evolved clusters at ~60 wt.% SiO<sub>2</sub>. However, at higher silica values the  
898 comparison between products becomes less consistent, the on land tephtras having higher  
899 SiO<sub>2</sub> contents for lower K<sub>2</sub>O. Thus it is unlikely that they correspond to correlatives.

900 As a summary for Group 2, eleven cryptotephtras were identified as either Puyehue-Cordón  
901 Caulle, Antillanca-Casablanca or Yate sourced. However, the paucity of literature  
902 information did not enable us to correlate with robustness any of these cryptotephtras. We  
903 suggest that T1/98 might be correlated to Ya2, and that more on land chronological  
904 constraints together with trace element analyses could help further constrain this  
905 correlation and better discriminate it from Puyehue-Cordón Caulle activity at this time.  
906 Additionally we suggest that T2/98 might be correlated to PCC2/Puyehue 2, if less evolved

907 products from the latter are analyzed; and T5/98 to PCC1/Puyehue 1, if more age  
908 constraints on land become available. Regarding the remaining cryptotephra in this group,  
909 no tentative correlations could be established. However the data presented here supports  
910 previous evidence for volcanic activity of Puyehue-Cordón Caulle and Yate during Holocene,  
911 deglacial and late glacial periods for which the marine tephra might represent the distant  
912 evidence of this explosive volcanic activity. A better correlation with the on land  
913 information however requires more geochronological and geochemical data, for example  
914 Ya1 eruption so far no age constraint exists and only two geochemical bulk analyses of the  
915 tephra are available (Watt et al., 2011b). Additionally, most deglacial evidence potentially  
916 correlated to the cryptotephra here analyzed corresponds to  $^{40}\text{Ar}/^{39}\text{Ar}$  dating in effusive  
917 products which does not allow for a correlation because of the high errors associated with  
918  $^{40}\text{Ar}/^{39}\text{Ar}$  dating relative to  $^{14}\text{C}$  dating.

### 919 **5.3 Group 3**

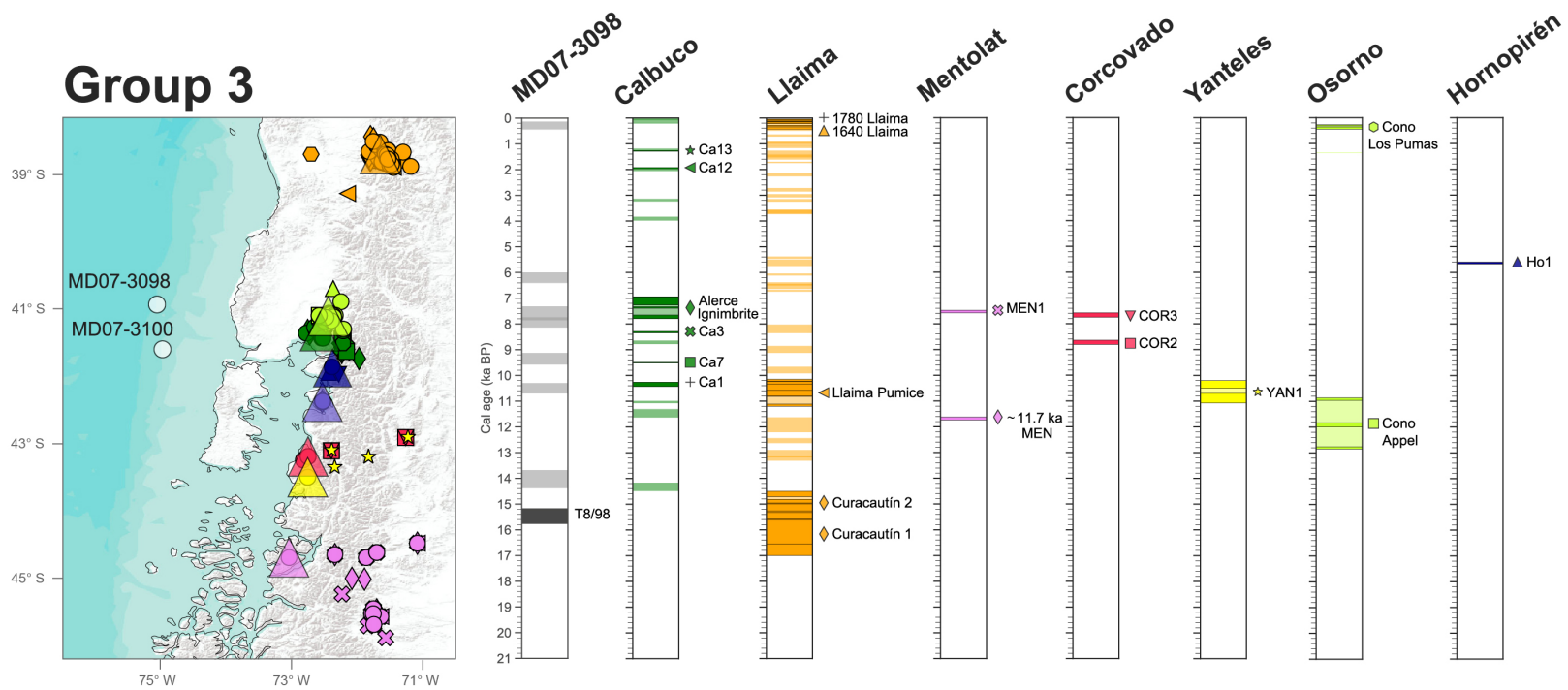
920 Cryptotephra layer T8/98 occurs at the beginning of the Deglaciation and has an estimated  
921 calibrated age of  $\sim 15.5$  cal ka BP ( $1\sigma$ : 15.8–15.2). The major element geochemistry of this  
922 cryptotephra is unique in both cores analyzed here, with much lower  $\text{K}_2\text{O}$  values ranging  
923 between  $\sim 0.5$  and 1.3 wt.% at variable  $\text{SiO}_2$  contents (Figure 5a and 3a, 4a for comparison).  
924 This particular geochemical fingerprint is shared by eight volcanic centers dispersed  
925 between 39 and 45° S along the SVZ (Figure 1): Llaima (Bouvet de Maisonneuve et al.,  
926 2012; Naranjo and Moreno, 1991, 2005; Rawson et al., 2016; Reubi et al., 2011;  
927 Schindlbeck et al., 2014; Lohmar, 2008), Osorno (Bertrand et al., 2008; Jacques et al., 2014;  
928 López-Escobar et al., 1993; López-Escobar et al., 1992; Moreno et al., 2010; Tagiri et al.,

929 1993), Calbuco (López-Escobar et al., 1995, 1992; Morgado et al., 2019; Watt et al., 2011b;  
930 Sellés and Moreno, 2011), Hornopirén (Watt et al., 2011b), Huequi (Watt et al., 2011a),  
931 Corcovado, Yanteles (López-Escobar et al., 1993; Naranjo and Stern, 2004) and Mentolat  
932 (López-Escobar et al., 1993; Naranjo and Stern, 2004; Stern et al., 2015; Weller et al., 2019,  
933 2017, 2015).

934 Among these volcanoes, one eruption from the Llama volcano has been well described and  
935 dated with a similar age to T8/98 (~15.5 cal ka BP). The Curacautín Ignimbrite, resulting  
936 from this eruption, is proposed to have occurred in two phases based on <sup>14</sup>C ages  
937 populations and field observations (Lohmar, 2008): Curacautín 1 (Table S6) with a mean  
938 age of ~16.1 cal ka BP and Curacautín 2 with a mean age ~15 cal ka BP (Figure 9). A direct  
939 geochemical comparison of individual glass shards major element geochemistry between  
940 T8/98 and Curacautín products indicates similar values below 60 wt.% SiO<sub>2</sub> (Figure 5b),  
941 however the more evolved values observed in T8/98 with low K<sub>2</sub>O are not present in  
942 Curacautín, for which glass shards with higher silica values display much higher K than  
943 tephra T8/98. The geochemical trend displayed by T8/98 seems more consistent with a  
944 poorly described pyroclastic deposit from the Calbuco volcano, dated at ~14.4 cal ka BP  
945 (Sellés and Moreno, 2011). For this deposit no geochemical data have been reported and  
946 the stratigraphic position of the charcoal used for <sup>14</sup>C dating is unknown, hampering further  
947 robust correlation. Regarding the remaining volcanoes, no deposits have been identified  
948 during this time interval (Figure 9) and the Osorno volcano is thought to be a volcano with a  
949 dominant effusive activity (Moreno et al., 2010), it might thus be a less likely possible  
950 source for T8/98.

951 From the latter, it is plausible that the cryptotephra T8/98 is correlated to either  
952 Curacautín<sup>1, 2</sup> or the unknown event from Calbuco volcano. Given the wind configuration  
953 in the area and the position of the volcanic centers relative to core MD07-3098, Calbuco  
954 volcano would be the more likely source of the cryptotephra (Figure 1). However, given the  
955 chaotic nature of atmospheric circulation as described in section 2, a Llaima source is not  
956 impossible.

957 Further information, especially regarding the volcanic activity of Calbuco could help clarify



958

959 Figure 9. Chronology for cores MD07-3098 and MD07-3100 highlighting the one cryptotephra associated to Group 3 (T8/98), as described in the text, compared with the  
 960 chronologies of explosive products from each volcanic center displaying a similar geochemical behavior (Figure 6). Only events dated by <sup>14</sup>C are plotted. To the left: position  
 961 of the marine sediment cores here studied together with the position of each volcanic center associated to Group 3 indicated as a triangle, from north to south: Llaima,  
 962 Osorno, Calbuco, Hornopirén, Huequi (no eruptive events have been dated for Huequi so far), Corcovado, Yanteles and Mentolat. Also shown are the position of  
 963 tephrochronological samples associated with each volcanic center which have been analyzed for geochemistry and/or dated, as smaller size symbols. Each symbol  
 964 corresponds to a different eruptive event indicated in the respective volcanic center chronology (to the right). Pyroclastic deposits which have been dated but no  
 965 correlation has been established are plotted but unlabeled in the chronology and the position is indicated as a dot of the corresponding volcanic center color. Details on the  
 966 chronologies of each volcanic center in Table S6.

967 this possible correlation in the future.

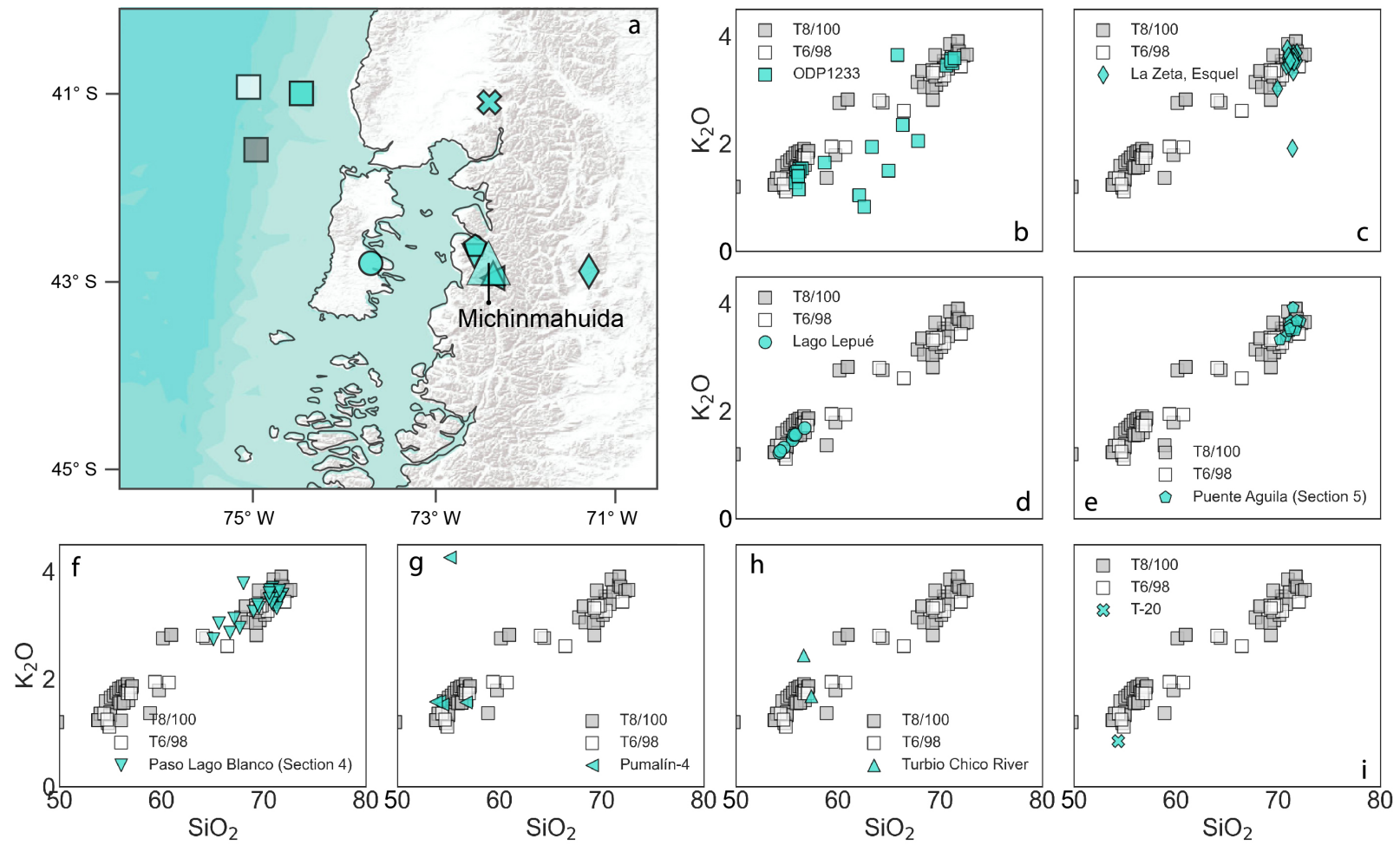
#### 968 **5.4 General comment**

969 As mentioned in the previous sections, all the cryptotephra and tephra layers analyzed here  
970 represent a trend, either bimodal or continuous, from basaltic andesite to dacite or  
971 rhyolite, which is in contrast with many records on land which represent clusters. This was  
972 also observed by Carel et al. (2011) when comparing marine tephra layers and on land  
973 tephtras from the Hudson volcano. This difference, which is observed in the case of T2/100  
974 versus LTT-9 and LTT-10, T3/100 versus H2 (Figure 3d, e), T4/100 or T5/100 versus LTT-14  
975 and LTT-15, LTT-20 compared to T6/100 (Figure 7a, c), T2/98 versus Puyehue 2/PCC2  
976 (Figure 4d) might indicate either that the deposits on land and the ocean arise from  
977 different eruptive events or it could also arise from biases associated to the dispersion of  
978 the ashes and/or measuring techniques. For example, it has been mentioned in the  
979 literature that analytical biases arise by the increased difficulty in analyzing glass shards  
980 with microlites, which are more abundant in less evolved melts (e.g. Alloway et al., 2017a;  
981 Fontijn et al., 2016). The latter could explain the lack of less evolved products in LTT-20  
982 compared to T6/100 (Figure 7c), however it would not explain the differences observed in  
983 T2/100 versus LTT-9 and LTT/10 (Figure 3d) and T4/100 or T5/100 versus LTT-14 and LTT-  
984 15 (Figure 7a).

985 An interesting situation regarding this point is observed in both identified eruptions in this  
986 study: HW7 and Lepu  tephra. For both these tephtras a number of different sections have  
987 been analyzed for geochemistry on land for which different sections display different  
988 geochemistry. In particular, in the case of Lepu  tephra, none of the analyzed sections on

989 land represent the whole trend (Alloway et al., 2017a; Amigo et al., 2013), however all

990 marine



991

992 Figure 10.  $SiO_2$  vs  $K_2O$  and position (a) of deposits correlated with Lepué tephra), indicated in the legend (b–i) are the given name for each section in which the Lepué tephra was  
 993 identified and analyzed. b–g deposits analyzed by Alloway et al. (2017a); i: Analyses of Amarillo ignimbrite, correlative of Lepué tephra (Amigo et al., 2013); h: Analysis of Lepué  
 994 correlative analyzed by Naranjo & Stern (2004), who initially identified this deposit as COR1.



995 tephtras do, including ODP-202 site 1233 and both tephtras here analyzed (Figure 10). In  
996 addition, a similar situation is observed for tephtra HW7, however in this case two of the  
997 studied sections on land: “Laguna Marcelo” and “Laguna Stibnite” also display the full trend  
998 (Haberle & Lumley, 1998), as do cryptotephtra T1/100 and cryptotephtra at 160 cm in core  
999 MD07-3088 (Carel et al., 2011), all of them located to the west of Hudson volcano. It is  
1000 difficult to pinpoint what give rise to the different geochemical compositions in the  
1001 different sections as this probably arises from a sum of different biases. For example: bias  
1002 associated with the depositional environment; bias associated with the analytical  
1003 techniques; biases associated with the very nature of the eruption i.e. its explosiveness  
1004 and/or duration; and finally, biases associated with the dispersal conditions of their  
1005 products, such as wind direction at the moment of the eruption. For these various reasons,  
1006 marine cryptotephtra might have the capacity of recording a more complete geochemical  
1007 trend from a single volcano than on land tephtras. Although the reasons are still unclear,  
1008 this would mean that marine tephtras/cryptotephtras can thus serve as robust compositional  
1009 reference for further correlations because they would be less geochemically biased. In  
1010 addition to the robust stratigraphic constraints here provided for each tephtra and  
1011 cryptotephtra, the geochemistry of the marine tephtras/cryptotephtras constitute robust  
1012 tephrochronological constraints to disentangle the complex tephrochronological record in  
1013 the SVZ.

## 1014 **6. Conclusion**

1015 Here we present a continuous marine (crypto) tephrochronological record associated with  
1016 the SVZ, derived from two marine sediment cores retrieved in the Chilean continental  
1017 margin. From the twenty five tephras and cryptotephras analyzed: one cryptotephra was  
1018 correlated with tephra HW7 from Hudson volcano previously identified by Carel et al.  
1019 (2011) and Haberle and Lumley (1998); and the Lepu  tephra (Alloway et al., 2017a; Amigo  
1020 et al., 2013), associated with Michinmahuida volcano, was identified in both studied cores  
1021 as a ~25–30 cm–thick visible tephra layer. The latter constitutes an important tie point to  
1022 align not only the chronologies of cores MD07-3098 and MD07-3100, but also the  
1023 chronologies of other available paleoceanographic record in the area, such as ODP-202 site  
1024 1233 (Lamy et al., 2004, 2015) and MD07-3088 (Carel et al., 2011; Siani et al.2013). Thus  
1025 representing a step forward towards producing a chronologically robust paleoceanographic  
1026 record in the area.

1027 Additionally, eight cryptotephra were identified as Michinmahuida sourced based on major  
1028 and trace elements for which robust stratigraphic constraints are provided. The latter is in  
1029 agreement with previous records suggesting continuous Holocene activity of  
1030 Michinmahuida (Amigo et al., 2013; Moreno et al., 2014), and provides new and robust  
1031 evidence for its deglacial and late glacial activity. Eleven cryptotephra were identified as  
1032 sourced either from Puyehue-Cord n Caulle, Antillanca-Casablanca or Yate, even though no  
1033 robust correlation could be established at this stage. Nevertheless, they represent evidence  
1034 for late glacial and deglacial explosive activity of either of these volcanic centers, for which  
1035 evidence on land has only been previously identified by Bertrand et al. (2008). The latter

1036 complements previous information from Puyehue-Cordón Caulle (Singer et al., 2008) and  
1037 Yate (Mella, 2008) for which effusive products have been dated indicating late glacial and  
1038 deglacial ages and provides further chronological constraints for the Puyehue-Cordón  
1039 Caulle activity.

1040 Overall, the marine sediment cores here studied provide robust geochemical and  
1041 stratigraphic evidence for a continuous volcanic activity in the SVZ since the late glacial (~20  
1042 cal ka BP). In the future, this information, together with an increased geochemical and  
1043 chronological on land database, might help further unveil the volcanic activity of the  
1044 different volcanic centers and provide additional tie points to synchronize different paleo  
1045 environmental records in the area.

## 1046 **7. Acknowledgements**

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1056 published data and whose comments helped improve the manuscript.

## 1057 8. Data Availability

1058 Datasets related to this article (Table S1-S5) can be found at \_\_\_\_.

## 1059 9. References

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