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Processes controlling a volcanoclastic turbiditic system during the last climatic cycle: Example of the Cilaos deep-sea fan, offshore La Réunion Island

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Abstract:

The present study focused on turbidite sedimentation in the Cilaos turbidite system, a volcanoclastic deepsea fan recently recognized offshore La Réunion Island. A set of piston cores was collected in order to establish the stratigraphy of this fan and to examine the processes controlling the turbidite sedimentation off the Cilaos cirque (Piton des Neiges volcanic massif) over the last climatic cycle. Two main phases of turbidite activity were identified, during the ca 140–127 ka and 30–0 ka periods, coinciding with the two last glacial–interglacial transitions (i.e., Terminations II and I). In addition to changes in climate and eustatic sea-level, these periods coincide with a low effusive volcanic activity of the Piton des Neiges volcano. The high erosional rates identified in the Cilaos cirque during these intervals of both low effusive volcanic activity and enhanced rainfall level are probably the main driver of sediment supply to the deep-sea depositional system. These new findings also highlight the important capacity of volcanoclastic turbidite systems to record rapid paleoenvironmental changes.

Keywords : Turbidites ; La Réunion Island ; Indian Ocean ; Late Quaternary ; Volcanoclastic system

1. Introduction

Recent studies based on high-resolution stratigraphy show that past climate and sea-level changes have had a significant influence on deep ocean turbidite sedimentation at both orbital and millennial timescales. These relationship has been demonstrated over the last few years along both siliciclastic ([Covault et al. 2007], [Jorry et al., 2008], [Toucanne et al., 2008], [Ducassou et al., 2010], [Jorry et al., 2011] and [Toucanne et al., 2012] among others) and carbonate margins (e.g., [Droxler and Schlager, 1985], [Schlager et al., 1994], [Andresen et al., 2003] and [Jorry et al., 2010]). Surprisingly, few studies have yet addressed the timing of gravity deposits around volcanic islands, and the forcing factors controlling the sediment delivery in such situations remain unclear ([Alibés et al., 1999] and [Frenz et al., 2009]).

52 The conventional sequence stratigraphy model for clastic systems states that deep marine
53 systems preferably grow during falls in sea-level and at lowstand. However, several studies
54 have demonstrated that some turbidite systems do not follow the classic sequence stratigraphy
55 concepts. Covault and Graham (2010) showed that deep-sea deposition occurs at all sea-level
56 states. Terrigenous sediment delivery to the deep-sea depends on many factors, such as the
57 tectono-morphologic character of the margin, climatic forcing and terrestrial sediment source.
58 The influence of climate and sea-level changes on sediment delivery to volcanoclastic basins
59 is poorly defined and remains a matter of debate. Quidelleur et al. (2008) and McMurtry et al.
60 (2004) suggested that most large volume landslides affecting volcanic islands occur at glacial-
61 interglacial transitions (Terminations) and concluded there was a causal relationship between
62 flank collapses of volcanic islands and global climate change. However, recent data contradict
63 these results, as these showed no link between climate-driven changes and volcanic flank
64 collapses (Harris et al., 2011; Longpré et al., 2011; Rodriguez-Gonzales et al., 2009). In
65 contrast, the influence of volcanic activity has been widely examined, especially off the
66 Canary Islands. In this area, the main turbidite activity has coincided with phases of high
67 volcanic activity (Schmincke and Sumita, 1998; Schneider et al., 1998).

68 Since 2006, several oceanographic cruises have been conducted on the submarine flanks of La
69 Réunion Island (Indian Ocean). These cruises led to the discovery of five volcanoclastic deep-
70 sea fans linked to major erosional structures visible on land (Saint-Ange et al., 2011; Sisavath
71 et al., 2011). La Réunion Island offers the opportunity to study a deep depositional system
72 related to an isolated oceanic island, situated far from continental influences. The aim of this
73 paper is to establish the first stratigraphy of the Cilaos turbidite system based on a set of
74 Küllenberg piston cores. We discuss how volcanic activity, climate and sea-level variations
75 have interacted and controlled the input of sediment offshore of La Réunion island over the

76 last 140 ka, leading to the building of a deep-sea fan spreading over hundreds of kilometers
77 on the sea floor.

78

79 **2. Physical setting**

80

81 *2.1. General setting of La Réunion Island*

82

83 La Réunion Island is an isolated volcanic system located 750 km from Madagascar in the
84 western part of the Indian Ocean (Fig. 1). It belongs to the Mascarene Archipelago and
85 resulted from the activity of the hotspot that formed the Deccan Trapps (65 Ma ago) and
86 subsequently the Mascarene Plateau and Mauritius Island (Bonneville et al., 1988; Duncan et
87 al., 1989; Morgan, 1981). It is the youngest and largest island in this group and the only one
88 that has active volcanism today. The island is located on an isolated compartment of the
89 oceanic crust (~67 Ma) bordered by two fracture zones (FZ) separated by 350 km: the
90 Mahanoro FZ to the west and Mauritius FZ to the east (Fig. 1).

91 La Réunion Island s.s. is the emerged part of a volcanic edifice rising from approximately -
92 4200 m (the oceanic floor) to more than 3000 m above sea level. The subaerial island
93 represents only three percent of the whole edifice (De Voogd et al., 1999). The submarine
94 flanks of La Réunion Island are mostly built by accumulation of debris avalanche deposits
95 (Lénat and Labazuy, 1990; Oehler et al., 2004). In the southwestern part of the island, no
96 major failures or flank collapses have produced deposits for 1 Ma (Oehler et al., 2008).

97 The island is elliptical in shape (50 × 70 km) and composed of two basaltic shield-volcanoes:
98 Piton des Neiges and Piton de la Fournaise (Fig. 1). Activity of Piton de la Fournaise (2632
99 m high) started less than 0.6 Ma ago and this volcano is still highly active (Gillot and Nativel,
100 1989). Its morphology is marked by a succession of calderas open to the sea on their eastern

101 sides (Fig. 1). The historic volcanic activity of Piton de la Fournaise has been described by
102 Bachelery et al. (1983), Lenat et al. (2009), Michon and Saint-Ange (2008), Peltier et al.
103 (2009; 2008) and Stieltjes et al. (1988).

104 The Piton des Neiges volcano (3070 m high) occupies the northwestern two thirds of the
105 island. The principal and most original feature of this volcano is the existence of three major
106 erosional depressions, called “cirques”, opened in the heart of the volcano (Fig. 1). The
107 cirques were partly filled by unconsolidated detrital rocks such as volcanic debris, avalanche
108 deposits, debris flow deposits and other breccia (Arnaud, 2005; Bret et al., 2003; Fèvre, 2005;
109 Oehler et al., 2005). Piton des Neiges started to grow more than 2.17 Ma ago, and has been
110 inactive for at least 0.012 Ma (Deniel et al., 1992; McDougall, 1971; Quidelleur et al., 2010;
111 Smietana et al., 2010). Its subaerial shield-building stage, known as the Oceanite Series,
112 extended from 2 Ma to approximately 430 ka (Deniel et al., 1992; McDougall, 1971; Upton
113 and Wadsworth, 1965). The second stage, known as the Differentiated Series, occurred
114 between 350 and 12 ka (Deniel et al., 1992; Gillot and Nativel, 1982; McDougall, 1971), with
115 the end of the main effusive activity at about 30 ka (Gillot and Nativel, 1982). The second
116 stage can be divided into three sub-stages. The first, between 350 and 180 ka, corresponded to
117 the beginning of Piton des Neiges magmatic differentiation. This effusive activity produced
118 lava flows that covered most of the edifice, filling existing valleys (Gillot and Nativel, 1982).
119 The second sub-stage corresponded to a low effusive activity period lasting 40 ka (Kluska,
120 1997), with mainly explosive activity (Kieffer, 1990). Kluska (1997) suggested that this was a
121 major erosional period corresponding to the formation of large and deep depressions: the
122 cirques. A second period of effusive activity took place between 140 and 30 ka.

123

124 *2.2. Local climate*

125

126 La Réunion Island is located in the subtropical zone, where it is influenced by the South
127 Equatorial Current and subjected to a prevailing southeasterly trade-wind regime. Trade
128 winds from the east induce highly variable precipitation regimes in time and space, which
129 lead to the island having a wet windward side (East) and a dry leeward side (West). Rainfall
130 also varies according to the elevation (dry coast - wetter upland), with maximum rain in the
131 mid-slope area. Average annual rainfall varies from over 12 000 mm per year between 1300
132 and 2000 m altitude on windward slopes, to as low as 600 mm near the leeward coast.
133 The late Quaternary climate of La Réunion Island is largely unknown, as no data are available
134 for this area.

135

136 *2.3. Drainage basins and rivers*

137

138 On La Réunion Island, the wet tropical climate and basaltic terrains cause high erosion rates,
139 which are amplified by seasonal cyclonic conditions (Louvat and Allegre, 1997; Rad et al.,
140 2007). These erosion rates are close to those estimated in active orogenic areas, with values
141 ranging between $0.47 - 3.4 \text{ m.kyr}^{-1}$. They result in a dense hydrographic network with more
142 than 750 gullies and rivers on the island, concentrated in drainage basins like the cirques and
143 the main valleys. These drainage basins are located between the planeze areas (Ollier and
144 Terry, 1999), which are relatively unaffected by erosion because water penetrates rather than
145 forming surface runoff (Fèvre, 2005). In the drainage basins, the rivers are torrential with
146 mechanical erosion rates among the highest measured in the world, ranging from 1200 to
147 $9100 \text{ t/km}^2/\text{yr}$ (Louvat and Allegre, 1997).

148 One of the major rivers of the island is the “Rivière Saint-Etienne”. It is a torrential river
149 formed by the junction of the “Bras de Cilaos”, which drains the inner part of the Cirque of
150 Cilaos, and the “Bras de la Plaine”, which drains the outer eastern part (Fig. 1). The Saint-

151 Etienne River has a drainage basin of about 200 km² reaching altitudes of 2500 to 3000 m. Its
152 mean fluvial solid load is estimated around 470 000 m³/yr; up to 1-2 million m³/yr during
153 large floods (SOGREAH, 1998). In addition to this drainage basin, the outer western part of
154 the Cirque of Cilaos is incised by many gullies (Fig. 1). The resulting drainage basin for the
155 Cirque of Cilaos (“Rivière Saint-Etienne” and gullies) is about 360 km².

156 The transition between the subaerial and submarine environments is marked by a narrow
157 shelf. The local absence of this shelf and the presence of steep submarine slopes around the
158 island imply a rapid transfer of sediment from the coast toward the base of the volcanic
159 edifice, allowing the formation of a volcanoclastic deep sea fan.

160

161 *2.4. The Cilaos turbidite system*

162

163 Recent oceanographic cruises over the submarine flanks of La Réunion Island and the
164 surrounding oceanic plate led to the discovery of five volcanoclastic turbidite systems (Fig. 1)
165 extending over 200 km away from the island (Saint-Ange et al., 2011; Sisavath et al., 2011).
166 On land, these systems are related to major erosional features that constitute the main
167 drainage area of the island. In each case, submarine canyons are directly connected to the
168 main river mouths. The Mafate fan is connected to the Cirque of Mafate, and the Saint-Joseph
169 fan is the only system connected to the Piton de la Fournaise volcano. The latter is considered
170 an embryonic stage fan. Finally, the Cilaos fan is the widest fan and is connected to the
171 Cirque of Cilaos.

172 The Cilaos turbidite system is located to the southwest of La Réunion Island. It is more than
173 250 km long and covers an area of about 15 000 km². This sedimentary system extends from
174 the Saint-Etienne river mouth to the Mahanoro fracture zone (Fig. 1). It starts at the coast,
175 with a 70 km long bypass area that directly feeds a deep-sea fan developing at about 4500 m

176 water depth. The Cilaos fan extends over a complex abyssal plain composed of NE-SW
177 volcanic ridges (Saint-Ange et al., 2011).

178 The canyon area (Fig. 1; outlined in blue) is composed of two main canyons, Saint-Etienne
179 and Pierrefonds. The Pierrefonds canyon is located in front of the older Saint-Etienne River
180 delta and is connected to the shelf by many tributary canyons. The Saint-Etienne canyon is 4
181 km wide and is directly connected to the present day Saint-Etienne River. Both canyons
182 merge into the single wide Cilaos canyon, which feeds the main deep sea fan body.

183 The fan can be divided into two parts: the proximal fan (Fig. 1; outlined in red) and distal fan
184 (Fig. 1; outlined in yellow). The proximal fan is broad, with a maximum width of 120 km and
185 low reflectivity of backscatter data. It is composed of elongated bodies, interpreted as small
186 lobes. The sedimentation in the proximal fan is characterized by coarse sandy turbidites
187 (Sisavath et al., 2011). The distal fan is divided into three parts, western, central and eastern,
188 by pre-existing volcanic ridges. The distal prolongation of the turbidite system is visible at the
189 ends of the western and central parts. It is characterized by elongated structures, extending via
190 narrow channels from the proximal fan. The sediments of the distal fan are composed of a
191 succession of fine sandy turbidites covered by a thick clay layer (about 3 m in thickness; Figs.
192 2 and 3).

193

194 **3. Materials and Methods**

195

196 In this paper, we used seven Küllenberg piston cores taken around La Réunion Island during
197 the oceanographic cruises ERODER 1, onboard the BHO *Beautemps-Beaupré* in 2006;
198 FOREVER, onboard the R/V *Atalante* in 2006; and ERODER2, onboard the R/V *Meteor* in
199 January 2008 (Fig. 1, Table 1). Five cores were taken from locations in the Cilaos fan
200 (KERO-09, KERO-16, KERO-12, KERO-15 and FOR-C1). Additional cores from the Mafate

201 fan (KERO-07, Fig. 1) and the Saint-Joseph fan (KERO-08, Fig. 1) were used to build a
202 regional age model. All the cores were situated and correlated using Parasound and 3.5 kHz
203 echosounder profiles acquired during the FOREVER and ERODER2 cruises (Fig. 3).
204 Sedimentary descriptions were made of all the cores, with a particular emphasis on sediment
205 color, visual grain size and turbidite/hemipelagite/pelagite differentiation. Two main types of
206 sediment were distinguished: volcanoclastic sandy turbidites and hemipelagic sediments.

207 A series of 1-cm-thick sediment slabs were collected from each split core section and
208 examined by X-radiography using a SCOPIX digital X-ray imaging system (Migeon et al.,
209 1999). Digital images were acquired to provide a precise identification of the sedimentary
210 structures. Sediment cores were sampled for grain-size analyses using a Coulter laser micro-
211 granulometer (LS130). The variation of Ca through each of the cores was measured with an
212 Avaatech XRF Core-Scanner equipped with a variable optical system allowing measurements
213 at resolutions between 10 and 0.1 mm. The selected measurement area was 8 mm and the
214 step-size was set at 1 cm.

215 Oxygen isotope analyses were conducted on small batches of *Globigerinoides ruber*, the
216 monospecific planktonic foraminifer that calcifies in the surface mixed layer, from cores
217 KERO-07, KERO-08, KERO-09 and KERO-16. Samples were collected at hemipelagic
218 intervals, representing intervals of continuous sedimentation, excluding turbidites. Cores were
219 sub-sampled with a sample spacing of 5 to 20 cm. On average, 15 specimens were picked out
220 from the >150 μm fraction. Using a common 100% phosphoric acid bath at 90°C, 20–50 μg
221 of sample were reacted and analyzed using a GV Isoprime isotope ratio mass spectrometer at
222 University of Pierre & Marie Curie (Paris). Isotope values are given in delta notation relative
223 to Vienna Peedee belemnite. Repeated analyses of a marble working standard (calibrated
224 against the international standard NBS-19) indicate an accuracy and precision of 0.1‰ (1 σ).

225 In core KERO-16, the last occurrence of pink-pigmented *G. ruber* indicates the transition
226 between Marine Isotopic Stage (MIS) 6 and MIS 5 (Thompson et al., 1979).

227 Nine AMS radiocarbon dates were obtained on the cores (Table 2). For each sample, about 10
228 mg of *G. ruber* and *G. sacculifer* specimens were picked out from the >150 mm fraction,
229 washed in an ultrasonic bath with distilled water, and dried. These samples were then
230 analyzed at the Poznan Radiocarbon Lab., Poland, and at the “Laboratoire de Mesure du
231 Carbone 14” at Saclay, France. Reported radiocarbon ages were corrected for a marine
232 reservoir effect of 400 years and converted to calendar years using CALIB Rev 6.0 (Reimer et
233 al., 2009). Calibrated kilo years before the present will be referred as ka.

234 The preservation of the test surface of the foraminifer *G. ruber* was examined by Scanning
235 Electron Microscopy (SEM, Philips XL30). The analysis was performed on *G. ruber* from
236 seven samples studied for $\delta^{18}\text{O}$ measurements (three in core KERO-09 and four in core
237 KERO-16). Foraminifera were placed on adhesive carbon tabs and coated with gold. The
238 observation of the test surfaces was done in the secondary electron mode at 10 kV voltage and
239 at a distance of 10 mm.

240 In core KERO-09, five representative samples associated with particular sedimentary facies
241 were also analyzed for calcareous nannofossil biostratigraphy (Table 3). Smear slides were
242 made directly from unprocessed samples and were examined with a polarized light
243 microscope at a magnification of 1000 \times .

244

245 **4. Results**

246

247 *4.1. Lithology and echosounding facies*

248

249 Based on the grain-size characteristics, internal sedimentary structures, erosive contacts with
250 underlying sediments and the abundance of glass shards and volcanic crystals, all the sandy
251 beds in the studied cores were interpreted as volcanoclastic turbidites (Saint-Ange et al., 2011;
252 Sisavath et al., 2011). These turbidite units ranged from a few centimeters up to 20 cm in
253 thickness (Fig. 2).

254 Cores KERO-09 and KERO-12, taken in the western part of the Cilaos distal fan at
255 about 215 km from the island, were 6.27 and 6.40 m long, respectively (Fig. 1). The lower
256 parts of these cores are characterized by a succession of four sandy units of 30 to 50 cm thick,
257 composed of well-sorted fine sand showing typical Bouma Tb to Te sequences (Bouma,
258 1962) (Fig. 3). On the echosounder profiles, this lowest unit corresponds to a stratified unit
259 named U1 (Fig. 3). A thick layer of clay (about 3 m thick), showing an alternation of light
260 brown clay and darker brown clay, overlies this unit. The light brown clay is dominated by
261 calcareous sediment (composed of nannoplankton and foraminifera), while the dark brown
262 clay mainly contains siliceous organisms (radiolarians and diatoms). On the echosounder
263 profiles, it corresponds to a semi-transparent unit, named U2 (Fig. 3).

264 Core KERO-15 was taken in the most distal part of the Cilaos distal fan at about 280
265 km from the island (Fig. 1). This 6.68 m-long core shows a sedimentary pattern similar to that
266 observed in cores KERO-09 and KERO-12. The base of the core shows a succession of seven
267 fine-sandy layers (20 to 80 cm thick) with typical Bouma Tb to Te sequences, and belongs to
268 unit U1 on the echosounder profiles (Fig. 3). The upper part of the core is composed of a clay
269 layer of about 2 m thick showing an alternation of light brown clay and darker brown clay,
270 corresponding to the unit U2 on the echosounder profiles (Fig. 3).

271 Core KERO-16 (4.95 m) was taken at a water depth of 4340 m, in the central part of
272 the Cilaos distal fan, on the right side of a channel. Clay layers (alternation of light brown
273 clay and darker brown clay) dominate the lithological succession in the lower part of the core,

274 which locally includes small bioturbation features. This unit corresponds to unit U2 on the
275 echosounder profiles. At the top of the core (0-1.4 meters below seafloor, mbsf), fine-grained
276 turbidite deposits of few centimeters thickness are visible (Fig. 2). They were interpreted as
277 overflow deposits. On the echosounder profiles, they correspond to the upper part of the
278 profiles, characterized by a stratified unit named U3.

279 Core FOR-C1 was taken at the top of a sedimentary ridge in the central part of the
280 distal fan at a water depth of 4074 m (Fig. 1). The 2.50 meters forming the base of the core
281 are composed of an alternation of clay and thin sandy turbidites (less than 10 cm thick). This
282 unit corresponds to the unit U1 on the echosounder profiles (Fig. 3). It is overlain by 1.50 m
283 of clay (alternation of light brown clay and darker brown clay) corresponding to the semi-
284 transparent unit U2 on the echosounder profiles (Fig. 3). The top 0.54 cm are composed of
285 clay interbedded with thin sandy layers (1 cm thick), interpreted as overflow deposits and
286 corresponding to the stratified unit U3 on the echosounder profiles (Fig. 3).

287 Core KERO-07 was taken in the Mafate fan at a water depth of 791 m. It is a 3.40 m-
288 long core from the left side of the canyon area (Fig. 1). The lower two thirds of the core
289 correspond to a succession of sandy and silty turbidite sequences that are variable in
290 thickness, while the upper third is composed of silty-clay (Fig. 2).

291 Core KERO-08 was taken near the volcanic ridge R4 at a water depth of 4126 m, in
292 the distal part of the Saint-Joseph fan. Its lithological succession is composed of a succession
293 of sandy and silty turbidites whose thicknesses range from 2 to 15 cm. These well-sorted
294 turbidites show normal grading and horizontal laminations. In the dark sandy layers,
295 laminations are underlined by white laminae characterized by abundant foraminifera and
296 bioclasts. Some clay and silty clay layers are highly bioturbated.

297

298 *4.2. Chronostratigraphic framework*

299

300 The chronostratigraphic framework of cores KERO-09, KERO-16, KERO-07 and KERO-08
301 was established through integration of radiocarbon dating, planktonic oxygen isotope,
302 biostratigraphic markers and XRF records.

303 The $\delta^{18}\text{O}$ curves show similar trends between all cores (Figs 5 and 6). The upper parts of the
304 cores show light $\delta^{18}\text{O}$ values (down to -2.0 ‰ in core KERO-07). A rapid increase of the $\delta^{18}\text{O}$
305 signal is observed thereafter, and heavier $\delta^{18}\text{O}$ values (from -1.0 to 0 ‰) characterized the
306 lower parts of the cores (Fig. 4). Peaks and troughs recognized in the oxygen isotope records
307 were correlated with the reference isotopic signal published by Fretzdorff et al. (2000) from
308 La Réunion area, in agreement with the trends of the $\delta^{18}\text{O}$ benthic stack record of Lisieky and
309 Raymo (2005) (Figs. 4 and 5).

310 The light $\delta^{18}\text{O}$ values observed in the upper part of the cores correspond to the Holocene (<11
311 ka). MIS2 is characterized by values of $\delta^{18}\text{O}$ between 0 and -0.5 ‰. A general decrease of the
312 $\delta^{18}\text{O}$ values from 0 to -1 ‰ is clearly shown in MIS3 (Fig. 5). It is followed by an increase of
313 $\delta^{18}\text{O}$ values identified as the MIS3/MIS4 transition. The relatively light $\delta^{18}\text{O}$ values (>-1‰)
314 observed in the lower part of cores from the Cilaos fan correspond to the last interglacial. The
315 oxygen isotope stratigraphy of all these cores provides a regional record of the last climatic
316 cycle around La Réunion Island.

317 In cores of the Cilaos fan (KERO-09 and KERO-16), MIS-5 is characterized by $\delta^{18}\text{O}$ values
318 ranging between -1 ‰ and 0 ‰, which are unusually low for the last interglacial compared
319 with those published by Fretzdorff et al. (2000) (Fig. 5). SEM observations of the test surface
320 of *G. ruber* in cores KERO-09 and KERO-16 reveal some dissolution pockets and
321 recrystallized areas (Fig. 6), which could explain these inconsistent $\delta^{18}\text{O}$ values. In addition,
322 periods of high carbonate dissolution have been identified in the western part of the Indian

323 ocean, mainly during interglacials (Divakar et al., 1993). However, this chronostratigraphy
324 was supported by the study of biostratigraphic markers. The nannofossil assemblage in core
325 KERO-09 contains abundant *E. huxleyi* at 65 cm below seafloor (bsf), suggesting an age
326 younger than 75-90 ka (Berggren et al., 1995). Samples from 115-117 and 361 cm bsf show
327 abundant well preserved nannofossils. The occurrence of *E. huxleyi* and the absence of *P.*
328 *lacunosa* suggest that these samples are younger than 260 ka (Berggren et al., 1995). The
329 sample from 608 cm bsf has abundant but a poorly preserved nannofossils. *Gephyrocapsa spp*
330 *cf caribbeanica* is dominant and two *P. lacunosa* are present, suggesting an age younger than
331 460 ka (Berggren et al., 1995). The last occurrence of pink pigmented *G. ruber* is also
332 observed in core KERO-16 at 4.70 m bsf, suggesting that the upper 4.70 m of KERO-16 is
333 aged 120 ka (Thompson et al., 1979).

334 All these data provide a consistent age model around La Réunion Island. This age model was
335 extended to other cores of the Cilaos fan (KERO-12, KERO-15 and FOR-C1) using the XRF
336 records. The age model of cores FOR-C1, KERO-12 and KERO-15 was established by
337 correlating the Ca variation of cores KERO-09 and KERO-16 (Fig. 7). In all the cores, light
338 brown clays, corresponding to high XRF Ca values, allow a reliable core-to-core correlation.

339

340 4.3. Lithostratigraphy

341

342 The lithostratigraphy in the Cilaos fan was established with cores KERO-09, KERO-12,
343 KERO-15, KERO-16 and FOR-C1. Based on the age model, the five cores retrieved from the
344 Cilaos fan extend from 10 ka to 130 ka, with the Holocene period only being recorded for
345 KERO-16 (Fig. 4). Although distances of tens of kilometers separate them, the variation of
346 the calcium XRF correlates well between the five cores through the last glacial-interglacial
347 cycle (Fig. 7). The cores exhibit a common sediment pattern and a fairly similar

348 sedimentation rate. They are all composed of a succession of turbidites covered by a thick
349 clay layer (Fig. 2). The sedimentation rate in the hemipelagic layer ranges between 1.8 and
350 5.2 cm/ka (Fig. 8). These results are comparable to the minimum sedimentation rate of 1.9
351 cm/ka observed by Ollier et al. (1998), based on micropaleontological analysis. They also
352 correlate with the sedimentation rates measured by Fretzdorff et al. (2000) in core S 17-666,
353 near the Mafate fan (Fig. 1), based on a $\delta^{18}\text{O}$ stratigraphy (Fig. 4). In this core (S17-666),
354 three stages can be observed in the sedimentation: 1) between 128 and 186 ka, with a
355 sedimentation rate of 4.14 cm/ka; 2) between 26 and 128 ka, with an average rate of
356 sedimentation of 2.27 cm/ka; and 3) between 14.5 and 26 ka, with a sedimentation rate of
357 4.35 cm/ka. If we calculate the mean of the sedimentation rates in our cores for each period,
358 we obtain sedimentation rates of about 4.9 cm/ka for the first stage, 2.73 cm/ka for the second
359 stage and 3.02 cm/ka for the third. The sedimentation rates lie between those of the Mafate
360 and Cilaos systems, with two stages of relatively high sedimentation rate interrupted by a
361 period of low sedimentation.

362 These three distinct phases of sedimentation correlate with the three sedimentary units
363 identified in the cores of the Cilaos turbidite system. The first unit corresponds to the turbidite
364 activity visible in the lower part of cores KERO-09, KERO-12 KERO-15 and FOR-C1,
365 characterized by sandy turbidites of 30 to 50 cm thickness (Fig. 3) and corresponding to the
366 stratified unit U1 on the echosounder profiles (Fig. 3). This first stage is characterized by
367 turbidity currents that spread over the entire fan (Sisavath et al., 2011). This first documented
368 phase of turbidite activity would have deposited until the end of MIS5 (Fig. 8). Therefore,
369 according to the age model of figure 8, the top of unit U1 – marking the interruption of this
370 first phase of turbidite activity – corresponds to an age of 125-127 ka.

371 The second phase is characterized by a thick clay layer observed in all the cores (Fig. 2) and
372 by the absence of major turbidite activity. Only a few thin sandy layers are observed in cores

373 FOR-C1 and KERO-16 (Fig. 8). On the echosounder profiles, this unit coincides with the
374 semi transparent unit U2, visible over the entire fan except in the more proximal part
375 (Sisavath et al., 2011). The timing of clay deposition ranged from MIS3 to MIS5 (Fig. 8). The
376 top of unit U2, visible in the upper part of cores KERO-16 and FOR-C1, lies within the lower
377 part of MIS2 at about 30 ka (Fig. 8).

378 The third phase is characterized by thin sandy layers observed in cores FOR-C1 and KERO-
379 16 (Fig. 3), with sediments coarser than in the older phase of turbidite activity. These turbidite
380 events were observed on the proximal fan and into the channels of the distal fan (Sisavath et
381 al., 2011). They are characterized, on the echosounder profiles, by a stratified unit (U3)
382 visible in the upper part of the profiles (Fig. 3). This phase corresponds to the most recent
383 activity of the Cilaos fan and was deposited during MIS1 and MIS2 until 30 ka.

384

385 **5. Discussion**

386

387 This discussion is based on cores KERO-09, KERO-12, KERO-15, KERO-16 and FOR-C1.
388 Cores KERO-07 and KERO-08 were used to build a consistent regional $\delta^{18}\text{O}$ stratigraphy
389 around La Réunion Island.

390

391 5.1. Sedimentation in the Cilaos fan over the last 140 ka

392

393 Three distinct episodes of sedimentation correlate with three sedimentary units identified in
394 the cores of the Cilaos turbidite system. The first unit corresponds to the oldest turbidite
395 activity, visible in the lower part of cores KERO-09, KERO-12 KERO-15 and FOR-C1. This
396 first unit, spreading over the entire fan (Sisavath et al., 2011), is characterized by sandy
397 turbidites of 30 to 50 cm thick (Fig. 3) and corresponds to the stratified unit U1 on the

398 echosounder profiles (Fig. 3). According to our age model (Fig. 8), the first period of turbidite
399 activity began before 140 ka and ended at ~ 125-127 ka (which coincides with the MIS5
400 highstand).

401 The second phase is characterized by a thick clay layer observed in all the cores (Fig. 2)
402 which illustrates the absence of major turbidite activity. Only a few thin sandy layers are
403 observed in cores FOR-C1 and KERO-16 (Fig. 8). On the echosounder profiles, this unit
404 coincides with the semi transparent unit U2, visible over the entire fan except in the more
405 proximal part (Sisavath et al., 2011). The timing of clay deposition ranged from MIS3 to
406 MIS5 (Fig. 8). The top of unit U2, visible in the upper part of cores KERO-16 and FOR-C1,
407 lies within the lower part of MIS2 at about 30 ka (Fig. 8).

408 The third phase is marked by the deposition of thin sandy layers as observed in cores FOR-C1
409 and KERO-16 (Fig. 3), with sediments coarser than those deposited during the older U1 unit.
410 These turbidite events were active in the proximal fan and into the channels of the distal fan
411 (Sisavath et al., 2011). They are characterized, on the echosounder profiles, by a stratified unit
412 (U3) visible in the upper part of the profiles (Fig. 3). This phase corresponds to the most
413 recent activity of the Cilaos turbiditic fan since 30 ka until Holocene.

414 A significant difference in sedimentation rate is observed between core KERO-16 (located
415 near the main channel of the central part of the Cilaos fan, showing the highest sedimentation
416 rate) and the other cores KERO-09, KERO-15, KERO-12 and FOR-C1 (Fig. 9). The location
417 of cores KERO-09, KERO-15 and KERO-12 at the termination of the distal fan might explain
418 a lower sediment supply compared to the upper/central part of the Cilaos fan. The location of
419 core FOR-C1 on top of a sedimentary ridge (about 200 m-high) explain sedimentation rates
420 significantly lower than those observed in KERO-16.

421 In the distal part of the Cilaos fan, sedimentation rate is rather homogeneous until ~ 60 ka and
422 decreases from ~ 60 ka to ~ 10 ka (Fig. 9), which corresponds to the progressive abandon of

423 the turbiditic sedimentation in the western distal part of the Cilaos fan since ~ 60 ka (Sisavath
424 et al., 2011). The increase of sedimentation rate from ~ 100 ka to ~ 45 ka in cores KERO-16
425 and FOR-C1 reflects a turbiditic activity restricted to the central part of the fan.
426 From ~ 45 ka to ~ 30 ka, decrease in sedimentation rates in all cores can be interpreted as a
427 major change in the sediment supply at the scale of the entire Cilaos deep-sea fan (Fig. 9).
428 This period coincides with the youngest phase of effusive volcanic activity of the Piton des
429 Neiges (Kluska, 1997, Salvany et al., 2012), which has probably contributed to fill the cirques
430 and the fluvial valleys. At ~ 30 ka, the major increase in sedimentation rates detected in core
431 KERO-16 corresponds to a new episode of turbidite activity (unit U3, Fig.9) restricted to the
432 central part of the fan (Sisavath et al., 2011).

433

434 5.2. Unravelling the forcing factors of turbidite sedimentation in the Cilaos fan over the last
435 140 ka

436

437 The three main forcing factors controlling sediment supply and transport offshore volcanic
438 islands are volcanic activity, climate and sea-level (Krastel et al., 2001, Quidelleur et al.,
439 2008). Recent works on the morphology and sedimentary architecture of the Cilaos fan show
440 that the sedimentary processes involved in the feeding of the Cilaos turbidite systems are
441 direct feeding by rivers and submarine slope instabilities (Saint-Ange et al., 2011; Sisavath et
442 al., 2011). The detailed stratigraphy of the Cilaos fan obtained in the present study allows us
443 to test this assumption over the last climatic cycle.

444

445 *5.2.1 Turbidites in relation to climate and sea-level fluctuations*

446 Offshore La Réunion Island, the two main intervals of turbidite activity coincide with the
447 transition from glacial lowstand to highstand condition. The first phase of turbidite deposits

448 coincided with lowstand and rising sea-level, at about 137 ka and between 137 and 130 ka,
449 respectively (Rohling et al., 2009; Fig. 8). The last recurrence of turbidite deposition in the
450 Cilaos system (unit U3) also coincided with such a sea-level pattern, the turbidite activity is
451 visible during the LGM lowstand (26 to 19.5 ka, Clark et al., 2009) and continuing until the
452 next highstand conditions. The intervening period did not show any turbiditic conditions.

453 A link between sea-level change and large scale landsliding is suggested by some authors (e.
454 g., McMurtry et al., 2004; Quidelleur et al., 2008). Sea-level variations can change the pore
455 pressure conditions, which are related to the location of the aquifer on top of a hydrothermal
456 unit (Join et al., 2005) , or influence the submarine and coastal boundary conditions that
457 control groundwater flow in the volcanic edifice. In his study of factors that could induce
458 large flank destabilization on shield volcanoes, Iverson (1995) had already concluded to the
459 minor role played by sea level changes. When the sea-level drops, the mechanical resistance
460 of the hydrothermal unit decreases as the pressure exerted by the water table at the base of
461 edifice increases. The mechanical resistance decreases until the rupture threshold is reached,
462 inducing a rapid lateral sliding of the volcano flank. This link between sea-level change and
463 large scale landsliding is not observed for the Cilaos fan because no flank collapses have
464 destabilized the studied area for 1 Ma (Oehler et al., 2008). Moreover, turbidite activity is
465 visible during lowstand and highstand conditions, suggesting that sea-level variation has little
466 influence on the development of turbidites in the Cilaos deep-sea fan.

467
468 The lack of palaeoclimatic records from La Réunion Island preclude the direct correlation of
469 the turbidite activity in the Cilaos Fan with climate changes. An alternative is to examine the
470 palaeoclimatic reconstructions from southern Africa. Intense debate persists about the
471 climatic mechanisms governing hydrologic changes in this area (e.g., Schefuß et al., 2012).
472 However, recent results suggest that mean summer insolation controls the atmospheric

473 convection, with higher insolation leading to higher rainfall (Schefuß et al., 2012). By
474 considering this orbital forcing over a geological timescale, this implies that glacial-
475 interglacial transitions in the southern African tropics were characterized by significant
476 changes in rainfall level, from wet to dry conditions. Such a pattern has been demonstrated for
477 the last Termination, through the runoff of the Zambezi river (Schefuß et al., 2012), and
478 corroborates previous rainfall reconstructions from South Africa and Madagascar over the last
479 150-200 ka (Partridge et al., 1997; Gasse and Van Campo, 2001).

480 Based on a geomorphological approach, Saint-Ange et al. (2011) showed that the Cilaos Fan
481 is directly connected to the Saint-Etienne river mouth. This implies that river runoff is a major
482 forcing factor on sediment delivery to the Cilaos Fan, and that a high rainfall period in the
483 southern African tropics would increase sediment supply to the deep-sea fan. A sediment
484 delivery process of this type to a deep-depositional system has been demonstrated from short,
485 mountainous river systems (e.g., Makran margin, Southern California; Bourget et al., 2010;
486 Covault et al., 2010). Based on the rainfall reconstructions of Partridge et al. (1997) (Fig. 8),
487 one would expect the turbidite sedimentation in the Cilaos Fan to increase during southern
488 hemisphere summer insolation maxima. However, no turbiditic activity was recorded in the
489 Cilaos Fan at the time of glacial rainfall maxima (e.g., ca 90, 70 or 50 ka; Partridge et al.,
490 1997). Moreover, the turbidite activity off the St Etienne river, centred on Termination II and
491 I, began at a time of wet (MIS 2) to very wet conditions (MIS 6) in the southern African
492 tropics and continued during both the climatic transition (i.e., Terminations) and the following
493 dry conditions (i.e., MIS 5 and MIS 1) (Partridge et al., 1997; Gasse and Van Campo, 2001).

494 To understand the impact of climate on the development of turbidite in the Cilaos fan, it
495 would be necessary to obtain more accurate data about the climatic variations in La Réunion
496 Island but, according to current knowledge, the results presented here call into question the
497 impact of rainfall level alone on sediment delivery to the Cilaos basin.

498

499 *5.2.2 Turbidites in relation to volcanic activity*

500

501 The two main periods of turbidite activity in the Cilaos system coincided with periods of low
502 effusive volcanic activity at La Réunion. Indeed, the first phase of turbidite activity (unit U1)
503 can be associated with the low effusive activity identified between 180 and 140 ka at Piton
504 des Neiges (Kluska, 1997; Salvany et al., 2012). The second period of turbidite activity (Unit
505 U3) began at the end of the effusive activity of the Piton des Neiges volcano, dated at about
506 30 ka by Gillot and Nativel (1982). These periods correlate with the major erosional episodes
507 interpreted by Kluska (1997) and Salvany et al. (2012) (Fig. 10). The decrease of volcanic
508 production coincided with the erosional formation of the cirques (Kluska, 1997; Salvany et
509 al., 2012). This strongly suggests that the low effusive activity contributed to increasing the
510 sediment input to the submarine flank and the deep Cilaos basin surrounding La Réunion
511 Island. Conversely, the interruption of turbidite activity (between 127 ka and 30 ka) coincided
512 with a resumption of the effusive and explosive activity of the Piton des Neiges volcano, with
513 large lava flows that filled the cirques and their drainage valleys (Fig. 10) (Kluska, 1997,
514 Salvany et al., 2012). In the cirque of Cilaos, the first lava flows related to this new volcanic
515 event are estimated to have appeared at 130 ka (Kluska, 1997). The filling of the drainage
516 basin by lava flows probably obstructed the pathways for sediment transfers to the deep
517 marine environment.

518 In the Canary Islands, for comparison, a clear relationship is observed between the timing of
519 emplacement of volcanic-rich turbidites and the period of explosive volcanic activity (Alibés
520 et al., 1999; Frenz et al., 2009). During volcanically active stages, an increase of sediment
521 transport from the volcanic island is observed, with major flank collapses feeding turbidite
522 currents (Frenz et al., 2009; Schneider et al., 1998). Schneider et al. (1998) also observed

523 increases of turbidite activity during a non-eruptive period on Gran Canaria, suggesting that
524 sediments were transported by low-density turbidity currents with some turbidites related to
525 the dynamics of the fluvial system. Funck and Schmincke (1998) showed that many of the
526 submarine canyons of the Canary Islands are the continuation of onshore canyons. Mitchell et
527 al. (2003) and Krastell et al. (2001) concluded that the dominant process feeding these
528 canyons was hyperpycnal flow.

529 For the Cilaos fan, the occurrence of turbidite activity during periods of low effusive volcanic
530 activity suggests that the fan was mainly fed by river sediment load. This means that the
531 turbidite activity occurred when erosional processes dominated, allowing a vast transport of
532 sediment over the submarine flanks of the volcano. These periods also correspond to phases
533 of explosive activity of Piton des Neiges, which could also have produced a large amount of
534 volcanoclastic material during eruption and a rapid transfer of sediments down to the adjacent
535 slope and basin, in a similar way to what was suggested for the Canary Islands (Schmincke and
536 Sumita, 1998).

537 As a result, the volcanic activity appears to be a major controlling factor influencing turbidite
538 development of the Cilaos deep-sea fan.

539

540 **6. Conclusions**

541

542 New stratigraphical data on the deep-sea Cilaos sedimentary system allow us to define the
543 timing of turbidite activity, which appears to have occurred close to the last two climatic
544 terminations. A first turbidite activity period occurred around 127 ka and a second one started
545 at 30 ka, which has continued until the present. The two main phases of turbidite activity
546 coincide with the last two transitions from glacial lowstands to subsequent sea level rises.
547 Nevertheless, our study demonstrates that sea level change played a minor role in the

548 triggering of turbidites in the Cilaos fan. On the other hand, the synchronicity between intense
549 turbidite deposition in the Cilaos deep-sea fan and periods of low effusive volcanic activity of
550 the Piton des Neiges volcano is clearly seen. We suggest that high erosional rates, identified
551 in the cirques during periods of low effusive activity, would mainly have contributed to
552 increase the seaward sediment transport. Conversely, the onset of volcanic activity would
553 have prevented intense erosion in the cirques, with the decrease of gravity deposits in the
554 Cilaos deep-sea fan resulting from the consequent low sediment transfer by rivers. Compared
555 with other volcanoclastic systems, explosive events of the Piton des Neiges might have also
556 contributed to feeding the Cilaos deep-sea fan.

557

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559

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571

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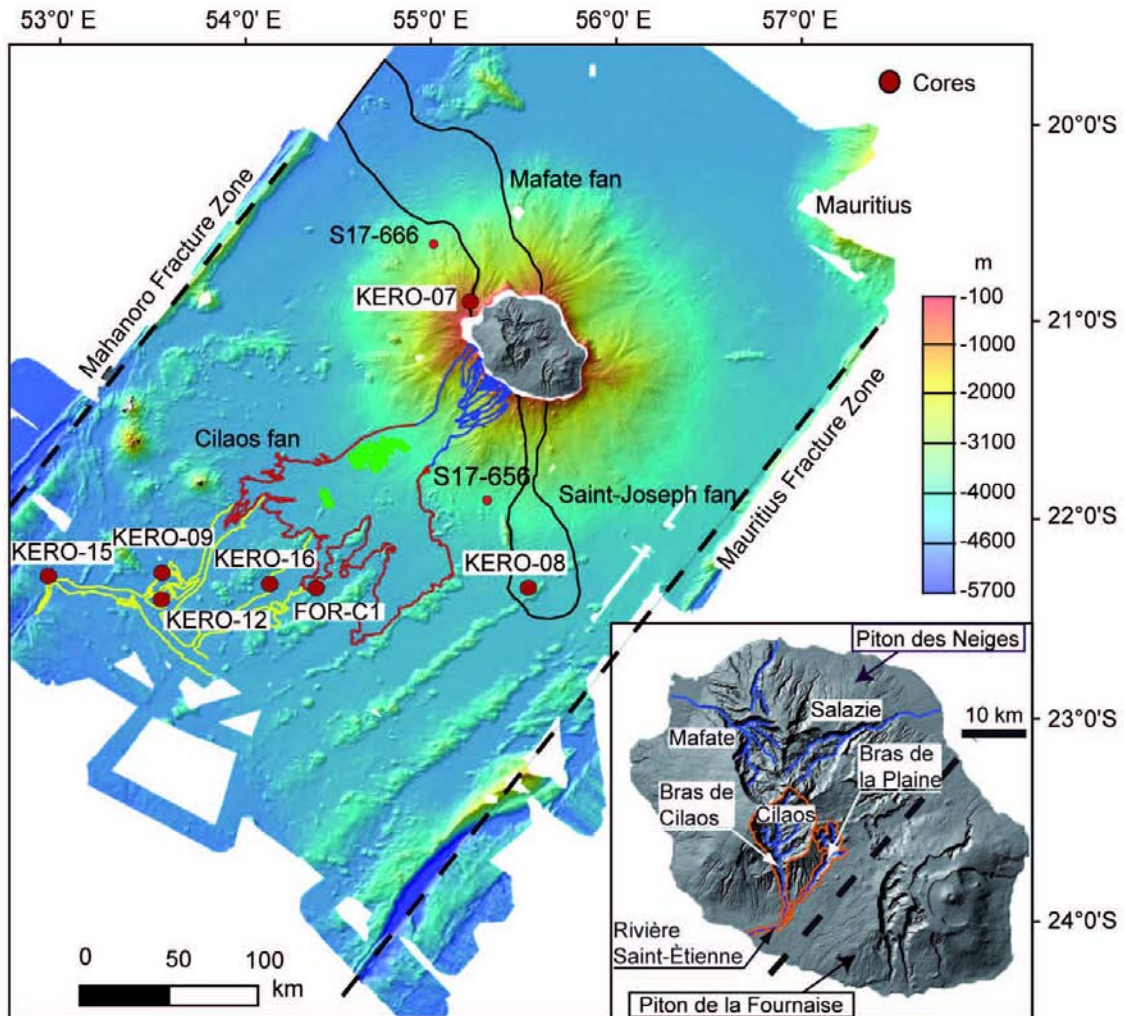
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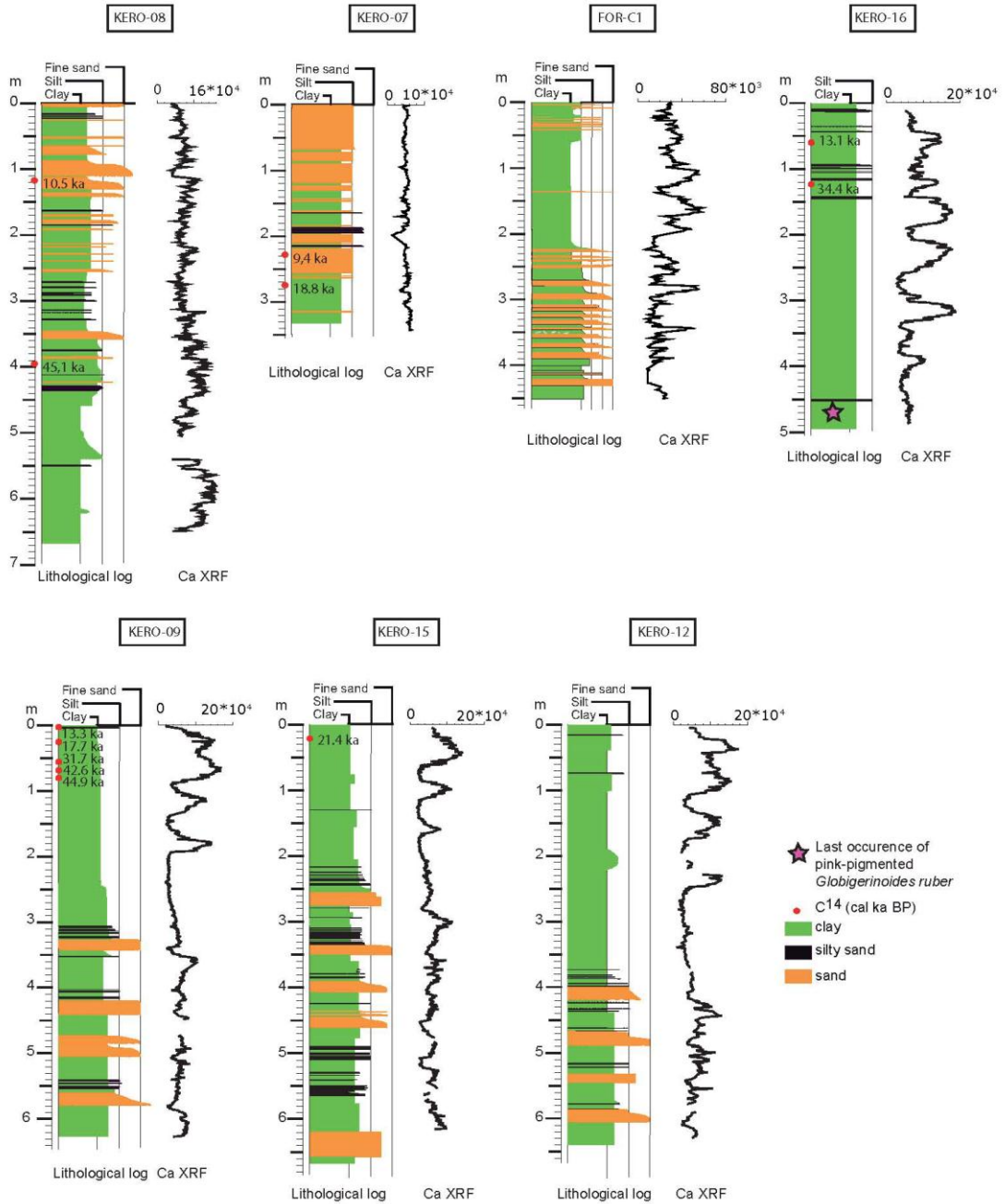
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780 **Figure captions**



781
 782 Fig. 1: Interpreted swath bathymetry image of the abyssal plain around La Réunion Island,
 783 compiled from the ERODER and FOREVER surveys. Red filled dots correspond to sediment
 784 cores presented in this paper. For the Cilaos fan, canyons are outlined in blue; the proximal
 785 fan in red and the distal part in yellow. Overview of the main geological structures of La
 786 Réunion Island (insert). The dashed line represents the separation between the two main
 787 volcanic edifices. The blue lines correspond to the main rivers draining the cirques. The
 788 orange lines delimit the main rivers feeding the Saint-Etienne river.

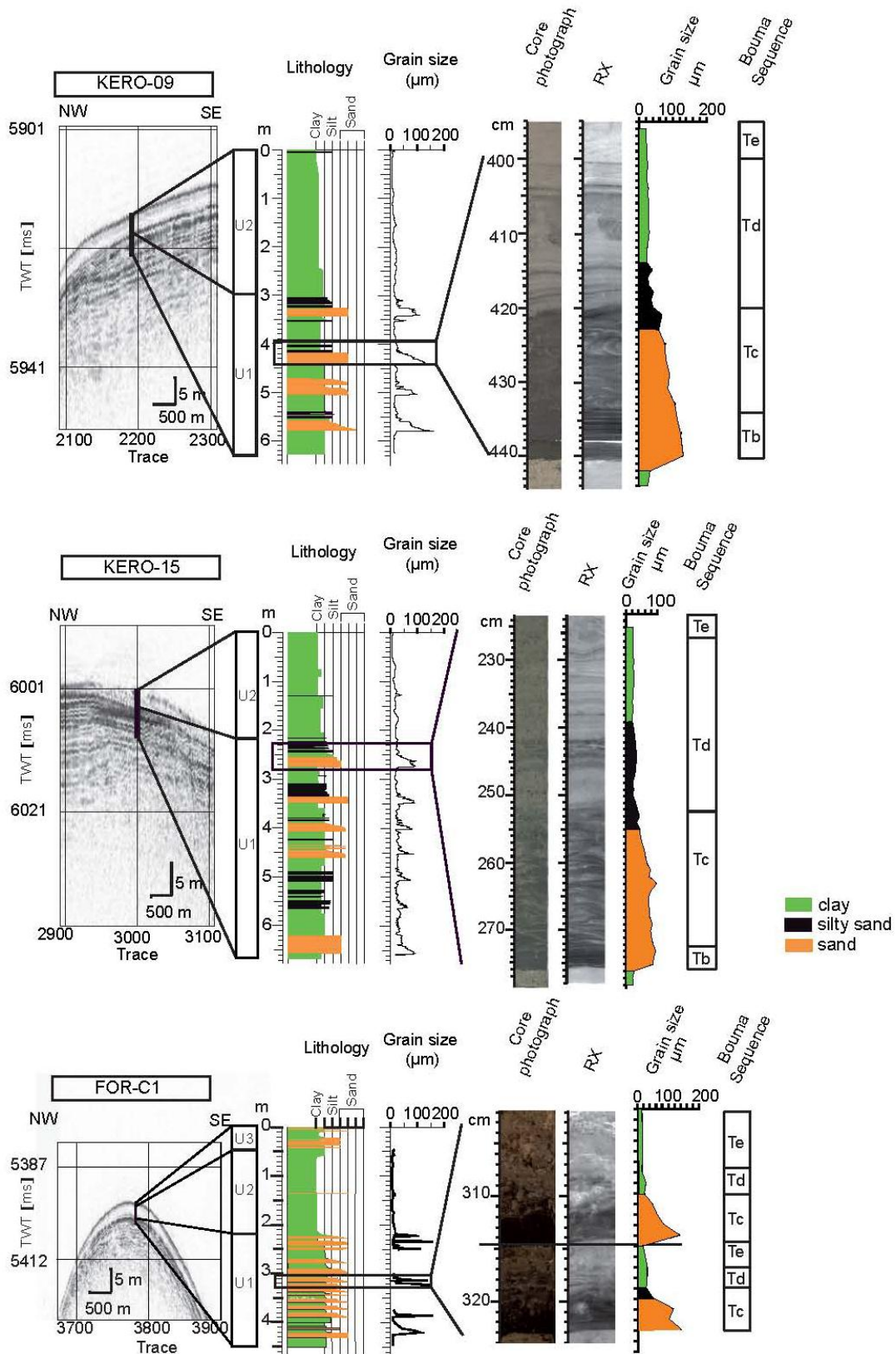


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791 Fig. 2: Lithological logs, fluctuations of Calcium XRF, and AMS ¹⁴C dates (cal ka BP) of

792 cores KERO-07, KERO-08, KERO-09, KERO-12, KERO-15, KERO-16 and FOR-C 1.

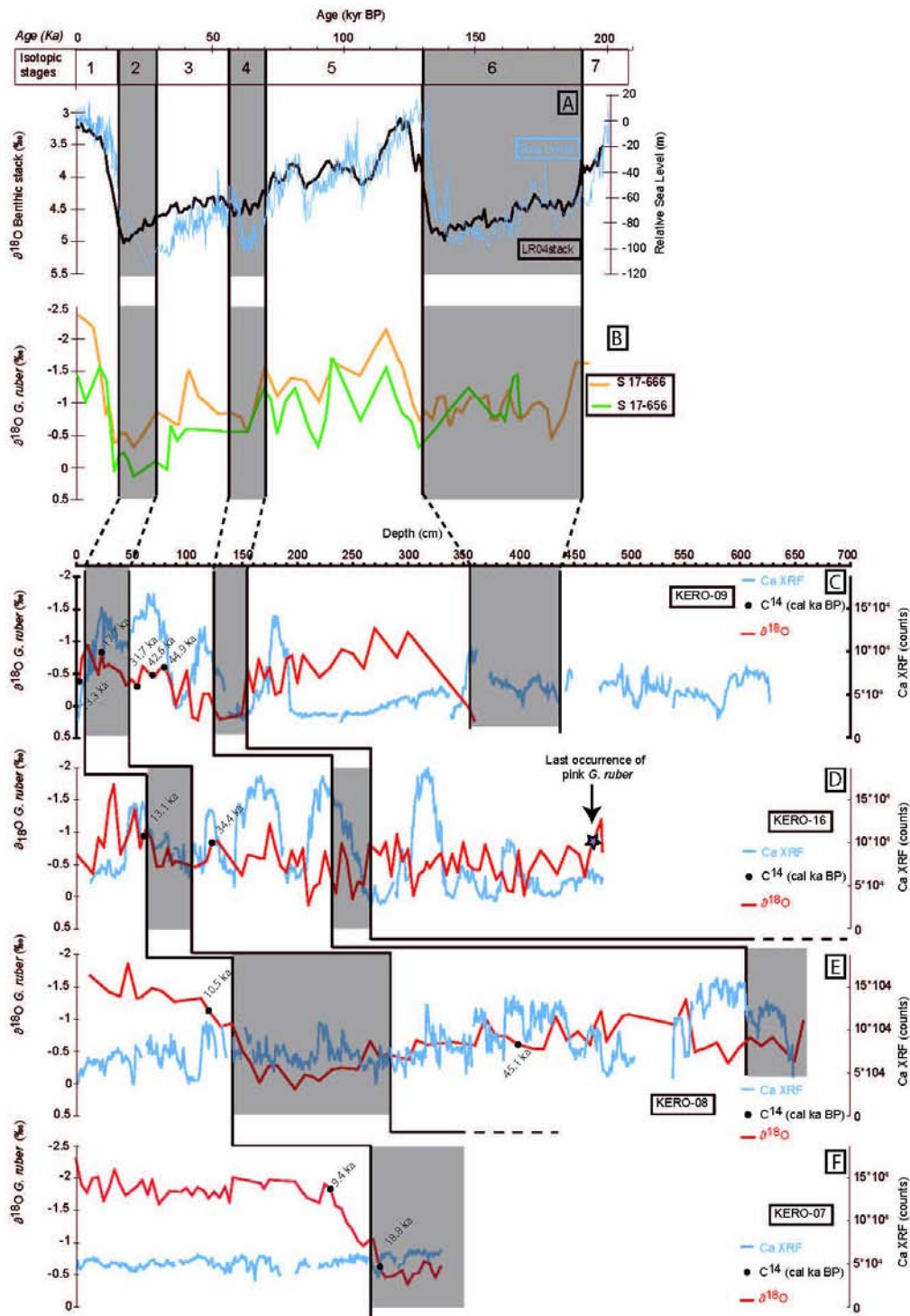


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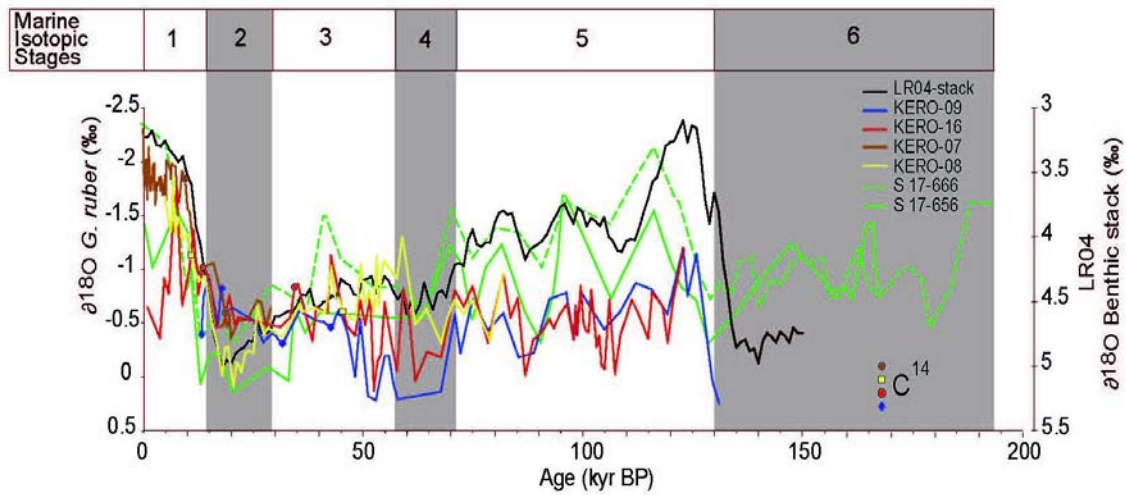
795 Fig. 3: Lithological log correlated with the corresponding echosounder profile, grain size

796 curve and X-ray image of cores KERO-09, KERO-15 and FOR-C .



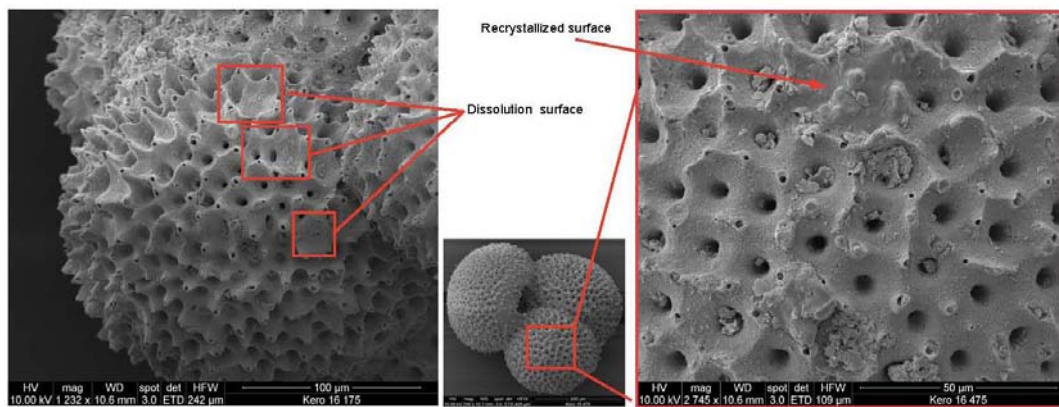
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798 Fig. 4: A: LR04 $\delta^{18}\text{O}$ benthic stack from Lisiecki and Raymo (2005). B: *G. ruber* $\delta^{18}\text{O}$ of
 799 cores S17-666 and S17-656 (Fretzdorff et al., 2000), locations shown on figure 1. C: *G. ruber*
 800 $\delta^{18}\text{O}$ and Ca XRF of core KERO-09. D: *G. ruber* $\delta^{18}\text{O}$ and Ca XRF of core KERO-16. E: *G.*
 801 *ruber* $\delta^{18}\text{O}$ and Ca XRF of core KERO-08. F: *G. ruber* $\delta^{18}\text{O}$ and Ca XRF of core KERO-07.



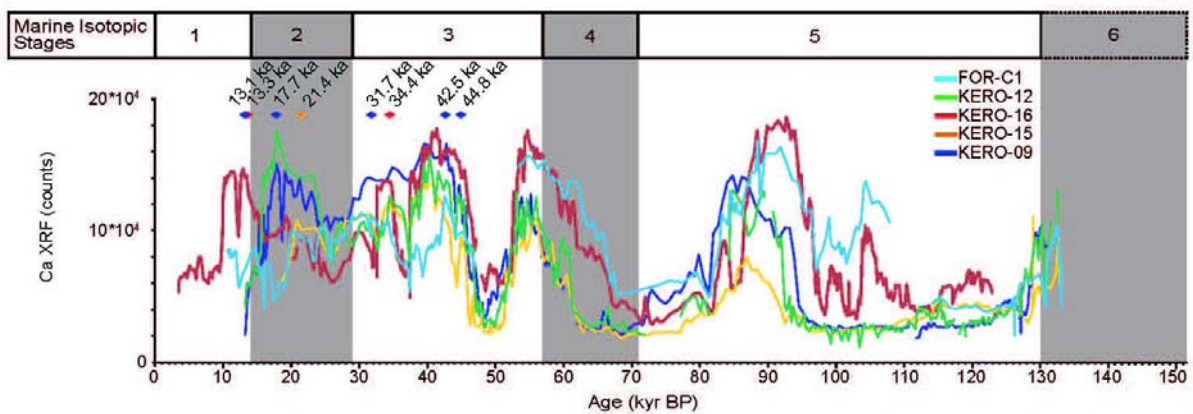
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803 Fig. 5: $\delta^{18}\text{O}$ curves of cores KERO-07, KERO-08, KERO-09 and KERO-16 compared with
 804 the LR04-stack curve of Lisiecki and Raymo (2005).



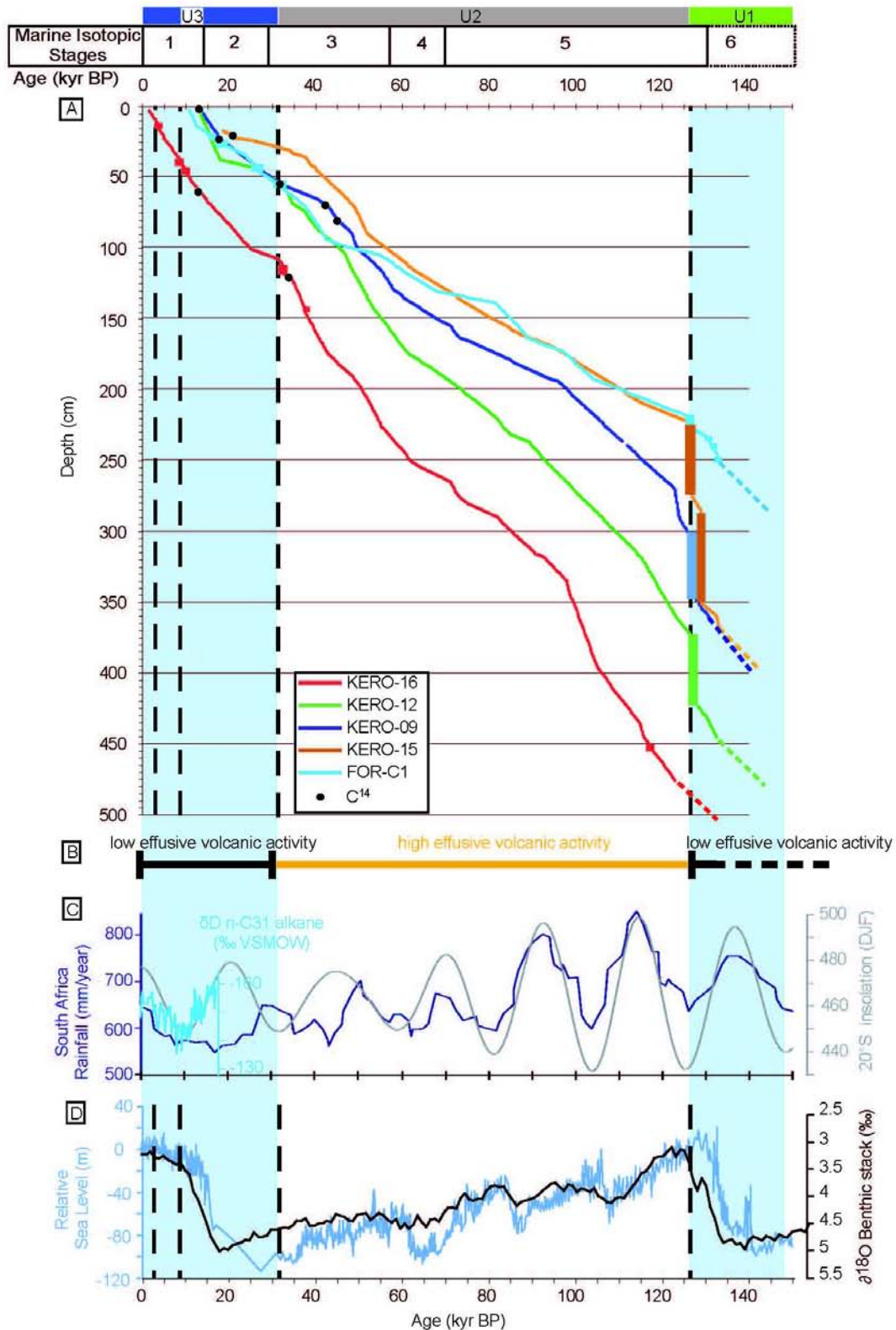
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806 Fig. 6: SEM microphotographs of *G. ruber* from core KERO-16. Recrystallization and
 807 dissolution surfaces are indicated on test surfaces.



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809 Fig. 7: Fluctuations of Calcium XRF for all the studied cores in the Cilaos deep-sea fan.

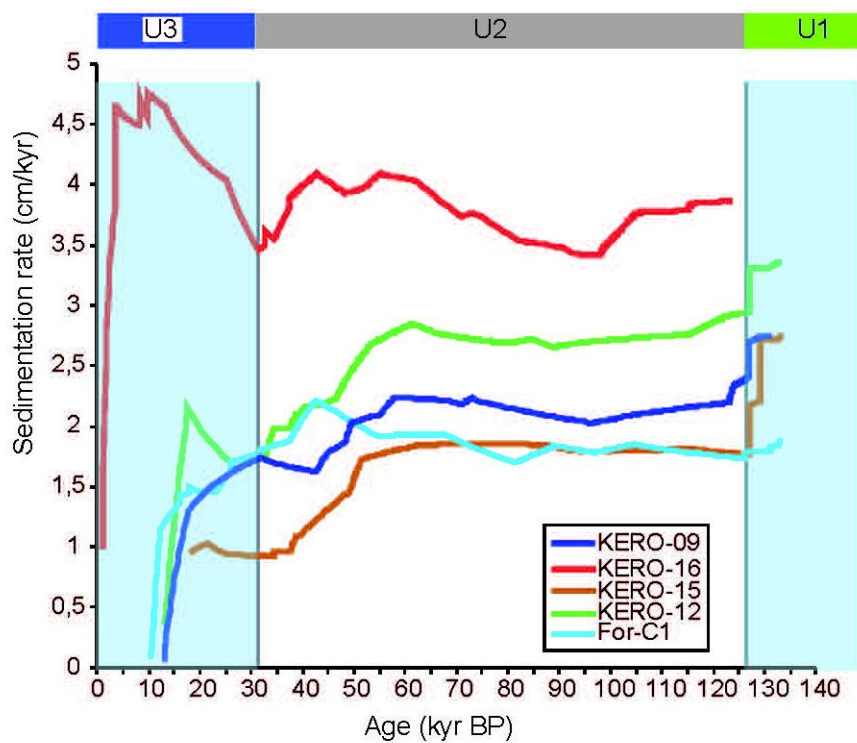


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811 Fig. 8: Timing of turbidite deposition in the Cilaos deep-sea fan. Blue areas indicate periods
 812 of major turbidite activity. The black dotted lines show the location of main turbidites. A)
 813 Age/depth model of the five cores KERO-09, KERO-12, KERO-15, KERO-16 and FOR-C1,

814 showing the location of turbidite beds in each core (turbidite beds are represented by
 815 rectangles along each age/depth curves; their vertical size is proportional to the turbidite
 816 thickness along the depth axis). Black dots show AMS ^{14}C dates. B) Major volcanic episodes
 817 of Piton des Neiges during the Late Quaternary. Dataset from Kluska (1997), Salvany et al.
 818 (2012). C) Hydrogen isotope compositions of the n-C31 alkane in GeoB9307-3, reflecting
 819 rainfall changes in the Zambezi catchment (Schefuß et al., 2012), Pretoria rainfall time series
 820 from Patridge et al. (1997) and insolation curve from Berger (1978). D) Red Sea sea level
 821 curve from core GeoTü-KL09 (Rohling et al., 2009) and LR04-stack curve of Lisiecki and
 822 Raymo (2005).

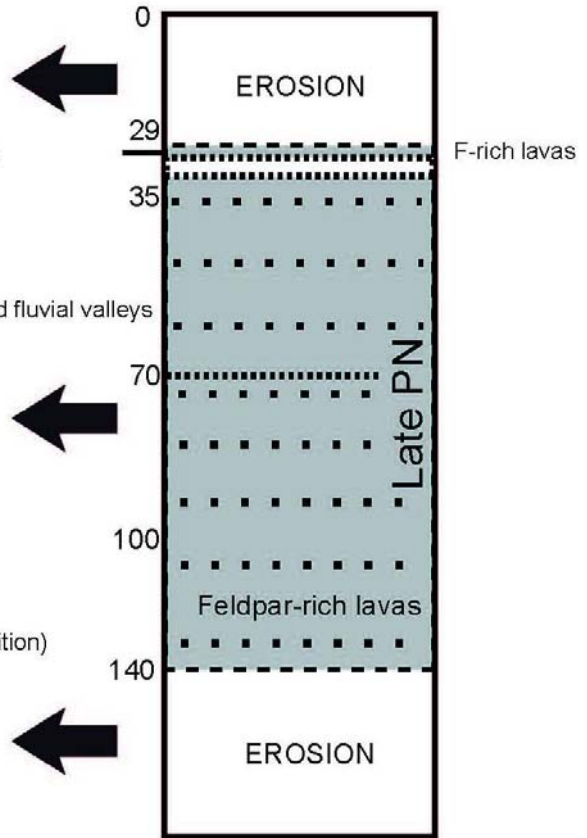
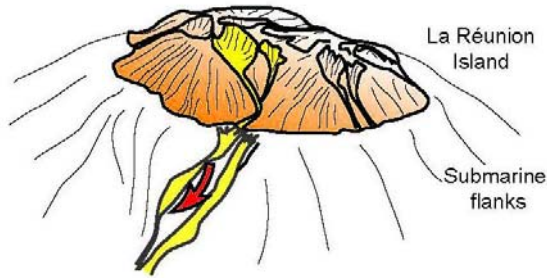
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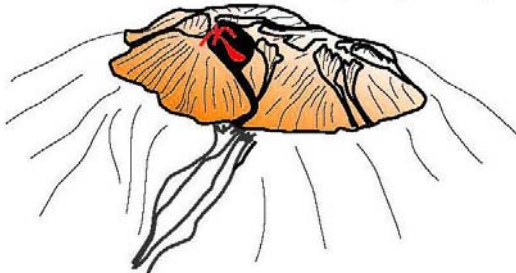
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825 Fig. 9: Sedimentation rate of the five cores of the Cilaos fan plotted versus time.

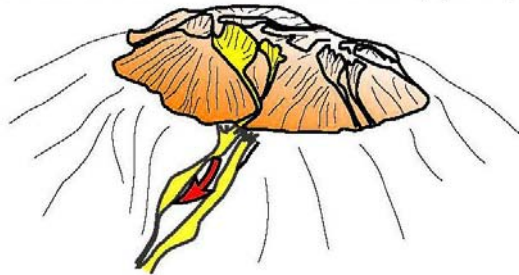
Since 30 ka: Erosion and turbidite activity (U3 deposition)



127-30 ka: Effusive volcanic activity filling the cirque and fluvial valleys



Before 127 ka: Erosion and turbidite activity (U1 deposition)



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827 Fig. 10: Schematic representation of the transport of sediments on the Cilaos deep-sea fan and
828 volcano-stratigraphic units of Piton des Neiges (modified from Salvany et al., 2012).

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Cruises	Cores	Lat (S)	Long (E)	Water Depth (m)	Location	Length (m)
FOREVER	FOR-C1	S22°20.95	E54°23.33	4074	Sedimentary ridge, Central Cilaos Fan	4.51
ERODER1	KERO-07	S20°53.59	E55°12.19	791	Levee, Mafate fan	3.40
ERODER 2	KERO-08	S22°20.89	E 55°31.12	4126	Distal part of Saint-Joseph fan	6.59
ERODER 2	KERO-09	S22°16.347	E53°33.060	4460	Levee, Occ. Cilaos Fan	6.27
ERODER 2	KERO-12	S22°23.550	E53°32.752	4461	Levee, Occ. Cilaos Fan	6.40
ERODER 2	KERO-15	S22°17.39	E52°56.10	4529	Distal levee, Occ. Cilaos Fan	6.68
ERODER 2	KERO-16	S22°19.51	E54°07.78	4340	Levee, Central Cilaos Fan	4.95

831

832 Table 1 : Location, bathymetry and length of the studied cores

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Laboratory number	Core	Depth <u>(cm bsf)</u>	AMS 14C age (yr)	AMS 14C age (-400yr)	Error yr	Calendar Age <u>(cal yr BP)</u>
Poz-35177	KERO-09	3	11 840	11 440	60	13 302
Poz-35178	KERO-09	23	14 980	14 580	70	17 739
Poz-35179	KERO-09	55	28 000	27 600	240	31 723
Poz-35180	KERO-09	69	38 500	38 100	600	42 587
Poz-35181	KERO-09	80	41700	41 300	1000	44 883
SacA 24240	KERO-08	118-119	9680	9280	40	10 543
SacA 24241	KERO-08	395-396	42 060	41 660	690	45 133
SacA 24239	KERO-07	229-230	8755	8355	40	9436
SacA 21880	KERO-07	274.5	16 110	15 710	50	18 834
SacA 21881	KERO-15	21.5	18 390	17 990	60	21 443
SacA 21882	KERO-16	60.5	11 610	11 210	35	13 118
SacA 21883	KERO-16	122.5	29 660	29 260	160	34 422

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837 Table 2 : Radiocarbon dates from cores KERO-07, KERO-08, KERO-09, KERO-15 and
838 KERO-16

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Sample	Depth (cm bsf)	Sedimentary Facies	Observation
KERO-09-65	65	Layer of light brown clay, located in the upper part of the core	<i>Emiliana huxleyi</i> abundant (<75-90 Ka)
KERO-09-115	115	Layer of light brown clay,	Abundant, good preservation, <i>Pseudoemiliana lacunosa</i> being absent (<460 ka) and contains <i>Emiliana huxleyi</i> (<260 ka)
KERO-09-117	117	Layer of light brown clay	Abundant, good preservation, <i>Pseudoemiliana lacunosa</i> being absent (<460 ka) and contains <i>Emiliana huxleyi</i> (<260 ka)
KERO-09-361	361	Layer of brown clay, located between two sandy layers	Abundant, good preservation, <i>Pseudoemiliana lacunosa</i> being absent (<460 ka) and contains <i>Emiliana huxleyi</i> (<260 ka). Reworked nannofossils were noted
KERO-09-608	608	Layer of brown clay, located in the lower part of the core	Abundant, poor preservation. <i>Gephyrocapsa spp cf caribbeanica</i> dominant. Two specimens of <i>Pseudoemiliana lacunosa</i> suggest age <460ka.

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Table 3 : Calcareous nannofossil data of core KERO-09. Age from Berggren et al. (1995).