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1 Unusual geologic evidence of coeval seismic shaking and tsunamis shows
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19 **Abstract**

20 An uncommon coastal sedimentary record combines evidence for seismic shaking and
21 coincident tsunami inundation since AD 1000 in the region of the largest earthquake
22 recorded instrumentally: the giant 1960 southern Chile earthquake (Mw 9.5). The record
23 reveals significant variability in the size and recurrence of megathrust earthquakes and
24 ensuing tsunamis along this part of the Nazca-South American plate boundary. A 500-m
25 long coastal outcrop on Isla Chiloé, midway along the 1960 rupture, provides continuous
26 exposure of soil horizons buried locally by debris-flow diamicts and extensively by tsunami
27 sand sheets. The diamicts flattened plants that yield geologically precise ages to correlate
28 with well-dated evidence elsewhere. The 1960 event was preceded by three earthquakes
29 that probably resembled it in their effects, in AD 898-1128, 1300-1398 and 1575, and by
30 five relatively smaller intervening earthquakes. Earthquakes and tsunamis recurred
31 exceptionally often between AD 1300 and 1575. Their average recurrence interval of 85
32 years only slightly exceeds the time already elapsed since 1960. This inference is of serious
33 concern because no earthquake has been anticipated in the region so soon after the 1960
34 event, and current plate locking suggests that some segments of the boundary are already
35 capable of producing large earthquakes. This long-term earthquake and tsunami history of
36 one of the world's most seismically active subduction zones provides an example of
37 variable rupture mode, in which earthquake size and recurrence interval vary from one
38 earthquake to the next.

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40

41 **Highlights:**

42 Unusual geologic evidence of shaking and coincident tsunami.

43 Evidence clarifies seismic and tsunami history in the area of the giant 1960 Chile
44 earthquake.

45 Three giant earthquakes similar to 1960 and five intervening earthquakes of lesser
46 effects in the last millennium.

47 Megathrust ruptures varied in length and recurrence interval.

48 A prehistoric recurrence interval only slightly exceeds the time already elapsed since
49 1960.

50 **Keywords:**

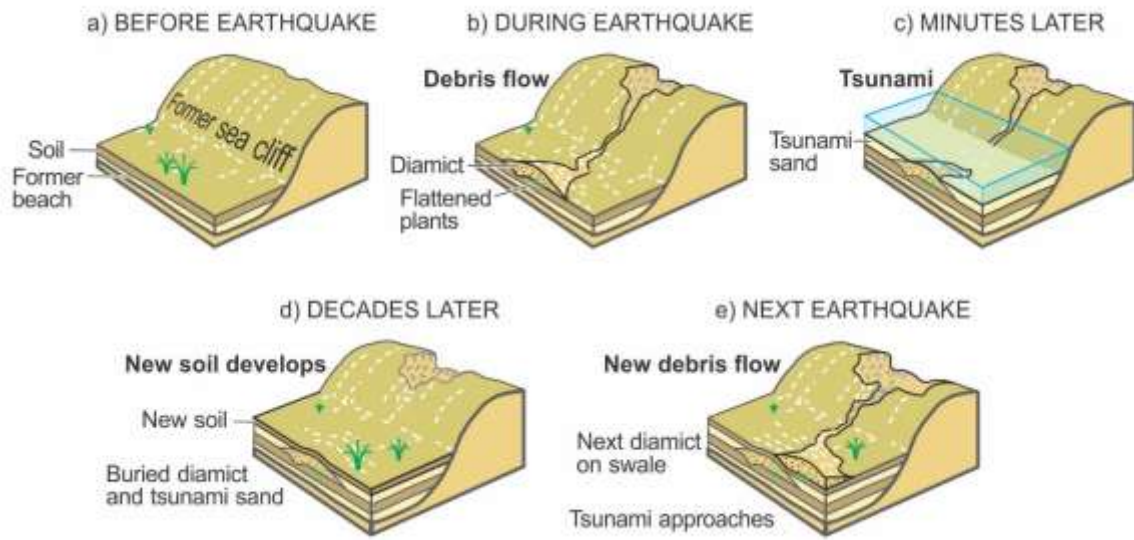
51 Paleoseismology

52 Tsunami geology

53 Earthquake recurrence

54 Variable rupture mode

55 1960 Chile earthquake



57 **1. Introduction**

58 As departures from recent historical experience, the 2004 Indian Ocean and 2011
59 Tohoku earthquakes and tsunamis underscore the importance of using geological evidence
60 to help estimate the variability among earthquakes that a given subduction zone can
61 produce (Satake, 2014). Variable rupture mode, in which a long segment of a subduction
62 zone sometimes ruptures in a single great earthquake, but in other times ruptures in a series
63 of relatively smaller earthquakes (Kanamori and McNally, 1982), was inferred decades ago
64 from written records of earthquakes between AD 684 and 1946 in southwest Japan (Ando,
65 1975), from instrumental records of earthquakes between 1906 and 1979 in Colombia and
66 Ecuador (Kanamori and McNally, 1982), and later from geological evidence for
67 earthquakes at several subduction zones (see Satake and Atwater, 2007).

68 This paper explores variation in rupture mode in the region of the largest earthquake
69 ever recorded instrumentally—the giant 1960 mainshock, Mw 9.5, in south-central Chile
70 (Fig. 1). A variable behavior was previously inferred for this region by comparing historical
71 earthquakes of 1575, 1737, 1837, and 1960 with stratigraphic evidence for land-level
72 change and tsunamis from estuaries (Cisternas et al., 2005; Ely et al., 2014; Garrett et al.,
73 2015; Hong et al., 2016), and with evidence for shaking in lakes of the Andean foothills
74 (Moernaut et al., 2007; Moernaut et al., 2014). These previous findings reinforce written
75 evidence indicating that the 1575 and 1960 earthquakes resembled one another, and that
76 both exceeded the relatively smaller 1737 and 1837 earthquakes in fault-rupture area.

77 Our work focuses on an uncommon combination of stratigraphic evidence for
78 seismic shaking and coincident tsunami inundation since AD 1000 at Cocotué on Isla
79 Chiloé (Figs. 2–5). Geological traces of shaking and tsunamis are rarely found together; in

80 a notable exception, sand blows from sediment liquefaction contributed to the 2011 tsunami
81 deposit on the Sendai Plain (Goto et al., 2012). The Cocotué stratigraphy suggests the
82 sequence of events cartooned in Fig. 2: (a) An earthquake triggers a debris flow on the
83 former sea cliff, either for the first time or by destabilizing a scarp made by a previous
84 earthquake. (b) The resulting debris-flow diamict covers soil at the foot of the slope, where
85 it flattens and buries plants that had been growing there. (c) An ensuing tsunami entrains
86 sand as it comes ashore and continues onto the toe of the debris-flow fan. The resulting
87 sand sheet tapers across the diamict and locally entrains some of its clasts. (d) A new soil
88 develops above both the tsunami sand and the debris-flow deposits. (e) This sequence
89 repeats each time an earthquake sets off another debris flow followed by a tsunami.

90 We combine the Cocotué stratigraphic record with previously reported geological
91 and historical evidence elsewhere in the 1960 region to reconstruct an earthquake history
92 spanning the last 1000 years (Fig. 6). We propose that three earthquakes since AD 1000
93 resembled the 1960 mainshock in fault-rupture area, and that five other earthquakes likely
94 radiated from smaller ruptures.

95 **2. Setting**

96 2.1. Cocotué outcrop

97 Our new geologic findings come from a previously undescribed outcrop along the
98 beach on the Pacific coast at Cocotué, on Isla Chiloé (Figs. 1, 3, S1). The site is midway
99 along the length of the 1960 mainshock rupture. The outcrop was incised by a migrating
100 creek and ocean waves that eroded into a low terrace at the foot of a former sea cliff. The
101 cliff, about 40 m high and composed of Pleistocene glacial outwash, was cut in the late

102 Holocene (Heusser and Foster, 1977; Porter, 1981; Heusser, 1990). The low terrace stands
103 about 3 m above present sea level (Fig. 4) despite about 1 m of subsidence in 1960 (Plafker
104 and Savage, 1970). Campaign GPS measurements 13 km to the northwest show that the
105 coast subsided 6 mm per year between 2005 and 2009 (Moreno et al., 2011).

106 Although the terrace is formed on beach deposits, its rear portion is covered by a
107 seasonal freshwater pond, and the rest by land plants, including shrubs and herbs (Figs. 3,
108 S1). Below we show that freshwater conditions have probably prevailed for most or all of
109 the roughly 1100 years since a soil began to form on the beach deposits.

110 The 22 May 1960 earthquake triggered debris flows on the face of the former sea
111 cliff and elsewhere on Chiloé as shown by airphotos taken in January 1961 (Fig. 3b). The
112 Cocotué outcrop cuts across the toes of the 1960 and earlier debris-flow fans, and also
113 exposes intercalated sand sheets (Fig. 3c).

114 2.2. Earthquake and tsunami written history in the 1960 area

115 Written records report at least three predecessors to the 1960 earthquake (Lomnitz,
116 1970; Cisternas et al., 2005). The earliest of these, in 1575, most closely resembles the
117 1960 mainshock based on the spatial extent and severity of its effects. The shaking in 1575
118 damaged Spanish settlements from Concepción to Castro (Fig. 1), and it triggered
119 landslides that blocked the outlet of Riñihue Lake in the Andean foothills. Five months
120 later an outburst flood killed thousands of natives and inundated Valdivia, 80 km
121 downstream (Lomnitz, 1970). The 1575 earthquake was also associated with a tsunami that
122 took more than 1000 lives north of Valdivia and wrecked galleons in that city's port.

123 The poorly known next predecessor, in 1737, reportedly damaged Valdivia and
124 towns of Isla Chiloé but no associated tsunami has been reported (Cisternas et al., 2005).
125 By contrast, the central Chile earthquakes of 1730 and 1751 farther to the north were both
126 associated with tsunamis that were noted not just in Chile (Lomnitz, 2004; Udías et al.,
127 2012) but also in Japan (Watanabe, 1998).

128 The most recent predecessor to the 1960 earthquake, in 1837, destroyed buildings
129 midway along the 1960 rupture area. It was accompanied by changes in coastal land level
130 farther south that lowered some islands (Vidal Gormaz, 1877) while raising others (Darwin,
131 1851). The ensuing tsunami appears to have been relatively small where documented at
132 Chiloé but crested 6 m high in Hilo, Hawaii (Coan, 1882) and caused flooding and damage
133 near Sendai, Japan (Watanabe, 1998). On the basis of the tsunami's height in Hilo, the
134 1837 earthquake was assigned a tsunami magnitude of 9¼ (Abe, 1979).

135 2.3. The 1960 earthquake

136 The mainshock of 22 May 1960 resulted from a rupture at the boundary between the
137 subducting Nazca plate and the overriding South America plate following an extraordinary
138 series of foreshocks (Cifuentes, 1989; Fig. 1a). The mainshock produced extensive damage
139 and abundant slope failures in most of south-central Chile (Weischet, 1963). Warping of
140 the overriding plate lowered one-quarter of Chile's outer coast by a meter or two (Plafker
141 and Savage, 1970; Fig. 1b). Concurrent displacement of the seafloor generated a tsunami
142 with peak heights of 15 m in Chile (Sievers, 1963), 10 m Hawaii (Eaton et al., 1961) and 6
143 m in Japan (Watanabe, 1998).

144 **3. Previous work**

145 Paleoseismological studies have extended the history of earthquakes and tsunamis
146 in south-central Chile thousands of years into the past. Some of these studies have relied on
147 evidence for coastal land-level changes and tsunamis, while others have used evidence for
148 shaking. So far none has combined both types of evidence for the same event.

149 3.1. Land-level changes and tsunamis

150 Midway along the 1960 earthquake rupture area, reconnaissance studies in the late
151 1980s uncovered coastal evidence for its predecessors in the Maullín estuary, located on the
152 mainland 50 km north of Cocotué (Fig. 1). Buried soils provided evidence for recurrent
153 subsidence despite net late Holocene emergence of tidal-marshes (Atwater et al., 1992).
154 Further study yielded a 2000-yr history of repeated coseismic subsidence and tsunamis
155 (Cisternas et al., 2005). The average interval between 1960-type events was two to three
156 times longer than the 128-year average interval separating the 1575, 1737, 1837, and 1960
157 events. The youngest four events in the Maullín sequence (D, C, B, and A), each marked by
158 subsidence and tsunami, occurred within the past 1000 years. Radiocarbon analyses dated
159 event D to AD 1020-1180 and event C to AD 1280-1390. Maullín event B likely
160 corresponds to 1575 and event A is unquestionably 1960. Traces of subsidence and
161 tsunamis in 1737 and 1837 were sought but not found at the estuary. Indeed, there is a
162 widespread absence of buried soils intermediate between those conspicuously preserved
163 from 1575 and 1960. In addition, tree-ring (see below) and written evidence argued against
164 19th-century subsidence.

165 Landforms and pits on a beach-ridge plain at Caille, 7 km southwest of Maullín,
166 provided stratigraphic evidence for three tsunamis, including the one in 1960 (Atwater et
167 al., 2013). The site, two kilometers inland, includes a breach in a Holocene beach ridge, a

168 pond that extends landward from the breach, and a fan that partly rims the pond. At least
169 two tsunamis prior to 1960 cut the breach. The 1960 tsunami reamed out the breach and
170 aggraded the fan with a sedimentary breccia of sand and soil clasts. Two older sand
171 deposits were found below the 1960 soil, each resting on buried soils. Dating gave an age
172 of AD 1270-1400 for the lower sand bed, similar to the age range of Maullín event C (AD
173 1280-1390). The intermediate sand sheet was stratigraphically ascribed to the 1575
174 tsunami.

175 Coastal evidence for recurring earthquakes was also identified at Chucalén, an
176 estuarine site 10 km north of Cocotué on the sheltered coast of northern Chiloé (Garrett et
177 al., 2015; Fig. 1c). Four buried soils, at least three of them capped with a sand sheet,
178 provide evidence of earthquakes and tsunamis in 1960 and 1575 and at two earlier times
179 overlapping the ages of Maullín's events C and D. Changes in diatom assemblages show
180 that subsidence coincided with at least three of the earthquakes. As in Maullín, no geologic
181 evidence of the historical earthquakes in 1737 and 1837 was found in Chucalén.

182 Evidence of land-level change and tsunamis also has been found in the northern half
183 of the 1960 rupture area. Cores in marshes of the Valdivia estuary (Fig. 1b), 230 km north
184 of Cocotué, encountered fragmentary evidence of subsidence and tsunamis before 1960
185 (Nelson et al., 2009). This evidence includes two successive buried soils that are each
186 overlain by sand and were dated to 2700-1700 and 1700-1300 cal yr BP.

187 Estuarine stratigraphy at Tirúa (Fig. 1b), 400 km north of Cocotué, shows four
188 instances of land-level change and accompanying tsunamis over the past 450 years within
189 the overlapping ruptures areas of the 1960 and 2010 earthquakes (Ely et al., 2014).
190 Negligible subsidence reported in 1960 at Tirúa (Plafker and Savage, 1970) contrasts with

191 the uplift of 0.5-1 m of this area during the 2010 earthquake. Likewise Tirúa appears to
192 have subsided during the southern Chile earthquake of 1575 and to have been uplifted
193 during the 1751 earthquake to the north. Based on multiple radiocarbon ages and
194 stratigraphic position, Ely et al. (2014) ascribe one of the sand sheets at Tirúa to the 1575
195 tsunami. Yet, OSL dating suggests older ages (AD 1390±80) for this and other sand sheets
196 in the sequence (Nentwig et al., 2015).

197 Evidence of five tsunamis was found in an abandoned meander and banks of
198 Quidico River, 10 km north of Tirúa (Hong et al., 2016). All events but the oldest were
199 correlated with historical tsunamis in 2010, 1960, 1835 and 1751. Although Quidico lies
200 near the northern boundary of the 1960 region, its sequence appears to record tsunamis
201 sourced in the 2010 earthquake region. A prehistoric tsunami dated to AD 1445-1490 is
202 marked by a thick and widespread sand sheet. Its source remains unknown.

203 Further clues about land-level changes come from arboreal evidence. Drowned
204 forests of dead or dying trees at San Rafael fiord (Fig. 1b), at the southern end of the 1960
205 rupture, provided evidence for subsidence in 1837. Counting the annual growth rings of one
206 drowned tree, still rooted to the floor of the fiord, Reed et al. (1988) noted it became
207 stunted and tilted in the 1830s. Comparing the level of the lowest healthy living trees with
208 the level of the lowest well preserved dead stumps, they inferred a subsidence between 2.0
209 and 2.5 m.

210 In contrast, old-growth trees in northern Chiloé and near Maullín, killed or damaged
211 by tidal submergence after subsiding in 1960, show no evidence of subsidence in 1837.
212 Measuring tree diameters and annual rings, Bartsch-Winkler and Schmoll (1993) suggested
213 that a forest drowned in 1960 at the Pudeto estuary of Chiloé (Fig. 1d) had failed to subside

214 in 1837. On the mainland at Misquihué (Fig. 1c) a similar story was inferred counting rings
215 of 15 trees killed in 1960. Ten of them were found to have lived through 1837 and two
216 through 1737 as well (Cisternas et al., 2005). Furthermore, on an 1874 nautical chart,
217 Chilean Navy surveyors schematically depicted the Misquihué trees as large and healthy.

218 3.2. Geologic evidence for shaking

219 Shaking-triggered turbidites at Puyehue Lake (Fig. 1b), in the foothills of the Andes
220 and about 150 km northeast from Cocotué, provide evidence for nine earthquakes
221 comparable to the 1960 event during the last 10000 years (Moernaut et al., 2007). Although
222 shaking had recurred at irregular intervals, spanning 1000 years on average, the two
223 youngest turbidites were linked to the historical earthquakes of 1575 and 1960.

224 By contrast, turbidites off the Chilean coast between 38°S and 40°S suggest a more
225 frequent earthquake recurrence, every 100-200 years, over the last glacial period (Blumberg
226 et al., 2008). Climate, sea-level change and proximity to the seismic source would explain
227 such a short recurrence interval by promoting the triggering of offshore turbidites, which
228 likely recorded earthquakes of different sizes. Similar influences were inferred by
229 Bernhardt et al. (2015) when studying offshore turbidites in central Chile, between 29°S
230 and 38°S. Although two of the Bernhardt et al. (2015) cores are located in the northern end
231 of the 1960 region, the scarcity of planktonic foraminifera in the hemipelagic intervals did
232 not allow a well resolved chronology to compare with other available paleoseismic records.

233 The 1575, 1837 and 1960 earthquakes have been invoked to explain a series of three
234 turbidites in another marine record at Reloncaví Fjord (Fig. 1d), 120 km east of Cocotué
235 (St-Onge et al., 2012). Turbidites were linked to those historical dates through average

236 sedimentation rates. A fourth and deeper turbidite was attributed to an earthquake around
237 AD 1400.

238 Turbidites found in lakes Villarica, Calafquén and Riñihue (Fig. 1b), along the
239 northern half of the 1960 region, suggest variation in rupture mode over the last 900 years
240 (Moernaut et al., 2014). The turbidites were stratigraphically correlated within each of the
241 three lakes. Successive turbidites differed in spatial extent and thickness, characteristics
242 interpreted as products of local earthquake intensity. Although lacking evidence from the
243 southern half of the 1960 rupture area, the four turbidites were ascribed to earthquakes that
244 ruptured its entire along-strike length in AD 1083-1171, 1310-1332, 1575 and 1960. This
245 turbidite sequence improves upon the less precisely constrained ages for the prehistoric
246 events D and C in the Maullín earthquake history (Cisternas et al., 2005). Additionally,
247 intervening smaller turbidites in one basin of Calafquén Lake were interpreted as three
248 smaller events, in AD 1280-1308, 1460-1470 and 1544-1548 (Moernaut et al., 2014). Of
249 the three, only the AD 1460-1470 has a counterpart in another lake, at Riñihue. Two
250 younger and smaller turbidites were linked to the historical earthquakes of 1737 and 1837.
251 The extent and thickness of these turbidites, compared with other paleoseismic and
252 historical records, suggest a northern rupture in 1737 and a southern rupture in 1837.

253 **4. Methods**

254 4.1. Stratigraphy

255 We found the Cocotué exposure during the austral summer of 2006 and studied it
256 further during the summers of 2007, 2008 and 2012. Using picks and shovels we cleaned an
257 outcrop 500 m long. Stratigraphic logs were made every 10 m for most of the exposure and

258 every 1 m at debris-flow fans. Grain size and texture were estimated in the field by hand
259 lens.

260 Airphotos taken before and soon after the 1960 earthquake aided in developing the
261 interpretations cartooned in Figure 2. Pre-earthquake oblique airphotos were taken by the
262 United States military in 1944 and post-earthquake vertical airphotos were taken by the
263 Chilean airforce in January 1961 (Figs. 1d, 3).

264 4.2. Topography and sea level

265 Height differences among features in the outcrop were measured during each field
266 season with a tripod-mounted level. The points surveyed in this manner were referred to a
267 common benchmark by means of third-order leveling with a closure error of 1 cm.

268 In January 2012 we linked the benchmark level to mean sea level. We estimated
269 mean sea level at Cocotué by measuring the water level continuously for 10 hours with a
270 portable acoustic tide gauge that we referenced to the benchmark. The resulting tide record,
271 obtained at a time of fair weather, was fitted to tidal predictions from the TPXO 8-atlas
272 tidal model, which is based on harmonic constituents extracted from 14 years of
273 TOPEX/Poseidon satellite altimetry (Egbert et al., 1994; Egbert and Erofeeva, 2002;
274 Melnick et al., 2012). The inferred total error in the estimate of mean sea level at Cocotué
275 was less than 0.1 m.

276 4.3. Dating by radiocarbon and introduced pollen

277 Accelerator mass spectrometry (AMS) radiocarbon analysis provided ages for the
278 plants flattened and buried by the debris flows. Well-preserved plant remains, mainly of
279 pineapple-like bromeliads (*Greigia sphacelata*) and rushes (*Juncus* sp.) were found bent

280 seaward, in about the same direction as the debris-flow runout in 1960. We collected the
281 samples by detaching the overlying diamict from the underlying soil (Fig. S2). Living *G.*
282 *sphacelata* plants, which still grow on the debris-flow fans, have hard and thorny leaves—a
283 characteristic that likely promoted their preservation. Other dated organic material included
284 a delicate twig, a small root of a bush likely overrun by the debris flow, and a piece of
285 wood within the second-youngest sand sheet. All the radiocarbon ages were calibrated to
286 years AD at two standard deviations. For calibration we used the Southern Hemisphere data
287 set of Hogg et al. (2013) and the software Calib Rev 7.0
288 (<http://www.calib.qub.ac.uk/calib/>).

289 We used pine pollen to assess the age of sediment deposited in the late 19th and
290 early 20th centuries (Cisternas et al., 2001). Chile had no pine trees until *Pinus radiata* was
291 first planted in the Concepción area in 1885 (Aztorquiza, 1929; Donoso and Lara, 1996). It
292 expanded early in the 20th century under a policy of the national government (Contesse,
293 1987). We therefore assume that pine pollen at Chiloé became abundant no earlier than
294 1900. We looked for pine pollen in the upper soils of the Cocotué stratigraphy by analyzing
295 1-g samples taken 1 cm apart. Samples were processed with standard palynological
296 techniques (Faegri and Iversen, 1975; Dupré, 1992). Pollen concentrations, in grains per
297 gram, were measured with the methods of Anderson (1974) and Kempt et al. (1974).

298 4.4. Diatom paleoecology

299 Fossil diatoms in coastal sedimentary sequences help to reconstruct past relative sea
300 levels and determine sediment provenance (Hemphill-Haley, 1996; Dawson, 2007; Dura et
301 al, 2016). Along the Cocotué's outcrop we selected an undisturbed stratigraphic column to
302 be sampled for diatoms (location in Fig. 4b). Diatoms were extracted, using standard

303 preparation methods (Palmer and Abbott, 1986), from 0.5 cm thick samples from the soils
304 and sand sheets that bracket them. We inspected slides at a magnification of x1000,
305 identifying at least 250 diatom valves per sample. Identifications followed Rivera and
306 Valdebenito (1979), Hartley et al. (1996) and reference collections held by Durham
307 University. We inferred the paleoecology of the assemblages by grouping species into five
308 categories based on their salinity preference, in the manner of Lowe (1974), Hemphill-
309 Haley (1993), and van Dam et al. (1994).

310 **5. The Cocotué stratigraphy**

311 The overall stratigraphy along the 0.5-km-long Cocotué outcrop includes four
312 nearly continuous, mostly tabular units of fine clean sand, each averaging about 10 cm in
313 thickness. These sand sheets alternate with dark organic soil layers, ≥ 10 cm thick, which
314 give the exposure a horizontally banded appearance (Fig. 4). This sequence rests on a basal
315 unit of beach sand. Since surveyed, the exposure has been eroded by waves and partially
316 collapsed.

317 Sharp contacts mark the boundaries between each sand sheet and underlying soil.
318 These four distinct and continuous sand-soil contacts, from youngest (top) to oldest
319 (bottom), are termed here contacts A, B, C, and D (Fig. 4). One of the four, contact C, is
320 distinctive lithologically because the sand above it contains pebbles of pumice (Figs. S4,
321 S5a). The youngest of the four, contact A, is distinctive paleontologically as the only
322 contact that overlies pine pollen (Fig. S3b).

323 Diamicts, composed of unsorted mixtures of clay, silt, sand, and gravel, locally
324 interrupt sand sheets above contacts A, B, C, and D. The largest clasts within the diamict

325 are cobbles and boulders as much as 40 cm in diameter. These are well rounded and a few
326 of them are so weathered that they can be cut with a knife. All these unsorted mixtures are
327 also exposed in a landslide scarp high near the top of the former sea cliff, pointing to the
328 cliff as the source of the diamicts. The diamicts themselves form lenticular beds that
329 coincide with the toes of two conspicuous debris-flow fans (Fans N and S; Fig. S1).
330 Although the diamicts locally interrupt the continuity of the sand-soil contacts, becoming
331 diamict-soil contacts, the lack of erosion on the tops of the soils, evidenced by the presence
332 of growth-position flattened plants (Fig. S2), persuaded us to keep the contact name
333 through the diamict-soil contacts, even though they likely mark a different process.

334 Four less distinct and discontinuous additional contacts, each capped by a sand sheet
335 and three of them also by diamicts, are present locally within the soils beneath contacts A,
336 B, and C. Most of the less conspicuous contacts are associated with fan S but some are also
337 found at fan N. The stratigraphically oldest of the additional contacts, termed cd, is between
338 contact C and D. Two others, contacts bc1 and bc2, are between contact B and C. Contact
339 ab, with a distinctively pink soil, underlies both contact A and the earliest (highest)
340 occurrence of pine pollen (Figs. S3b, S5).

341 5.1. Basal beach deposits

342 Beach deposits thicker than 2 m underlie the eight contacts described below (Fig. 4).
343 We ascribed this basal unit to beach because it contains scattered marine shells and heavy
344 mineral laminae similar to those in the modern foreshore beach. On this old beach an
345 organic dark soil developed. As the soil contains mostly freshwater, salt-intolerant diatoms
346 and was vegetated with plants before burial (Figs. S2, S3), it probably resulted from a
347 significant land-level change that raised the beach out of the tidal zone. The soil likely

348 developed shortly before 1000 years ago, when its plants were buried (see age of contact D
349 below). Freshwater species dominated diatom assemblages in the overlying sediments (Fig.
350 S3c), implying that the terrace has never returned to the tidal reaches since the formation of
351 the soil on the beach. Despite this broad trend, during the last 1000 years the freshwater,
352 salt-intolerant diatoms have diminished upwards in favor of diatoms that tolerate or are
353 stimulated at low salinity. This subtle trend would suggest either a relative sea-level rise, a
354 net landward advance of the shore that exposed the terrace to salt spray, or both.

355 5.2. Contact D

356 Contact D marks the boundary between the soil developed on the old beach and
357 both sand and diamict deposits (Fig. 4b). The contact is clearly delineated by a ~10-cm
358 thick tabular sand sheet, which rests on the soil along the entire 500-m outcrop and overlaps
359 both pinched-out edges of the diamict lens. Freshwater diatoms predominate in this sand
360 sheet (Fig. S3c). The diamict lens, located at fan N, is the second largest in exposed area at
361 Cocotué. No coeval diamict is evident at fan S.

362 Plants flattened and buried by the diamict at fan N date contact D to AD 907-1149
363 (*Juncus* sp.) and AD 862-1131 (*G. sphacelata*). Their pooled mean age, 1083 ± 31 14C yr
364 BP, corresponds to AD 898-1128 at 2σ (Fig. 5; Tables 1, S1).

365 5.3. Contact cd

366 Contact cd is defined only by a sand-soil discontinuous boundary. No diamict lens
367 was observed interrupting the sand sheet or occurring at the same stratigraphic level. The
368 thin (~3 cm) tabular and discontinuous sheet of clean sand is found 3-5 cm below contact
369 C. Its clearest exposure is at fan S, where it extends continuously for ~4 m below the

370 diamict that marks contact C (Fig. 4c). Contact cd also occurs discontinuously at two other
371 locations: ~25 m northward from the first place in fan S, and hundreds of meters northward,
372 in fan N, below the diamict of contact C (Fig. 4b).

373 We were not able to find adequate material to date contact cd; however, by its
374 position, between D and C, contact cd postdates AD 898-1128 and predates AD 1300-1398
375 (see age of contact C below).

376 5.4. Contact C

377 Contact C is clearly marked by both sand-soil and diamict-soil boundaries. While
378 the sand sheet is thinner (~4 cm) and broken in fan N, as it is probably trampled (Figs. 4b,
379 S3), it is thicker (~10 cm) and continuous in fan S (Fig. 4c). The sand overlaps only the
380 lower pinched edge of the diamict in fan N but both edges in the diamict of fan S. Of all the
381 sand sheets at the outcrop, the sand over contact C contains the largest proportion of marine
382 diatoms (30% of the total assemblage; Fig. S3c). Remarkably, diamict lenses over contact
383 C occur at both of the debris-flow fans. The diamict at fan N is laterally longer than the one
384 in fan S (Figs. 4b, 4c).

385 A distinctive but discontinuous ~5-cm-thick layer of rounded, gravel-sized pumice
386 (1-5 cm in diameter) conformably caps the sand sheet over contact C at different places but
387 mainly at fan S (Figs. 4a, 4c, S4). No soil nor erosion is evident between the sand and the
388 pumice. Additionally, no other conspicuous pumice layer is evident in the 1000-year-long
389 stratigraphic record. These absences suggest a closeness in time between the pumice
390 deposition and the process that produced contact C. A short discussion about the pumice

391 sources and a possible volcanic eruption associated to event C is in the Supporting material
392 (Text S1).

393 The diamict over contact C, at fan N, flattened a twig from AD 1300-1420 and a *G.*
394 *sphacelata* leaf from AD 1286-1394. The pooled mean age constrains contact C to AD
395 1300-1398 (Tables 1 and S1).

396 5.5. Contacts bc1 and bc2

397 Two closely spaced contacts, bc1 and bc2, are overlain by both sand and diamict
398 deposits between contacts C and B. They are delineated by two thin (~3 cm) sand sheets
399 that pinch out on the flanks of the soil-covered diamict that overlies contact C in fan S (Fig.
400 4c). The sandy pair also occurs in the middle of fan S and at the southern end of fan N. A
401 pair of overlapped diamict lenses likely also mark contacts bc1 and bc2 between contact C
402 and B. They show an architecture similar to that of fan N: the younger diamict occupies an
403 area beside the higher ground created by the preceding diamict (Figs. 4c and S1d). We
404 could not trace laterally the sand-soil contacts with the diamict-soil contacts; however, their
405 similar stratigraphic level and spatial closeness strongly suggest their correlation.

406 By their position between contacts C and B, contacts bc1 and bc2 postdate AD
407 1300-1398 and predate AD 1411-1625 (see age of contact B below). However, radiocarbon
408 dates obtained for bc1 and bc2 overlap those of the bracketing contacts (Table S1). The
409 earlier diamict in the couplet, resting on contact bc1, flattened and buried a *Juncus* stem
410 from AD 1304-1435, an age mostly indistinguishable from that of the preceding contact C
411 (AD 1300-1398). Likewise, the later diamict, resting on contact bc2, flattened a *Juncus*

412 stem from AD 1505-1802, a range that overlaps the age of the overlying contact B (AD
413 1412-1625).

414 5.6. Contact B

415 Contact B is defined by both sand-soil and diamict-soil boundaries. This contact is
416 clearly marked by a ~8-cm thick tabular sand sheet, which rests on the soil along the entire
417 outcrop. The sand sheet overlies only a distal edge of the diamict lens at fan N. Like contact
418 D, contact B is marked by only one diamict lens.

419 The age of contact B is limited by a woody root that was 2 cm below the surface of
420 the soil buried by the diamict in fan N. If the root was part of a plant killed by the debris
421 flow, the contact dates to AD 1412-1625 (Tables 1 and S1).

422 5.7. Contact ab

423 The penultimate contact at Cocotué is overlain by a relatively thick (2-6 cm) and
424 continuous sand sheet that lies 4-8 cm below contact A at fan N (Figs. 4b, S3, S5c), and by
425 diamict lenses in fan S. Both the diamicts and the sand overlie a conspicuous change in
426 color in the soil (pink when wet, paler when dry; Fig. S5). On this basis, the sand at fan N
427 was correlated with two small and one medium-size diamicts in fan S (Fig. 4c). The
428 diamicts and sand sheet are stratigraphically closer to contact A than to contact B, which
429 suggests a temporal proximity between those deposits and the process that produced contact
430 A.

431 A piece of wood included in the sand sheet gave a maximum calendar age of AD
432 1505-1949, with most of the probability density between AD 1700 and 1800 (Tables 1 and
433 S1). A further minimum age for contact ab is provided by *P. radiata* pollen (Fig. S3b).

434 Because the sand sheet is ~4 cm below the soil that contains the deepest pine pollen,
435 contact ab should predate AD 1900 by several decades, when widespread planting of *P.*
436 *radiata* began in southern Chile.

437 5.8. Contact A

438 The youngest contact at Cocotué is clearly marked by both sand-soil and diamict-
439 soil boundaries. It underlies a widespread and thick (~7-12 cm) sand sheet and the largest
440 diamict lens (Fig. 4). The sand sheet is continuous through the entire outcrop and at its
441 northeastern end is mixed with or overlapped by eolian sand, reaching almost 1 meter thick.
442 The sand sheet overlaps the southern edge of the diamict. Remarkably, this diamict extends
443 laterally more than 30 m and, while lapping onto older diamicts, projects mainly to the
444 south of them. The age of this contact is discussed below, in section 6.2.

445 **6. Interpretations and correlations**

446 6.1. Stratigraphic interpretations

447 We interpret each of the Cocotué sand sheets as a tsunami deposit. We invoke
448 tsunamis, instead of tidal flat, eolian or storm washover deposits, for reasons similar to
449 those used at Maullín (Cisternas et al., 2005): similar sand grain size and texture to those of
450 the sand composing the modern beach, widespread tabular shape, abrupt and continuous
451 lower contacts, and absence of bioturbation in the underlying soils. We also discount the
452 alternative hypothesis of tidal-flat deposition because the terrace stood above high-tide
453 level even after subsiding in 1960.

454 Diatoms associated with the sand sheets neither confirm nor refute a tsunami source
455 (Fig. S3c). Freshwater assemblages predominate in most of the sand sheets. The

456 predominance of freshwater species need not imply a terrestrial sediment source, however,
457 because the former cliff lacks clean sand similar to that of the sand sheets. We hypothesize
458 that sand derived from the beach may have contained few diatoms to begin with, while
459 post-depositional freshwater ponding and infiltration may have introduced freshwater
460 diatom taxa into the sand sheets.

461 We interpret each of the diamict lenses as an earthquake-triggered debris-flow
462 deposit. This interpretation is supported by their mix of clast types and their coincidence
463 with fans at the bases of debris-flow channels. Airphotos taken eight months after the 1960
464 earthquake show fan N covered by a fresh, large debris flow (Fig. 3b). Furthermore, the
465 1960, 1837 and 1575 earthquakes are known to have produced abundant slope failures in
466 south-central Chile (Weischet, 1963; Davis and Karzulovich, 1963, Cisternas et al., 2005).

467 Where a diamict lens and a sand sheet buried the same soil, and where the sand
468 extends onto unweathered debris-flow deposits, we infer that a tsunami was triggered
469 during the same earthquake that set off the debris flow (Fig. 2).

470 When a laterally large diamict lens is associated with a continuous and relatively
471 thick sand sheet, comparable with those assigned to the 1960 earthquake and tsunami (see
472 below), we simplistically infer that they were more likely created by a 1960-type
473 earthquake and tsunami. When the diamict lens is smaller and the sand sheet is
474 discontinuous and thin, we infer that they more likely resulted from a relatively smaller or
475 perhaps more distant event. In making a final interpretation of the relative sizes of the
476 events at Cocotué, we also considered the latitudinal extent of correlative deposits at other
477 sites.

478 Given these interpretations, and the corresponding radiocarbon ages, the series of
479 eights contacts gives evidence of four possible 1960-type earthquakes and tsunamis (from
480 old to young, events D, C, B, and A), and of four intervening relatively lesser events (cd,
481 bc1, bc2, and ab) in the span of about 1000 years.

482 6.2. Chronological correlation with well-dated evidence elsewhere

483 In most cases the ages of the plants killed by debris flows predate the inferred
484 earthquake by no more than a few years. This geological precision facilitates correlation
485 with well-dated coastal and subaqueous evidence elsewhere and with historical accounts
486 (Tables 1 and S1; Fig. 5).

487 The oldest 1960-type earthquake and tsunami recorded in Cocotué (event D),
488 marked by the lowest debris-flow deposit and associated sand sheet, occurred in AD 898-
489 1128. Notably this age coincides with evidence for a 1960-type earthquake and tsunami at
490 Maullín (Cisternas et al., 2005) and Chucalén (Garrett, et al., 2015), and for a comparable
491 earthquake in Calafquén Lake (Moernaut et al., 2014). The Cocotué age range overlaps the
492 age of two rushes killed by tidal submergence at Maullín, the modelled age of submergence
493 at Chucalén and the age of the oldest large turbidite in Calafquén Lake (Fig. 5; Tables 1 and
494 S1).

495 Sometime after event D, a relatively smaller tsunami was recorded in Cocotué (cd).
496 Although we could not date it, by its stratigraphic position, between D and C, it postdates
497 AD 898-1128 and predates AD 1300-1398. Interestingly, a local turbidite in Calafquén
498 marks a smaller earthquake that coincides, by stratigraphic position and age range, with cd
499 (Tables 1 and S1). Event cd likely also correlates with a smaller tsunami recorded in

500 Maullín. There, at the seaward end of two perpendicular transects, a thin 1-cm sand sheet is
501 exposed between Maullín events D and C (Figs. S2, S3 of Cisternas et al. (2005)). In both
502 stratigraphic level and thickness, the thin sand sheet at Cocotué greatly resembles the sand
503 sheet at Maullín.

504 The age of the next prominent earthquake and tsunami at Cocotué is AD 1300-1398
505 (event C), probably several decades after cd. This date overlaps the ages of plants killed by
506 subsidence in Maullín and in Caulle (Atwater et al., 2013) and an inferred age in Chucalén.
507 The age of C also overlaps the date of the second oldest large turbidite of Calafquén Lake
508 and the oldest of Riñihue Lake (Moernaut et al., 2014; Tables 1 and S1).

509 Following event C, a pair of lesser earthquakes and tsunamis left less distinct layers
510 in Cocotué (bc1 and bc2). Notably, turbidites in Calafquén show a similar pattern with
511 evidence of a pair of smaller earthquakes between the events linked to events C and B at
512 Cocotué (Table 1). Radiocarbon dating of the earlier event in Cocotué, bc1, yielded an age
513 (AD 1305-1435) that is indistinguishable from that of event C, while the age of the later
514 bc2 event (AD 1505-1802), is also indistinguishable from that of event B. The similarity in
515 ages together with their stratigraphic proximity (see Fig. 4c) suggests events C, bc1, bc2
516 and B all occurred close in time. Varve counting in Calafquén Lake supports this inference.
517 The counterpart of bc1 in Calafquén was dated at AD 1460-1470, roughly a century after
518 event C, and bc2 was dated at AD 1544-1548, a few decades before the occurrence of event
519 B.

520 Event B, another 1960-type earthquake and tsunami, struck Cocotué in AD 1412-
521 1625 (Tables 1 and S1). This age range overlaps with ages of the plant remains that
522 underlie the penultimate tsunami deposit at Maullín and the penultimate large turbidites in

523 Calafquén, Riñihue and Villarrica lakes (Moernaut et al., 2014). Event B at Cocotué likely
524 represents the giant and well-documented Chilean earthquake and tsunami of 1575.

525 The penultimate event recorded in Cocotué (ab), with a maximum age of AD 1505-
526 1949 and most of its probability density between AD 1700 and 1800, represents a relatively
527 smaller earthquake and tsunami. This event likely corresponds to the south-central Chile
528 earthquake and trans-Pacific tsunami of 1837. While the historical earthquake of 1737 also
529 falls within the age range, there is no known report of a tsunami either in Chile or Japan in
530 1737. Further, written records suggest the effects of the 1737 earthquake were focused in
531 the northern half of the 1960 region, around Valdivia farther north of Cocotué, while the
532 1837 effects were more evident in the southern half, closer to Cocotué. In addition, the
533 stratigraphic position of contact ab is closer to the upper contact than to the lower 1575
534 contact, suggesting a temporal proximity to 1960. Finally, because the tsunami deposit is
535 around 4 cm below the soil that contains the deepest *P. radiata* pollen, event ab likely
536 occurred only several decades before AD 1900.

537 If contact ab records the 1837 earthquake and tsunami, the 1737 earthquake is the
538 only case, in a 1000-year comparison between Cocotué and the lacustrine record farther
539 north, of an incomplete match. From north to south, lakes Villarica, Calafquén and Riñihue
540 recorded both 1737 and 1837 earthquakes (Moernaut et al., 2014). However, turbidites
541 suggest that the 1737 shaking was stronger in the northernmost lake, Villarica, than in the
542 lakes to the south. This agrees with the northern location of the 1737 earthquake indicated
543 by written records (Cisternas et al., 2005). Thus, the lack of 1737 evidence in Cocotué is
544 consistent with a more northerly location.

545 Finally, we infer that event A corresponds to the 1960 earthquake and tsunami.
546 Three lines of evidence support this assumption. First, post-1960 airphotos show both a
547 fresh landslide scarp on the slope directly above fan N and its resulting debris-flow deposit
548 resting on the top of the fan itself (Fig. 3b). Second, the soil beneath the correlative sand
549 sheet contains pollen of the introduced *P. radiata*, making the debris-flow and tsunami
550 deposits younger than AD 1900 (Fig. S3). Third, the 1960 mainshock and associated
551 tsunami remain abundantly recorded by geologic evidence of tsunami at Maullín, Caulle
552 and Chucalén, and by evidence of shaking in lakes Calafquén, Riñihue and Villarrica.

553 **7. Discussion and synthesis**

554 7.1. Greater sensitivity at Cocotué than at nearby coastal sites

555 The coastal stratigraphy presented above reveals a sequence of four giant
556 earthquakes and tsunamis that likely matches one-for-one with the events dated within the
557 last 1000 years in nearby coastal sites, at Maullín (Cisternas et al., 2005) and Chucalén
558 (Garrett et al., 2015). Additionally, the Cocotué stratigraphy recorded four relatively
559 smaller events, perhaps 1837-type, that match evidence for four smaller earthquakes in
560 three Andean lakes 270-330 km to the northeast (Moernaut et al., 2014).

561 Unlike Maullín and Chucalén, the ability of Cocotué to record smaller events
562 probably reflects its exposure to open water together with a physiographic setting that
563 promotes the preservation of thin sand layers. Cocotué is more directly exposed to tsunamis
564 (Fig. 1c). While Cocotué faces the open Pacific, Maullín and Chucalén fringe inland
565 waters. Maullín is 8 km upstream along a tidal river and Chucalén is in a semi-enclosed
566 inland water body.

567 Although open to the sea, the Cocotué sequence was deposited above the highest
568 tides, as judged from its post-1960 elevation and from diatom assemblages, which indicate
569 that sediments accumulated above the reach of tides throughout the last 1000 years. This
570 elevation prevented tides and perhaps storms from reworking the thin sand sheets and the
571 smaller lenses of diamict. Similarly at Maullín, the 1960 sand sheet is preserved best where
572 pasture soils beneath it barely entered the intertidal zone upon subsiding in 1960. Sand on
573 lower surfaces was widely reworked by waves at high tide that have stripped the underlying
574 soil as well (Cisternas et al., 2005). Cocotué escaped such reworking, despite having
575 subsided about a meter in 1960.

576 Preservation of the evidence of the smaller events at Cocotué was also likely
577 promoted by later debris-flow deposition. The thin sand sheet marking contact cd was later
578 covered by the debris flow of contact C on both fans S and N (Fig. 4). Contacts bc1 and
579 bc2, as a pair, are similarly buried by the debris flow that caps contact ab in two places on
580 fan S, and the sand sheet on contact ab is protected by the 1960 diamict in fan N. Thus,
581 these thin surviving sheets coincide mainly with the debris-flow fans. It is likely that
582 diamicts, being both thick and hard, shielded the underlying sand sheets from later
583 bioturbation and erosion.

584 7.2. Sources and relative sizes of the largest earthquakes between AD 1000 and 1960

585 While the new findings are consistent with the previously assumed megathrust
586 origin for the three giant pre-1960 earthquakes, namely D, C and B (1575), their absolute
587 size remains ambiguous due to the lack of geologic and historical evidence from the
588 southern half of the 1960 region (Fig. 6).

589 The latitudinal extent of the geologic evidence for each of the three giant
590 predecessors to 1960 in Cocotué, Chucalén, Maullín and Calafquén Lake suggests fault
591 rupture at the boundary between the Nazca and South American plates. No intraplate fault
592 could explain strong shaking for 300 km between Cocotué and Calafquén or coastal land-
593 level changes for 350 km between Maullín and Tirúa, in 1575 (Fig. 6b). The tsunami
594 evidence at Cocotué, Chucalén, Maullín, Caulle, and Tirúa also points to megathrust
595 ruptures that warped the floor of the continental shelf and slope to their west.

596 Could the size of the debris-flow deposits at Cocotué be indicative of the relative
597 size of the earthquakes? Among these deposits, only those of event D rival the one from
598 1960 in thickness and continuity along the 500-m outcrop. The size of the 1960 diamict,
599 however, was perhaps enhanced by the land-use change and the season when the event
600 occurred. Spanish colonists initiated widespread clearing of the native rain-forest on
601 Chiloé, promoting soil erosion and slope instability (Torrejón et al., 2004). This process
602 significantly increased and accelerated after Chile's independence in 1810 (Torrejón et al.,
603 2011). Additionally, the 1960 earthquake occurred during the austral winter, a very rainy
604 season at Chiloé. By contrast, the 1575 earthquake occurred early in the era of land clearing
605 and during the drier austral summer.

606 Written evidence for far-field tsunamis in Japan, as in 1837 and 1960, could help to
607 define the size of the predecessors to the 1960 earthquake. However, no such evidence has
608 been found in Japan for the 1575 Chilean tsunami (Ninomiya, 1960; Iida et al., 1967;
609 Soloviev and Go, 1984; Watanabe, 1998), despite Chilean written and geologic evidence
610 indicating that the effects of the 1575 earthquake resembled those in 1960. Similarly, there
611 are no accounts in Japan of flooding and damage at times corresponding to the earlier

612 Cocotué events D or C. This lack of Japanese written records is inconclusive, however,
613 because during most of the time interval spanned by events D, C and B (1575), Japan
614 produced few records of natural disasters as the Japanese were immersed in feudal wars
615 (Ueda and Usami, 1990). Still, at least one written record of flooding north of Sendai, at
616 Minami Sanriku, was linked to the 1586 tsunami from Peru. Notably, the 1960 Chilean
617 tsunami was as much as 5 m high in this area (Ninomiya, 1960; Watanabe, 1998).

618 7.3. Sources and sizes of the relatively smaller prehistoric earthquakes and tsunamis

619 Rupture on the megathrust also best explains all three of the relatively smaller,
620 1837-type, prehistoric events (cd, bc1, and bc2) evidenced at Cocotué and in Calafquén
621 Lake. We consider other offshore sources, such as outer-rise or splay fault ruptures, as
622 unlikely, because the earthquakes were able to produce strong shaking not only on the coast
623 but also in the far inland Andean lakes; a result difficult to explain through secondary
624 offshore faults.

625 The earliest 1837-type event (cd) is marked by a tsunami deposit in Cocotué and
626 likely in Maullín. However, no evidence for coincident shaking was found in Cocotué. This
627 points to a more distant source, likely in the northern half of the 1960 region where the
628 Calafquén Lake did record correlative shaking. Alternatively, dry conditions not favoring
629 slope instability at Cocotué at the time of the earthquake could also explain such absence.

630 The pair of events bc1 and bc2, between events C and 1575, is marked by evidence
631 for both shaking and tsunami at Cocotué and for shaking at Calafquén. In this respect each
632 of them resemble the 1837 earthquake (Fig. 6).

633 7.4. Notably short earthquake recurrence intervals between AD 1300 and 1575.

634 The earthquake recurrence intervals among C, bc1, bc2, and 1575 are remarkably
635 short (Table 1; Fig. 6c). If the date range for event C as indicated by the turbidite record is
636 correct (AD 1310-1332), and event B is in fact 1575, the middle and northern thirds of the
637 1960 region were struck by at least four megathrust earthquakes and four ensuing tsunamis
638 in just two and a half centuries.

639 The average recurrence interval for this period (85 yr) is shorter than the historical
640 average in the series 1575, 1737, 1837, 1960 (128 yr) and about three times shorter than the
641 earthquakes geologically recorded at Maullín (285 yr). Furthermore, the interval between
642 event bc2 and 1575 appears to be even shorter than the roughly 85-year average. Correlated
643 with a Calafquén turbidite, bc2 is bracketed by AD 1544-1548. It would imply that the
644 giant 1575 event was preceded by a 1837-type event (bc2) only 30 years before. Although
645 close in time to the foundation of Valdivia in 1552, the first Spanish settlement south of
646 Concepción (Cisternas et al., 2012), bc2 was probably a few years too early to have entered
647 written history through Spanish colonists.

648 Alternatively, the AD 1300-1575 recurrence story is oversimplified if events bc1
649 and bc2 ruptured different portions along the area broken in 1960, as likely occurred in
650 1737 and 1837. While the rupture in 1737 focused in the north, the one in 1837 was in the
651 center and south. If bc1 and bc2 behaved in a similar way, the AD 1300-1575 average
652 recurrence interval for a given portion of the megathrust would be longer than 85 years.
653 However, the pair bc1-bc2 left a geologic imprint far different from that of the 1737-1837
654 pair in Cocotué. While the 1737 event did not leave any record, the 1837 event is marked
655 by evidence of shaking (debris-flow deposit) and tsunami (sand sheet). In this respect, the
656 geologic record at Cocotué differentiates between these two ruptures. On the contrary, both

657 bc1 and bc2 are clearly and similarly marked by both shaking and tsunamis at Cocotué. It
658 strongly suggests that the bc1 and bc2 ruptures were both similar to that in 1837 and that
659 they occurred if not in the same area, at least nearby.

660 Consequently, the short bc2-1575 interval provides an important precedent for a
661 recurrence interval comparable to the 56 years that have elapsed since 1960. This inference
662 gives some cause for concern, because by 2011 the accumulated interplate moment deficit
663 showed that locked segments of the 1960 region were already capable of nucleating
664 earthquakes of magnitude 8 (Moreno et al., 2011).

665 **8. Conclusions**

666 The unusual coastal sedimentary record at Cocotué, which combines evidence for
667 seismic shaking and coincident tsunami inundation since AD 1000 in the region of the giant
668 1960 Chile earthquake, reveals differences in earthquake size and recurrence between
669 successive ruptures.

670 In the millennium preceding 1960, south-central Chile was struck by three
671 earthquakes that probably resembled the 1960 mainshock (in AD 898-1128, 1300-1398 and
672 1575) and by five intervening earthquakes that were likely smaller in fault-rupture area. All
673 but one, in 1737, generated obvious tsunamis. This inferred seismic history begins with a
674 giant, 1960-type, earthquake early in the 12th century AD. A relatively smaller earthquake
675 with a tsunami followed before the end of the 13th century. Soon after, early in the 14th
676 century, another giant earthquake was accompanied by a tsunami and may have occurred
677 close in time with a volcanic eruption as well. Two relatively smaller earthquakes and
678 tsunamis ensued between this event and the giant 1575 earthquake and tsunami. The

679 smaller historical events of 1737 and 1837, rupturing different portions of the plate
680 boundary, then failed to expend most of the accumulated plate convergence that contributed
681 to the enormity of the 1960 earthquake.

682 Since AD 1000, megathrust earthquakes and tsunamis recurred most often between
683 AD 1300 and 1575. The recurrence intervals averaged about 85 years, a little more than the
684 time already elapsed since 1960. This conclusion is of serious concern for two reasons.
685 First, no earthquake has been expected to occur so soon in the area struck only 56 years ago
686 by an earthquake as large as the 1960 event. Second, plate locking suggests that some
687 segments of the region are already capable of nucleating earthquakes as large as magnitude
688 8.

689 This millennial earthquake and tsunami history of one of the world's most active
690 subduction zones adds another example of variable rupture mode, in which earthquake size
691 and recurrence interval vary from one earthquake to the next.

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883 **Figure captions**

884 Figure 1. Index maps and pre-1960 aerial view of the study area. a) Plate-tectonic setting of
885 Chile. Purple paired arrows indicate plate convergence of 6.6 cm yr⁻¹ (Angermann et al.,
886 1999). b) Area of 1960 earthquake. Brown elongated ellipse outlines the area that subsided
887 tectonically in 1960 (Plafker and Savage, 1970). Black dots mark places mentioned in the
888 text. Blue dots mark Andean lakes studied by Moernaut et al. (2014). c) Location of
889 Cocotué. Dashed blue curve shows western limit of the last glacial advance (Porter, 1981).
890 d) Oblique 1944 air view of Cocotué and surroundings.

891

892 Figure 2. Schematic view of how seismic shaking and tsunami inundation were recorded at
893 Cocotué. a) An earthquake triggers a debris flow on a former sea cliff. b) The diamict
894 resulting from the debris flow covers the soil at the foot of the slope, where it flattens and
895 buries plants. c) An ensuing tsunami entrains sand as it comes ashore, and subsequently
896 deposits the sand on the toe of the debris-flow fan. d) A new soil develops on the tsunami
897 deposit and the diamict. e) The next cycle begins when another earthquake triggers another
898 debris flow.

899

900 Figure 3. Setting of the Cocotué outcrop shown by aerial views before and after the 1960
901 earthquake and tsunami (see also Figure S2). a) Oblique 1944 airphoto shows the slope of
902 the former sea cliff and the elongated freshwater pond at its foot. All three views show the
903 same fence line. In 1944, the sea stack known as Roca Runrun was surrounded by beach
904 sand. b) Vertical airphoto taken eight months after the 1960 earthquake shows a series of

905 fresh debris-flow channels and associated fans. Some of them are enlarged in the inset.
906 Yellow dotted line and small red arrows indicate the distal limits of the debris-fan fronts.
907 Coseismic subsidence (~1.5 m) and consequent shoreline retreat have left Roca Runrun in
908 the surf. c) An image from 2009 shows that most of the southern pond area had been eroded
909 and the shoreline had retreated to intersect the debris-flow fans. Dashed white line shows
910 the outcrop mapped in Figure 4a (image © Digital Globe).

911

912 Figure 4. Stratigraphy at Cocotué. a) Stratigraphic cross section of the entire surveyed
913 outcrop. b and c) Details of the toes of fans N and S. Most of the ages were measured on
914 the remains of plants in growth-position that had been flattened by debris flows.

915

916 Figure 5. Chronologic comparison, from north (top) to south (bottom), of evidence for
917 earthquakes and tsunamis along the length of the 1960 region. No data coverage south of
918 Chiloé. Written records from Lomnitz (1970), Cisternas et al. (2005) and Fig. 6 of this
919 report. Filled rectangles for Tirúa, Maullín, Caille, Chucalén and Cocotué show age ranges
920 of plant fossils at two standard deviations (Ely et al., 2014; Cisternas et al., 2005; Atwater
921 et al., 2013; Garrett et al., 2015). Cocotué ages from coastal sand sheets and diamicts,
922 interpreted as recording giant and merely great events (this report). Colors of the rectangles
923 match the events that have been previously correlated, by stratigraphy and dating, to the
924 events of Maullín—Blue: event D; purple: C; green: B (1575); red: A (1960). Gray dots for
925 Reloncaví indicate dates of earthquakes inferred from marine turbidites and average
926 sedimentation rates (St-Onge et al., 2012). Barred rectangles show dates and uncertainty

927 derived from varve counts for turbidites in Calafquén Lake (Moernaut et al., 2014). Basin-
928 wide and local turbidites interpreted as produced by large and smaller events respectively.

929

930 Figure 6. Graphical summary of the inferred 1000-year earthquake history along the length
931 of the 1960 region. a) Age sequence of the 1960-type and relatively smaller earthquakes
932 recorded at Cocotué compared with ages from Calafquén Lake (Moernaut et al., 2014) and
933 historical dates (Lomnitz, 1970). b) Plot of the stratigraphic and historical evidence for each
934 event. Geologic evidence from Ely et al. (2014), Moernaut et al. (2014), Cisternas et al.
935 (2005), Garrett et al. (2015) and this report. Historical evidence from Lomnitz (1970) and
936 Table S1 of Cisternas et al. (2005). c) Timeline of giant and great earthquakes that struck
937 south-central Chile during the last 1000 years. Age ranges from Calafquén Lake (Moernaut
938 et al., 2014).

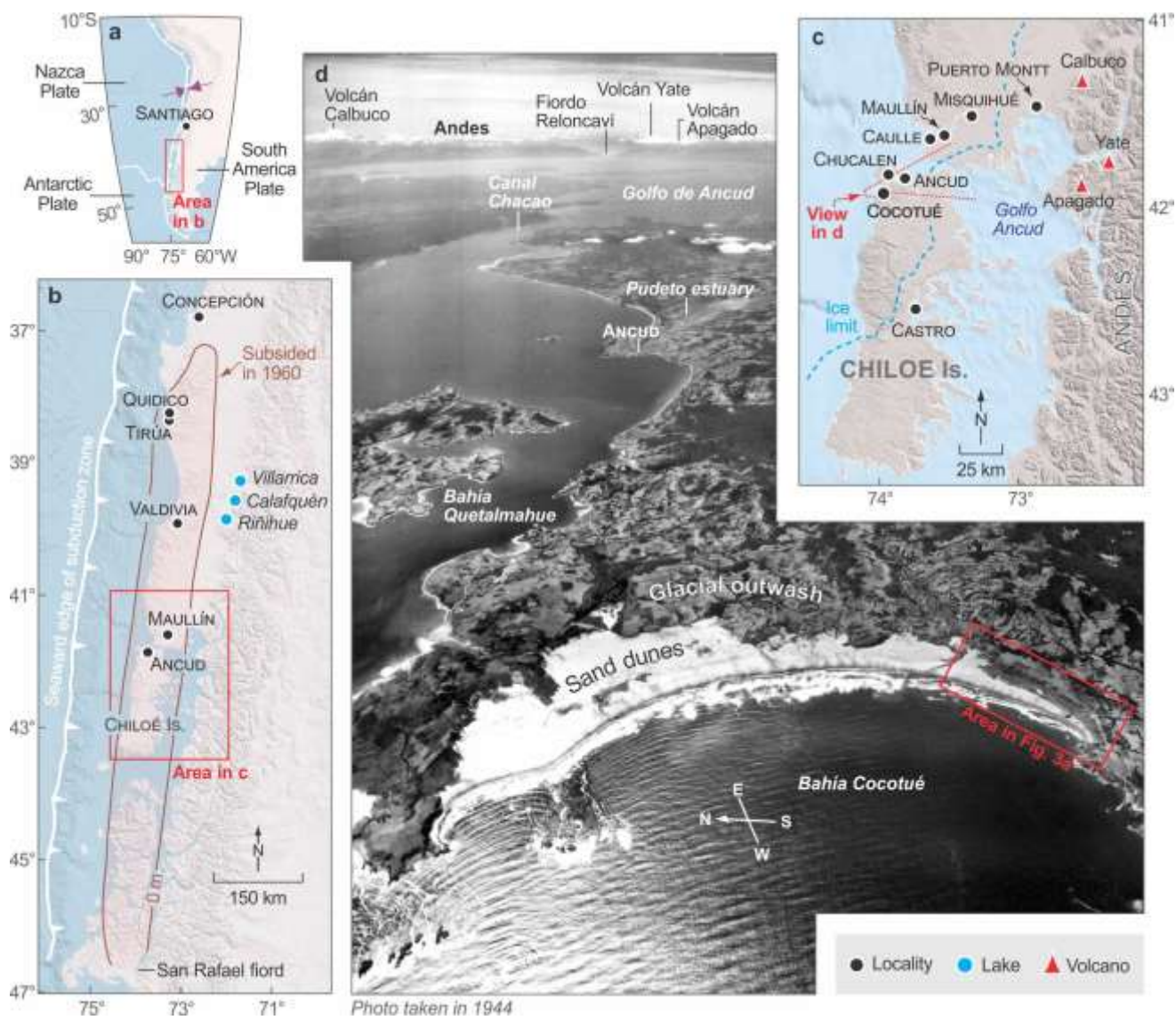
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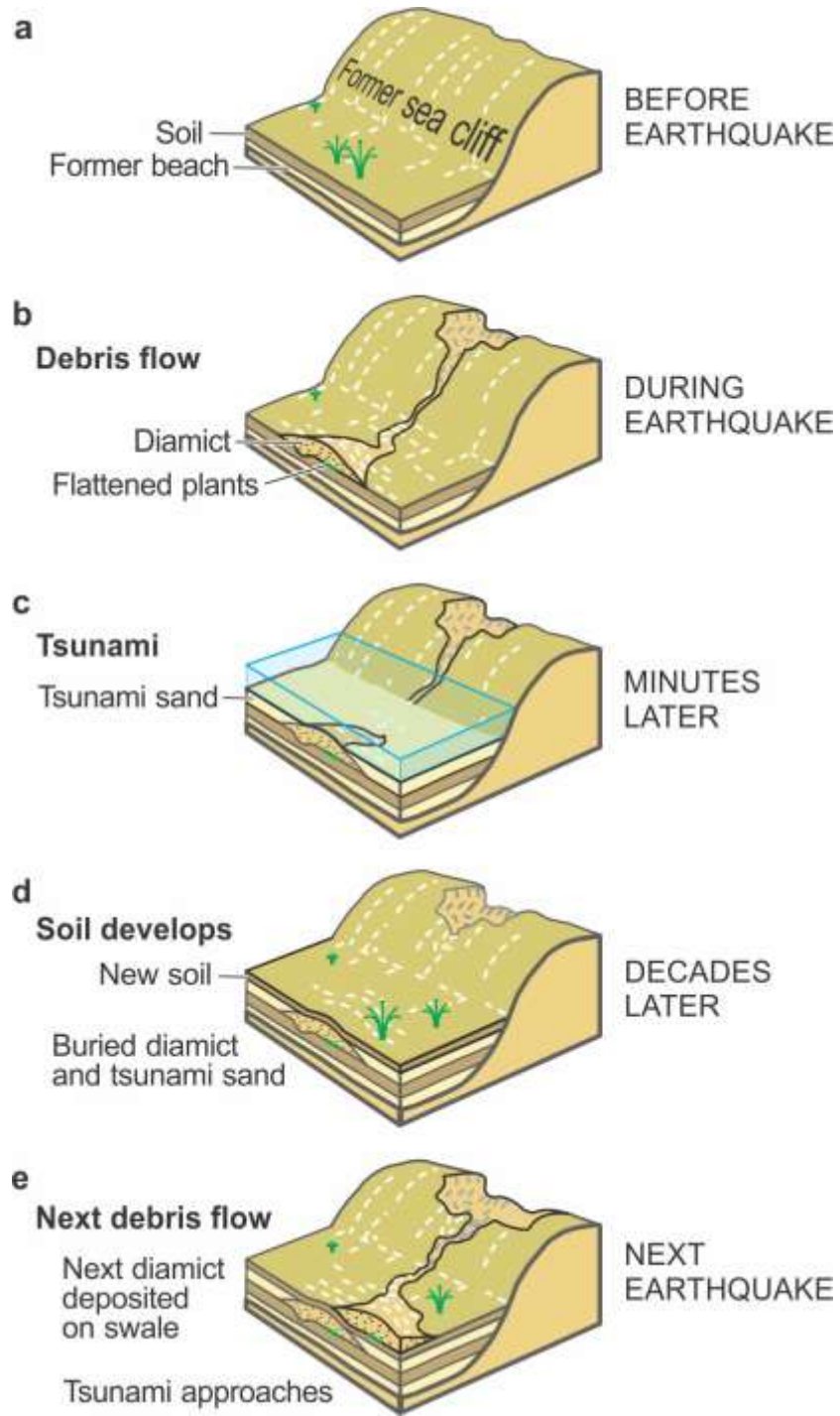
940 **Table caption**

941 Table 1. Stratigraphic and chronological correlation between the Cocotué events (this
942 report) and coastal (Cisternas et al., 2005; Atwater et al., 2013; Garrett et al., 2015) and
943 subaqueous lacustrine evidence (Moernaut et al., 2014) elsewhere. Bold and regular
944 lettering denote events interpreted as giant and relatively smaller earthquakes respectively.
945 Asterisks indicate dates inferred from stratigraphy, written records, or airphotos. While
946 ages from Cocotué Maullín, Caille, and Chucalén were obtained through radiocarbon
947 analyses, the Villarica, Calafquén, and Riñihue lakes were derived counting varves.

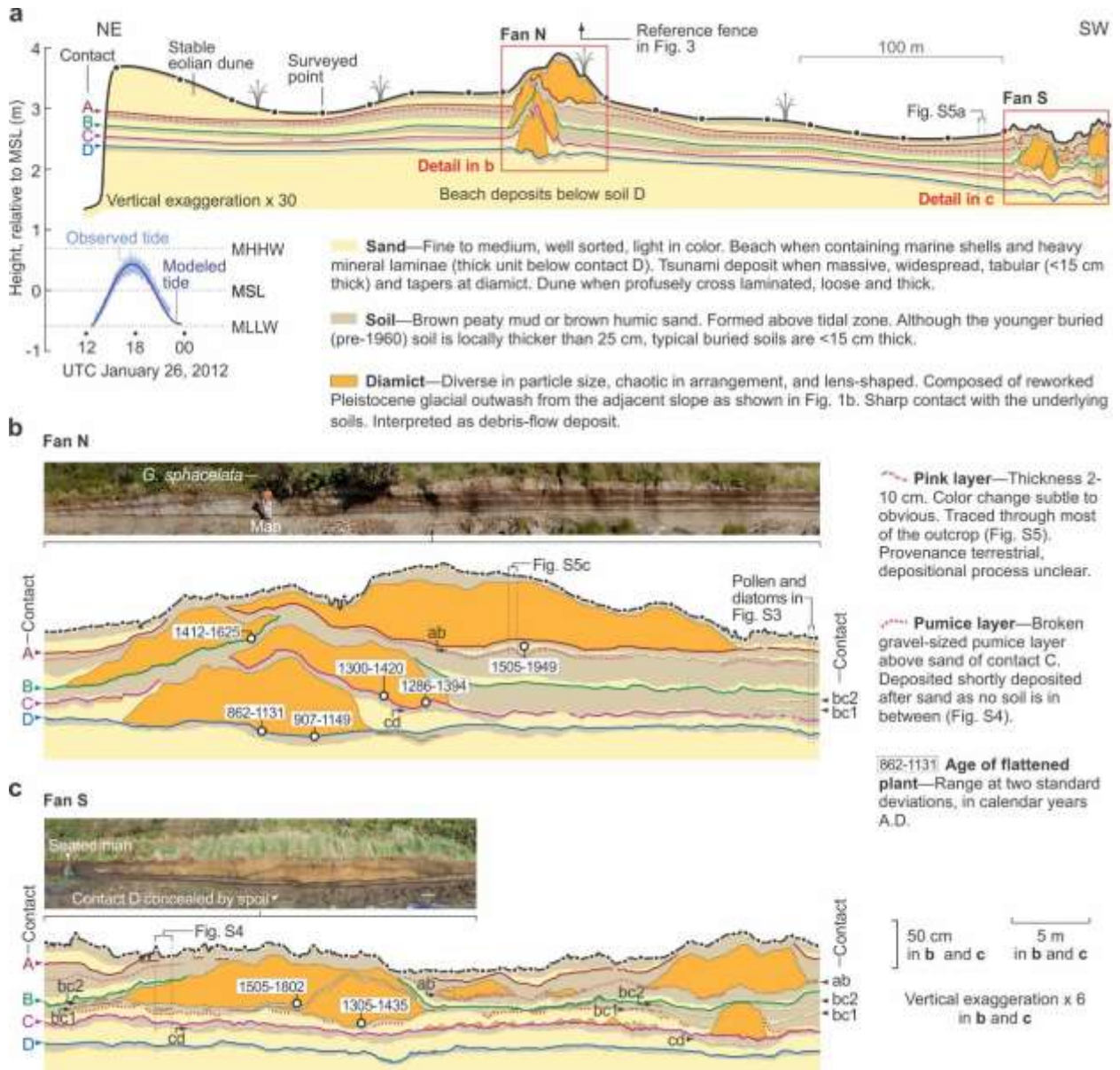
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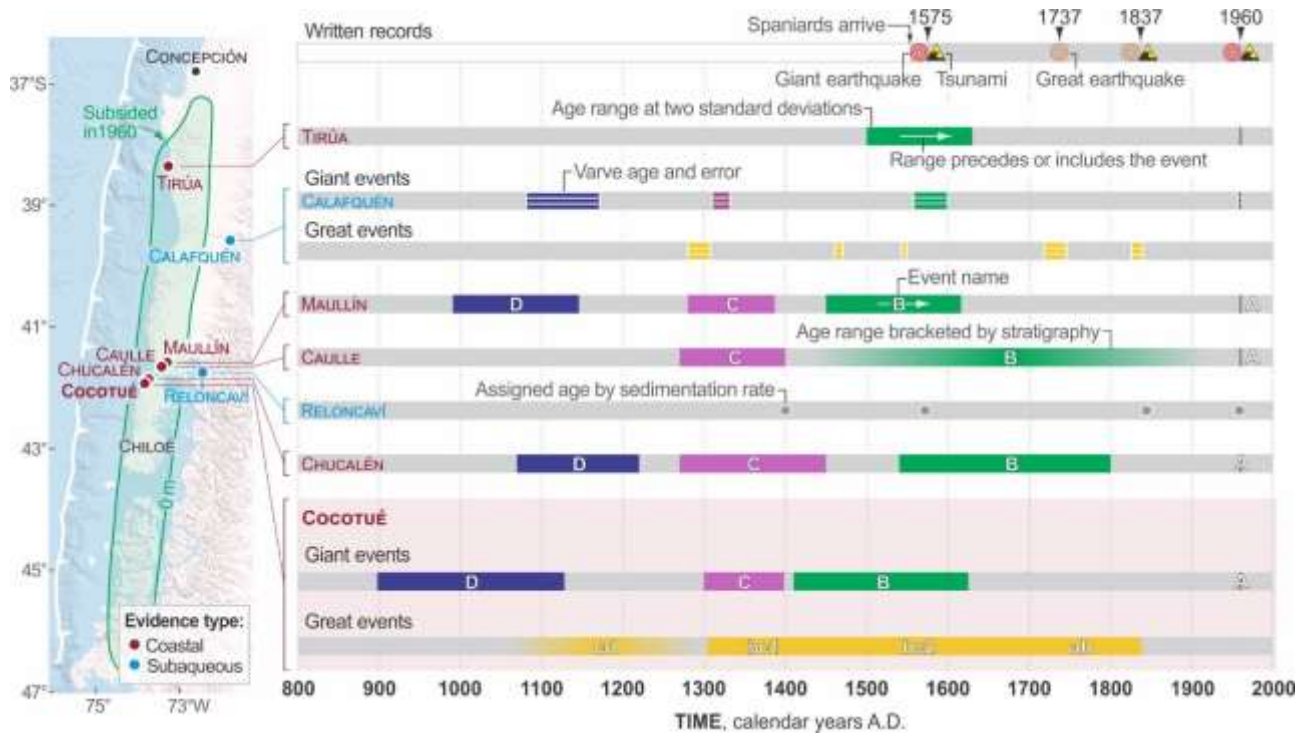
949 Figure 1

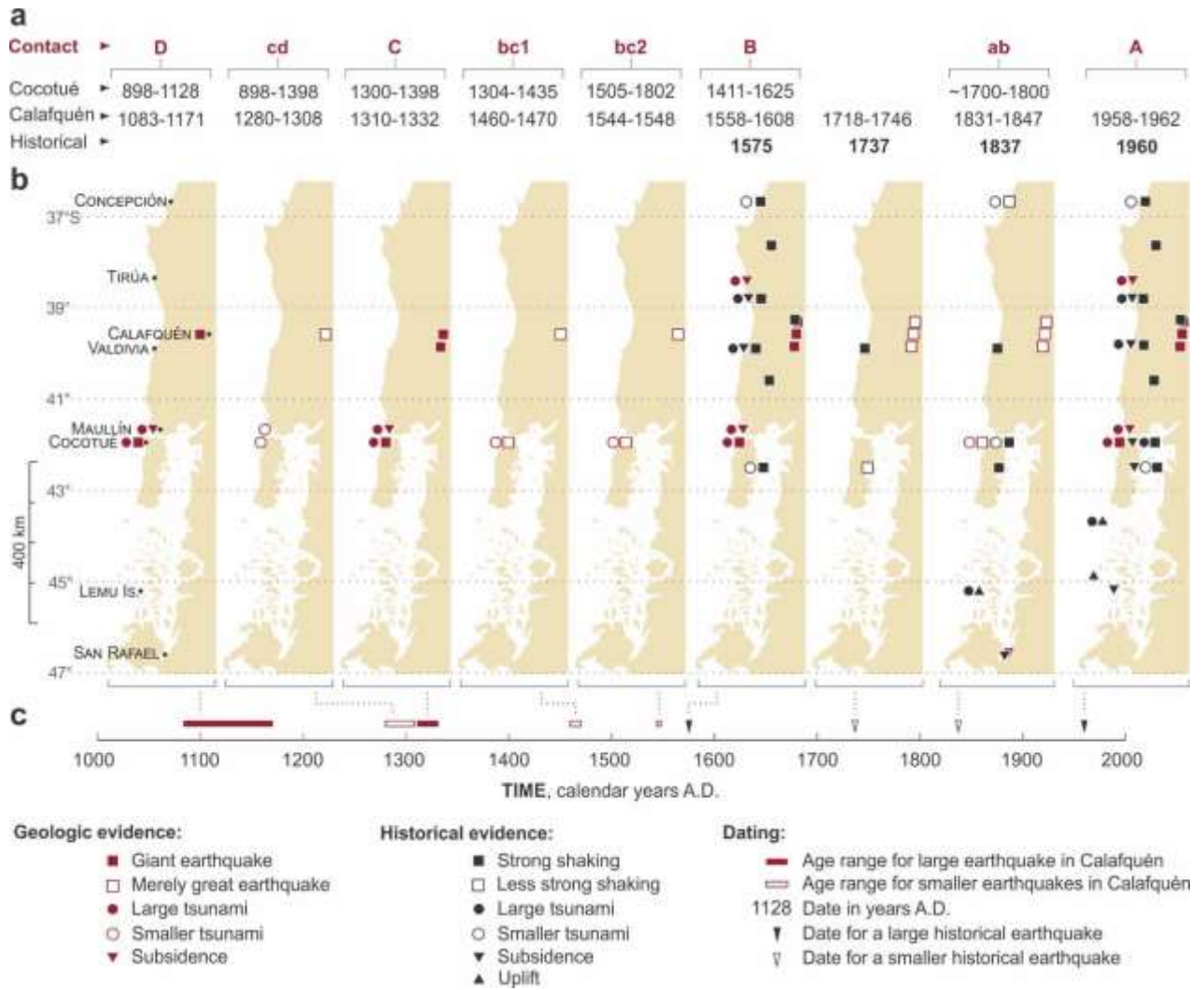












COASTAL EVIDENCE					LACUSTRINE EVIDENCE			
Cocotué	Dates in	Dates in	Dates in	Dates in	Dates in	Dates in	Dates in	Historical
event	Cocotué	Maulín	Caulle	Chucalén	Villarica	Calafquén	Riñihue	dates
	(AD at 2 σ)	(AD at 2 σ)	(AD at 2 σ)	(AD at 2 σ)	(AD)	(AD)	(AD)	(AD)

D	898-1128	1020-1180		1070-1220		1083-1171		
cd	between 898-1128 and 1300-1398	between 1020-1180 and 1280-1387				1280-1308		
C	1300-1398	1280-1387	1270-1400	1270-1450		1310-1332	1307-1327	
bc1	1305-1435					1460-1470	1464-1471	
bc2	1505-1802					1544-1548		
B	1412-1625	1450-1616	1575*	1540-1800	1558-1608	1558-1600	1562-1596	1575
					1723-1755	1718-1746		1737
ab	1505-1949				1820-1840	1831-1847	1826-1840	1837
A	1960*	1960*	1960*	1955-1971	1957-1963	1958-1962	1959-1963	1960