

1 Frequency and intensity of palaeofloods at the interface of 2 Atlantic and Mediterranean climate domains

3
4 **B. Wilhelm^{1,2}, H. Vogel², C. Crouzet^{3,4}, D. Etienne⁵, F.S. Anselmetti²**

5
6 [1]{Univ. Grenoble Alpes, LTHE, F-38000 Grenoble, France}

7 [2]{Institute of Geological Sciences and Oeschger Centre for Climate Change Research,
8 Univ. of Bern, CH-3012 Bern, Switzerland}

9 [3]{Univ. Savoie Mont Blanc, ISTerre, F-73376 Le Bourget-du-Lac, France}

10 [4]{CNRS, ISTerre, F-73376 Le Bourget-du-Lac, France}

11 [5]{UMR INRA 42 CARRTEL, Univ. Savoie Mont Blanc, F-73376 Le Bourget du Lac,
12 France}

13 Correspondence to: B. Wilhelm (bruno.wilhelm@ujf-grenoble.fr)

14 15 **Abstract**

16 The long-term response of the flood activity to both Atlantic and Mediterranean climatic
17 influences was explored by studying a lake sequence (Lake Foréant) of the Western European
18 Alps. High-resolution sedimentological and geochemical analysis revealed 171 event layers,
19 168 of which result from past flood events over the last millennium. The layer thickness was
20 used as a proxy of intensity of past floods. Because the Foréant palaeoflood record is in
21 agreement with the documented variability of historical floods resulting from local and
22 mesoscale, summer-to-autumn convective events, it is assumed to highlight changes in flood
23 frequency and intensity related to such events typical of both Atlantic (local events) and
24 Mediterranean (meso-scale events) climatic influences. Comparing the Foréant record with
25 other Atlantic-influenced and Mediterranean-influenced regional flood records highlights a
26 common feature in all flood patterns that is a higher flood frequency during the cold period of
27 the Little Ice Age (LIA, AD 1300-1900). In contrast, high-intensity flood events are apparent
28 during both, the cold LIA and the warm Medieval Climate Anomaly (MCA, AD 950-1250).
29 However, there is a tendency towards higher frequencies of high-intensity flood events during
30 the warm MCA. The MCA extremes could mean that under the global warming scenario, we
31 might see an increase in intensity (not in frequency). However, the flood frequency and
32 intensity in course of 20th century warming trend did not change significantly. Uncertainties in

33 future evolution of flood intensity lie in the interpretation of the lack of 20th century extremes
34 (transition or stable?) and the different climate forcing factors between the two periods
35 (greenhouse gases vs. solar/volcanic eruptions).

36

37 **Key-words:** palaeoflood record, flood frequency, flood intensity, Atlantic influence,
38 Mediterranean influence, last millennium

39

40 **1. Introduction**

41 Heavy rainfall events trigger mountain-river floods, one of the most significant natural
42 hazards, causing widespread loss of life, damage to infrastructure and economic deprivation
43 (e.g. Kundzewicz et al., 2014). This is especially the case for the Alpine area in Europe,
44 where tourism and recent demographic development with an increasing population raise the
45 vulnerability of infrastructure to natural hazards (e.g. Beniston and Stephenson 2004).
46 Moreover, the current global warming is expected to lead to an intensification of the
47 hydrological cycle and a modification of flood hazard (IPCC et al., 2013). Hence, a robust
48 assessment of the future evolution of the flood hazard over the Alps becomes a crucial issue.

49 A main limitation for robust flood-hazard projections is the scarce knowledge on the
50 underlying natural climate dynamics that lead to these extreme events (IPCC, 2013). Indeed,
51 the stochastic nature and the rare occurrence of extreme events make the identification of
52 trends based on instrumental data alone difficult (e.g. Lionello et al., 2012). One way of
53 overcoming this issue is to extend flood series beyond observational data and compare these
54 datasets with independent climatic and environmental forcing. In this purpose, many types of
55 sedimentary archives have been studied (e.g. Luterbacher et al., 2012 and references therein).
56 Among them lake sediments are being increasingly studied because they allow to reconstruct
57 flood records long enough to identify the natural variability at different time scales (e.g.
58 Noren et al., 2002; Oslegger et al., 2009; Wilhelm et al., 2012a; Czymzik et al., 2013; Glur et
59 al., 2013; Corella et al., 2014).

60 In the western Alps, many lake-sediment sequences have been studied to better assess the
61 response of the flood activity to climate variability. These studies revealed higher flood
62 frequency of mountain streams in many regions during multi-centennial cold phases such as
63 the Little Ice Age (Giguët-Covex et al., 2012; Wilhelm et al., 2012a; 2013; Glur et al., 2013;
64 Wirth et al., 2013b; Amann et al., 2015). However, regarding flood intensity/magnitude,

65 opposite patterns appear with the occurrence of the most extreme events during warmer
66 periods in the north (Giguet-Covex et al., 2012; Wilhelm et al., 2012b; 2013), while they
67 occurred during colder periods in the south (Wilhelm et al., 2012a; 2015). These north-south
68 opposite flood patterns were explained by flood-triggering meteorological processes specific
69 to distinct climatic influences: Atlantic in the north versus Mediterranean in the south. In the
70 north-western part of the Alps, floods at high altitude are mainly triggered by local convective
71 events (i.e. thunderstorms) and seem to mainly depend on the temperature that would
72 strengthen vertical processes (e.g. Wilhelm et al., 2012b; 2013). In contrast, floods in the
73 south are mostly triggered by mesoscale events and may strongly depend on pathways and
74 intensity of storm-tracks (e.g. Trigo and Davis, 2000; Boroneant et al., 2006; Boudevillain et
75 al., 2009). By analogy with these results over past warm periods, the mountain-flood hazard
76 might be expected to increase in the north-western Alps, mainly because of an enhanced flood
77 magnitude associated to stronger convective processes. Hence, better assessing the spatial
78 extent of the Atlantic-influenced flood pattern at high-altitude appears a crucial issue to
79 appropriately establish hazard mitigation plans and prevent high socio-economic damages.

80 In this context, the present study was designed to reconstruct the flood pattern at an
81 intermediate situation between the north-western and south-western Alps, i.e. at the climate
82 boundary between Atlantic and Mediterranean influences. This is undertaken by
83 reconstructing a millennium-long flood chronicle from the sediment sequence of the high-
84 altitude Lake Foréant located in the Queyras Massif (France).

85

86 **2. Regional setting**

87

88 **2.1. Hydro-climatic setting and historical flood record**

89 The Queyras massif is located in between the northern and southern French Alps where the
90 climate is influenced by the Atlantic Ocean and the Mediterranean Sea (Fig. 1). As a result,
91 the Queyras mountain range corresponds to a transition zone of Alpine precipitation patterns
92 in the meteorological reanalyses (Durant et al., 2009; Plaut et al., 2009) and in the simulations
93 (Frei et al., 2006; Rajczak et al., 2013). In the Queyras, heavy precipitation events are related
94 to either local convective phenomena (i.e. summer thunderstorms) or mesoscale convective
95 systems. The mesoscale systems called “Lombarde-Type” or “East Return” events occur
96 mainly from late spring to fall and result from Mediterranean humid air masses flowing

97 northward into the Po Plain and then westward to the Queyras Massif (e.g. Gottardi et al.,
98 2010; Parajka et al., 2010). The humid air masses are then vigorously uplifted with the steep
99 topography of the Queyras massif, causing an abrupt cooling of the air masses and intense
100 precipitations. Such mesoscale precipitation events, typical of the Mediterranean climate (e.g.
101 Buzzi and Foschini, 2000; Lionello et al., 2012), affect extensive areas and may lead to
102 catastrophic floods at a regional scale as shown in June 1957 or October 2000 over the
103 Queyras massif (Arnaud-Fassetta and Fort, 2004). Other numerous past flood events were
104 documented from studies of local historical records. These data have been compiled in a
105 database managed by the ONF-RTM (<http://rtm-onf.ifn.fr/>). They show that the village of
106 Ristolas (located 8 km downstream from Lake Foréant, Fig. 1C) has been affected at least 34
107 times over the last 250 years by floods of the Guil River and its five main tributaries (see
108 Supplementary Material).

109

110 **2.2. Lake Foréant and its tributaries**

111 Lake Foréant (2620 m a.s.l., 44°42'20"N, 6°59'00"E) is located in a cirque of 3 km² in the
112 upper part of the Queyras Massif, adjacent to the Italian border (Fig. 1). It is located
113 approximately 60 km north from Lake Allos and 100 km south-west from Lakes Blanc,
114 whose hydro-climatic settings are characterized by the south-western and north-western flood
115 pattern, respectively (Wilhelm et al., 2012a,b; Fig. 1B). The catchment rises up to 3210 m.
116 a.s.l. and is made up of three lithologies from the Queyras schistes-lustrés nappe (e.g.
117 Schwartz et al., 2009); (i) marble in the eastern part, (ii) calc-schist in the western part and
118 (iii) a narrow band of arkose in between (Fig. 1D). The main stream of the catchment, the
119 Torrent de Bouchouse, drains mainly the central band of arkose. Before entering the lake, this
120 stream has built an alluvial plain (Fig. 1E). This suggests that the Bouchouse stream is a
121 major source of sediment entering the lake. In addition, two minor and non-permanent
122 streams drain the western part of the catchment. In contrast, they enter the lake through only
123 small deltas compared to the Bouchouse inflow area, suggesting limited detrital inputs. There
124 is no evidence that the catchment was glaciated in the past, i.e. no moraine or other glacial
125 deposits occur. Thereby, detrital inputs only result from the erosion and transport of these
126 lithologies. Detrital inputs from these streams are limited to summer and fall because the
127 catchment is covered by snow and the lake is frozen from mid-November to the beginning of
128 June. The Bouchouse stream flows downstream into the Guil River and reaches
129 approximately 8 km further the village of Ristolas (Fig. 1C).

130

131 **3. Method**

132

133 **3.1. Lake coring**

134 In summer 2013, a bathymetric survey was carried out on Lake Foréant and revealed a well-
135 developed flat basin in the centre of the lake with a maximum water depth of 23.5 m (Fig.
136 1E). An UWITEC gravity corer was used to retrieve four cores along a north-south transect in
137 the axis of the two main inlets of the Bouchouse stream. Cores FOR13P3 and FOR13P4
138 correspond to the proximal locations of the two different branches of the Bouchouse stream
139 and aim at investigating their respective sediment inputs during floods. Cores FOR13P2 and
140 FOR13P1 correspond to the depocenter and to the most distal position, i.e. the opposite slope
141 to the Bouchouse inflows, respectively.

142

143 **3.2. Core description and logging**

144 Cores were split lengthwise and the visual macroscopic features of each core were examined
145 to identify the different sedimentary lithofacies. The stratigraphic correlation between the
146 cores was then carried out based on these defined lithofacies. The stratigraphic correlation
147 allows identifying the depositional pattern of the event layers within the lake basin.
148 Depositional patterns of event layers may help to decipher their trigger, i.e. mass-movements
149 vs. flood events (e.g. Sturm and Matter, 1978; Wilhelm et al., 2012b; Van Daele et al., 2015).

150 High-resolution color line scans and gamma-ray attenuation bulk density measurements were
151 carried out on a Geotek™ multisensor core-logger (Institute of Geological Sciences,
152 University of Bern). Bulk density was used as a proxy of event layers, e.g. flood layers,
153 characterized by higher density due to the high amount of detrital material (e.g. Støren et al.,
154 2010; Gilli et al., 2012; Wilhelm et al., 2012b).

155 Geochemical analysis and X-ray imaging were carried out using an Itrax™ (Cox Analytical
156 Systems) X-ray fluorescence (XRF) core scanner (Institute of Geological Sciences, University
157 of Bern), equipped with a Molybdenum tube (50 keV, 30 mA) with a 10-s count-time using
158 sampling steps of 1 mm (XRF) and 0.2 mm (X-ray imaging). The areas of the element peaks
159 obtained are proportional to the concentrations of each element (Tachikawa et al., 2011).
160 Geochemical data were applied to identify event layers at high resolution through their higher

161 content in detrital material (e.g. Arnaud et al., 2012; Wilhelm et al., 2012b; Czymzik et al.,
162 2013; Swierczynski et al., 2013) and/or as high-resolution grain-size proxies (e.g. Cuven et
163 al., 2010; Wilhelm et al., 2012a; 2013). Geochemical analyses were carried out on core
164 FOR13P2. X-ray images highlighting the variability of the sediment density have been
165 acquired for the four cores.

166 Grain size analyses on core FOR13P2 were performed using a Malvern Mastersizer 2000
167 (Institute of Geography, University of Bern) on sub-samples collected at a 5-mm continuous
168 interval. Before grain-size analysis, the samples were treated in a bath of diluted (30%)
169 hydrogen peroxyde during 3 days to remove organic matter. After treatment, microscopic
170 observations were performed to control that organic matter was totally dissolved. These grain-
171 size analyses of the detrital material were performed to characterize event layers and, when
172 event layers are induced by floods, to establish a proxy of flood intensity. Grain-size
173 variability is assumed to be related with the stream flow energy of the river entering the lake
174 and, thereby, with the peak discharge reached during the flood event (Campbell, 1998;
175 Lapointe et al., 2012; Wilhelm et al., 2015). The flood intensity may also be reconstructed
176 based on the amount of sediment transported and deposited during floods (e.g. Schiefer et al.,
177 2011; Jenny et al., 2014; Wilhelm et al., 2015). This approach appeared particularly relevant
178 in case of an insignificant variability of the flood-sediment grain size (e.g. Jenny et al., 2014;
179 Wilhelm et al., 2015). When the depositional pattern of the flood layers within the lake basin
180 (assessed through the stratigraphic correlation) shows a high variability, many cores are
181 required for a reliable assessment of the flood-sediment volumes (e.g. Page et al., 1994;
182 Schiefer et al., 2011; Jenny et al., 2014). However, when the depositional pattern of the flood
183 layers is stable over time, the layer thickness in one core may be sufficient (e.g. Wilhelm et
184 al., 2012b; 2015). As a result, grain size and sediment volume of the event layers were both
185 explored as two distinct proxies of flood intensity.

186

187 **3.3. Coprophilous fungal spores analysis**

188 Erosion processes in high-altitude catchments may be modified by grazing activity and,
189 thereby, the climatic signal in flood reconstructions may be altered (e.g. Giguet-Covex et al.,
190 2012). The variability of grazing intensity in a catchment area can be reconstructed from the
191 sedimentary abundance of coprophilous fungal ascospores, i.e. *Sporormiella* (HdV-113) (e.g.
192 Davis and Schafer, 2006; Etienne et al., 2013). To test the potential impact of grazing

193 intensity on the reconstructed flood activity, *Sporormiella* abundance was determined in
194 subsamples collected all along the core FOR13P3 with an approximate step of 3 cm. This core
195 was chosen because it was the sequence with the thinnest potentially-erosive event layers.
196 During the sampling, event layers were avoided because they may correspond to layers
197 induced by flood or mass-movement events that may have transported unusual quantities of
198 *Sporormiella* ascospores, or induced the remobilization of older sediments. Subsamples were
199 chemically prepared according to the procedure of Fægri and Iversen (1989). *Lycopodium*
200 *clavatum* tablets were added in each subsample (Stockmarr, 1971) to express the results in
201 concentrations (number.cm⁻³) and accumulation rates (number.cm².yr⁻¹). Coprophilous fungal
202 ascospores were identified based on several catalogues (Van Geel and Aptroot, 2006; Van
203 Geel et al., 2003) and counted following the procedure established by Etienne and Jouffroy-
204 Bapicot (2014).

205

206 **3.4. Dating methods**

207 For dating the lake sequence over the last century, short-lived radionuclides (²¹⁰Pb, ¹³⁷Cs)
208 were measured by gamma spectrometry at the EAWAG (Zürich, Switzerland). The core
209 FOR13P4 was sampled following a non-regular step of 1±0.2 cm. The non-regular step aims
210 at matching the facies (i.e. sedimentary background or event layer) boundaries for
211 homogeneous samples. The ¹³⁷Cs measurements allowed two main chronostratigraphic
212 markers to be located: the fallout of ¹³⁷Cs from atmospheric nuclear weapon tests culminating
213 in AD 1963 and the fallout of ¹³⁷Cs from the Chernobyl accident in AD 1986 (Appleby,
214 1991). The decrease in excess ²¹⁰Pb and the Constant Flux/Constant Sedimentation (CFCS)
215 allowed a mean sedimentation rate to be calculated (Goldberg, 1963). The standard error of
216 the linear regression of the CFCS model was used to estimate the uncertainty of the
217 sedimentation rate obtained by this method. The ¹³⁷Cs chronostratigraphic markers are then
218 used to control the validity of the ²¹⁰Pb-based sedimentation rate.

219 To date the sequence beyond the last century, small-size vegetal macro-remains were sampled
220 in core FOR13P4. Terrestrial plant remains were isolated at the Institute of Plant Sciences
221 (University of Bern) and sent for AMS ¹⁴C analysis to the AMS LARA Laboratory
222 (University of Bern). ¹⁴C ages were calibrated using the Intcal13 calibration curve (Reimer et
223 al., 2013; Table 1). In addition, palaeomagnetic chronostratigraphic markers were also used.
224 These markers can be obtained by comparing the characteristic declination and inclination of

225 remanent magnetization (ChRM) versus depth to global geomagnetic models or to known
226 secular variations of the geomagnetic field (e.g. Barletta et al., 2010; Wilhelm et al., 2012a).
227 Palaeomagnetic investigations were performed at the CEREGE laboratory (University Aix-
228 Marseille) and are detailed in supplementary material.

229 Using the R-code package “clam” (Blaauw, 2010), an age-depth model was generated from
230 the ^{210}Pb ages, the ^{14}C ages and the palaeomagnetic chronological markers.

231

232 **4. Results**

233

234 **4.1. Sedimentology**

235 The sediment consists of a homogeneous, brown mud mainly composed of silty detrital
236 material and aquatic organic remains (small fragments of plants and anamorphous organic
237 matter), representing the background hemi-pelagic sedimentation. These fine grained deposits
238 are interrupted by 171 rather coarser-grained layers, which are interpreted to represent short-
239 term depositional events, i.e. event layers (Fig. 2).

240 The 171 event layers correspond to graded layers, characterized by their higher density, a
241 slight fining-upward trend and a thin, whitish fine-grained capping layer (Fig. 2). There is no
242 evidence of an erosive base. According to the stratigraphic correlation, almost all these graded
243 layers (168 out of 171) extend over the entire lake basin with a regular deposition pattern. The
244 thickness of these 168 graded layers is systematically larger in cores FOR13P2 and FOR13P4,
245 and decreases respectively in cores FOR13P1 and FOR13P3. This suggests that the southern
246 branch of the Bouchouse stream is the main sediment input over the studied period (Fig. 1).
247 The grain-size of these graded layers is dominated by silt-sized grains with only small
248 amounts of clay / fine silt present in the whitish capping layer and to coarse silt in their basal
249 part (Fig. 2). The origin of these 168 is discussed in section 5.1. The three other graded layers
250 show a higher variability in thickness, grain size and depositional pattern. Above all, they
251 overlie three cm-thick coarse-grained layers, present at 75 cm in core FOR13P2 and at 9 and
252 42 cm in core FOR13P4 (Fig. 2). The coarse-grained layers consist of pebble gravels and
253 aquatic plant remains embedded in a silty matrix. The high porosity in the sediment due to the
254 presence of gravels generates a partial loss of XRF signal, preventing a reliable geochemical
255 characterization. X-ray images show chaotic sedimentary structures. The stratigraphic
256 correlation revealed that two centimetres of sediment are missing below the thickest coarse-

257 grained layer in core FOR13P4, suggesting an erosive base for this layer. The stratigraphic
258 succession of a coarse-grained layer overlain by a graded layer (labelled MMIL in Fig. 2 and
259 3) suggests that the two layers were induced by a single event. Their origin is discussed in
260 section 5.1.

261

262 **4.2. Geochemistry**

263 Among the core scanner output parameters, the scattered incoherent (Compton) radiation of
264 the X-ray tube (Mo_{inc}) may vary with the sediment density (Croudace et al., 2006) and,
265 thereby, offer a high-resolution proxy for sediment density. Mo_{inc} values were averaged at a
266 5-mm resolution to be compared to the density obtained at a 5-mm resolution with the
267 gamma-ray attenuation method. A positive and significant correlation ($r=0.85$, $p<10^{-4}$)
268 between the two density parameters was found and allowed using Mo_{inc} as a proxy of
269 sediment density for identifying millimetre-scale event layers (Fig. 3).

270 The variability of other elements within the event layers was then investigated to assess i) a
271 high-resolution grain-size proxy and ii) distinct sediment sources of the event layers between
272 the littoral (i.e. mass-movement origin) and the catchment area (i.e. flood origin). The
273 variability of potassium (K) intensities vs. sediment depth (Fig. 3) shows increased K
274 intensities in the capping layers of the event layers, suggesting K enrichment in the finest
275 sediment fraction. Variability in silicon (Si) intensities is correlated to K intensities ($r=0.77$,
276 $p<10^{-4}$). Variations in iron (Fe) intensities show an opposite pattern with Fe enrichments in
277 the basal and coarser part of the graded beds. Interestingly, Fe is the only element which
278 elevated in event layers. These results suggest that the Fe/K ratio may be used as a millimetre-
279 scale proxy for relative grain-size distribution and hence for detecting millimetre-scale event
280 layers. However, since grain-size variability is insignificant, the information that can be won
281 from this proxy in regard to flood-intensity reconstruction is minor.

282 Ca intensities are most of the time very low (< 900 counts), except for several sharp peaks and
283 two well-marked excursions (> 1200 counts) at 30 and 75 cm in core FOR13P2 (Fig. 3).
284 These two well-marked excursions correspond to the two thickest graded layers (labelled
285 MMIL2 and 3; Fig. 2). In addition, manganese (Mn) intensities also vary within a low value
286 range ($< 10^4$ counts) interrupted by sharp, well-marked peaks (up to $4 \cdot 10^4$ counts). All those
287 Mn peaks are located at the base of the 168 graded layers (those that not overlain a coarse-
288 grain layers). However not every base of graded layers corresponds to a Mn peak. To better

289 assess the relationships between those elements, the Ca intensities were plotted against Fe, K
290 and Mn intensities (Fig. 3). These plots clearly highlight two groups of deposits. The
291 background sediments and the 168 graded layers (those labelled FIL in Fig. 3) are
292 characterized by i) low Ca intensities and ii) a high variability in Mn intensities. The three
293 graded layers labelled MMIL in figure 2 show a distinct geochemical pattern with (i) high Ca
294 intensities regardless of Fe and K intensities and (ii) very low Mn intensities.

295

296 **4.3. Chronology**

297 The down-core ^{210}Pb excess profile for core FOR13P2 shows a continuous decrease to low
298 values ($\sim 50 \text{ Bq.Kg}^{-1}$), punctuated by sharp excursions to low values for three layers (2-3.5
299 cm, 7.5-10.5 cm and 15-17 cm) corresponding to graded layers (Fig. 4A). In line with Arnaud
300 et al. (2002), these values were excluded to build a corrected sedimentary record without
301 event layers (Fig. 4B). The CFCS model (Goldberg, 1963), applied on the event-free ^{210}Pb
302 excess profile, provides a mean sedimentation rate of $1.3 \pm 0.1 \text{ mm.yr}^{-1}$ (without the event
303 layers). Ages derived from the CFCS model were transposed to the original sediment
304 sequence to provide a continuous age-depth relationship (Fig. 4C). The event-free ^{137}Cs
305 profile indicated two peaks at 3.5 cm and 5.5 cm (Fig. 4B), interpreted as the result of the
306 Chernobyl accident in AD 1986 and the maximum fallout of the nuclear weapon tests in AD
307 1963. These independent chronological markers are in good agreement with the ^{210}Pb excess
308 ages, supporting the age-depth model over the last century (Fig. 4C).

309 In order to evaluate the reliability and efficiency of the palaeomagnetic results several points
310 need to be verified: i) the preservation of the sedimentary magnetic fabric, ii) the stability of
311 magnetic mineralogy, and iii) the stability of the magnetic components. Results of Anisotropy
312 of Magnetic Susceptibility for core FOR13P4 show a well preserved sedimentary fabric, i.e.
313 K_{min} inclination close to the vertical, except in the thickest event layers (labelled MMIL2
314 and MMIL3, Fig. 2) where the K_{min} inclination is clearly deviated (Fig. S1). For the 3 cores,
315 the mean destructive field of ARM and IRM is very similar (between 20 and 30 mT)
316 indicating a magnetic mineralogy mainly composed of low coercivity phase. The S-ratio
317 (Bloemendal et al., 1992) is always between 0,86 and 0,95 indicating lower coercivity and a
318 ferrimagnetic mineralogy. This suggests a good stability of the magnetic mineralogy, except
319 in event layers where other parameters such as the relative palaeointensity (calculated as
320 NRM intensity divided by ARM intensity) are clearly different, highlighting a different

321 magnetic mineralogy. PCA have then been performed using puffin plot software (Lurcock
322 and Wilson, 2012) to calculate the ChRM. A careful examination of demagnetization
323 diagrams shows a unidirectional behaviour (Fig. S2). The Mean Angular Deviation (MAD) is
324 usually lower than 6 revealing a good stability of the magnetization direction. In most cases,
325 the calculated component is not straight to the origin. This is particularly the case in the event
326 layers. This implies the occurrence of a high coercivity component of unknown origin. All
327 cores show quite large variations of the declination and inclination *vs.* depth. Because of the
328 deviation of the Kmin and changes in magnetic mineralogy, measurements from the thickest
329 event layers (i.e. MMIT2 and MMIT3) have been removed to build event-free declination and
330 inclination signals (Fig. 5A). Based on the stratigraphic correlation, the event-free
331 palaeomagnetic profiles obtained for each core were all corrected to a reference depth, i.e. the
332 event-free depth of core FOR13P2 (Fig. 5B). Finally, all magnetic profiles were averaged to
333 obtain unique curves of declination *vs.* depth and inclination *vs.* depth (Fig. 5C), smoothing
334 small artefacts and making it easier for comparison to the reference curve (ARCH3.4k model;
335 Donadini et al., 2009; Korte et al., 2009). From the variations of the reference curve over the
336 last millennium, magnetic declination minima and maxima can be identified at AD 1810 \pm 20,
337 1540 \pm 70 and 1365 \pm 25 (D-1 to D-3, respectively). For the inclination, two tie points at AD
338 1700 \pm 30 and 1330 \pm 40 can be used (I-1 and I-2, Fig. 5D). Furthermore the ChRM
339 declination profile presents 3 declination features and the ChRM inclination profile presents 2
340 inclination features over this period, allowing the correlation proposed (Fig. 5). These well-
341 correlated declination and inclination features can thus be used as additional chronological
342 markers.

343 The ^{210}Pb and the ^{14}C ages (Fig. 4 and Table 1) were then combined with the palaeomagnetic
344 chronomarkers (Fig. 5) to construct an age-depth model covering the whole sequence (Fig. 6).
345 As noted above, the age-depth model was calculated on an event-free depth using a smooth
346 spline with the “clam” R-code package (Blaauw, 2010). This revealed that the sequence FOR
347 covers the last millennium with a mean sedimentation rate of 1 mm.yr⁻¹ (without event
348 layers).

349

350 **5. Discussion**

351

352 **5.1. Different triggers for event layers**

353

354 **5.1.1. Mass movements**

355 The unusual presence of gravel and aquatic plant remains, in combination with the chaotic
356 sedimentary structures and the localized deposition areas, suggest that the coarse-grained
357 layers result of a mass movement originating from sediment-charged slopes (e.g. Sauerbrey et
358 al., 2013). The three graded layers overlying the coarse-grained layers are then resulting from
359 the sediment that is transported in suspension during sliding of subaquatic slope sediments
360 and then deposited over the coarse-grained layers and further into the lake basin (e.g.
361 Girardclos et al., 2007; Moernaut et al., 2014). As a result, each stratigraphic succession of a
362 graded and a coarse-grained layer is interpreted as a mass-movement-induced layer (MMIL).
363 These MMILs are well characterized by higher Ca intensities that suggest a distinctly
364 different sediment source when compared to the sedimentary background and to the 168
365 graded layers that not overlain a coarse-grain layer. The coarse-grained layer of MMIL3 is
366 only present in core FOR13P3, suggesting a littoral origin of the mass movement (Fig. 2). The
367 two coarse-grained layers of MMIL1 and MMIL2 are located in core FOR13P4 (Fig. 2).
368 These layers may thus originate either from the delta or from the littoral slopes. Slope angles
369 of $< 10^\circ$ and $\sim 15^\circ$ for delta and littoral slopes of Lake Foréant, respectively, point to a littoral
370 origin as suggested by many studies showing that slope angles $> 10^\circ$ are favourable for the
371 generation of mass-movements (e.g. Moernaut et al., 2007; Strasser et al., 2011; Van Daele et
372 al., 2013). In addition, higher Ca intensities are often an indicator of littoral sediments as a
373 result of increased fluxes of endogenic calcite when compared to the open-water endogenic
374 production.

375

376 **5.1.2. Flood events**

377 The 168 graded layers may be induced by either mass movements or flood events (e.g. Sturm
378 and Matter, 1978). Their extents over the whole basin with a relatively homogeneous
379 deposition pattern, their frequent occurrence and a different geochemical pattern suggest a
380 distinct origin from that of the MMIL graded layers. The low Ca intensities suggest a minor
381 sediment contribution of the marble and calc-schists in favour of a major contribution of the
382 arkose band, which is the lithology drained by the main inflow (Fig. 1). The 168 graded layers
383 are also characterized by sharp peaks of Mn only located at their bases. This location suggests

384 that the punctual enrichment in Mn is related to the occurrence of these event layers. Mn is a
385 redox sensitive element and more soluble under reducing conditions (e.g. Davison, 1993;
386 Torres et al., 2014). The punctual presence of detectable Mn at the base of the graded layers
387 suggests that hyperpycnal turbidity currents carry oxygen to the deeper parts of the basin.
388 Dissolved oxygen is probably also trapped in pore waters of the individual graded layers.
389 Based on these considerations we suggest that dissolved and reduced Mn is, in part due to the
390 rapid increase in loading from the graded layers, migrating from pore waters of the buried
391 sediments into oxygenated graded layers where it is oxidized and precipitated likely in form
392 of an Mn-oxyhydroxide (e.g. Davison, 1993; Deflandre et al., 2002). The fast sediment
393 deposition during the event-layer formation and the low reactive organic matter
394 concentrations would then prevent reductive dissolution of the Mn-oxyhydrxide precipitates
395 (e.g. Torres et al., 2014). According to these layer characteristics, flood events are the most
396 probable candidate to trigger the 168 graded layers because i) these events may be frequent
397 (e.g. Czymzik et al., 2013), ii) these events may bring both high oxygen and detrital inputs in
398 a short time (e.g. Deflandre et al., 2002), and iii) the nature of the sediment correspond the
399 most to the main lithology drained by the inflow. Hence, the 168 graded layers likely
400 correspond to flood-induced layers (FIL).

401

402 **5.1.3. Chronological controls**

403 MMILs can be triggered by spontaneous failures due to overloading/oversteepening of
404 sediments-charged slopes, snow avalanches, rockfalls, earthquakes or fluctuations in lake
405 levels (e.g. Monecke et al., 2004; Girardclos et al., 2007; Moernaut et al., 2014). In case of
406 Lake Foréant, changes in lake level can be excluded because water levels of Lake Foréant are
407 well controlled by bedrock outlets. In addition, there is no geomorphological evidence of
408 major rockfalls in the catchment area. Regarding earthquakes, many events occurred in the
409 region and affected the population and infrastructure. Historical earthquakes are well
410 documented thanks to the database SisFrance (<http://www.sisfrance.net>, Lambert and Levret-
411 Albaret, 1996; Scotti et al., 2004). An earthquake trigger for the MMILs can then be
412 investigated by comparing ages of the MMILs to the dates of the closest and/or strongest
413 historical earthquakes (e.g. Avşar et al., 2014; Howarth et al., 2014). The three MMILs are
414 respectively dated to AD 1963 (± 6), AD 1814 (+50/-39) and AD 1456 (+19/-56) (Fig. 6). The
415 age of the most recent deposit is consistent with the Saint-Paul-sur-Ubaye earthquake (AD
416 1959), characterized by an epicentre at ca. 20 km from the lake where the MSK intensity

417 reached VII-VIII. The age of the second deposit is consistent with the Piemont (Torre Pellice)
418 earthquake (AD 1808), characterized by an epicentre at ca. 20 km from the lake and a MSK
419 intensity of VIII (Fig. 6). For the older period of the third deposit, data of documented
420 earthquakes are sparser in the catalogue, precluding a reliable assignment. The earthquakes of
421 Saint-Paul-sur-Ubaye and Piemont are both the closest and strongest historically-known
422 earthquakes around the lake, suggesting that they are the most probable trigger of the
423 temporarily corresponding subaquatic landslides. Overall, there is a good agreement between
424 major historical events and the calculated ages of the MMILs supporting their sedimentologic
425 interpretation and the chronology over the last centuries.

426

427 **5.2. Palaeoflood record**

428 A flood chronicle of the Bouchouse stream was built by dating the 168 FILs over the last
429 millennium. Changes in flood frequency are highlighted through a running sum of flood
430 occurrences with an 11-year (Fig. 7) or 31-year window (Fig. 8). The absence of significant
431 grain-size variability precludes the use of grain size to assess changes in flood intensity (e.g.
432 Giguet-Covex et al., 2012; Lapointe et al. 2012; Wilhelm et al., 2013, 2015). The relatively
433 homogeneous grain size of the FILs makes the sediment accumulation per event a more
434 suitable proxy of flood intensity (e.g. Jenny et al., 2014). In addition, the relatively
435 homogeneous flood-sediment deposition pattern within the lake basin makes it possible to use
436 the FIL thickness as a proxy of the flood-sediment accumulation as shown by previous
437 calibration in Alpine environments (Wilhelm et al., 2012b; Jenny et al., 2014; Wilhelm et al.,
438 2015). Hence, the FIL thickness is here assumed to represent the flood intensity, under the
439 condition that erosion processes and availability of erodible materials in the catchment did not
440 change significantly over time.

441

442 **5.2.1. Proxy validation**

443 To control the reliability of our reconstruction, the Foréant palaeoflood record is compared to
444 the historical floods at Ristolas located around 8 km downstream the lake (Fig. 1C and
445 Supplementary Material). The almost absence of documented flood event for the Bouchouse
446 stream (outlet of Lake Foréant) precludes an event-to-event comparison as undertaken with
447 the Lake Allos record (Wilhelm et al., 2015). Hence, the 21 flood events having affected the
448 village of Ristolas and occurring during the ice-free season of the lake (mid-June to mid-

449 November) have been considered to reconstruct a historical flood record (Fig. 7). This
450 includes 6 floods considered as ‘local’ because they only affected the village of Ristolas
451 (catchment area of ca. 80 km²) and 15 floods considered as ‘sub-regional’ because they also
452 affected other villages downstream (Abriès, Aiguilles, Chateau-Vieille-Ville, catchment area
453 of ~320 km²). Through comparison of the historical chronicles and the lake records, we
454 observe that the ranges of flood-frequency values are similar, i.e. between 0 and around 4
455 floods per 11 years. We also observe strong similarities in the two flood records with common
456 periods of low flood frequency in AD 1750-1785, 1820-1860 and 1910-1945 and common
457 periods of high flood frequency in AD 1785-1820, AD 1945-1970 and AD 1985-2000. Only a
458 slight time lag (~5 years) appears for the latter period in the lake record. Overall, there is then
459 good agreement with the historical data, supporting that Lake Foréant sediments record the
460 variability of past flood events that impacted societies over the last 250 years relatively well.
461 A major inconsistency, however, appears from 1860 to 1910 since numerous floods are
462 documented in the lake record but there is missing evidence for flood in the historical record.
463 A high hydrological activity is documented for the region at this time (e.g. Miramont et al.,
464 1998; Sivan et al., 2010; Wilhelm et al., 2012a; 2015), suggesting that this may result of a
465 historical database locally incomplete.

466

467 **5.2.2. Potential influences of environmental changes**

468 The Foréant flood record may be considered as relevant over the entire studied period if
469 erosion processes are stable over time. Erosion processes in the Foréant catchment may be
470 affected by modifications in the river system and/or by land-use changes.

471 The main inflow, the Bouchouse stream, has built an alluvial plain upstream the lake where it
472 is divided in two main meandering branches. An alternate activity of these branches during
473 floods may disturb the flood record by triggering variable sediment dispersion within the lake
474 basin (e.g. Wilhelm et al., 2015). However, such processes seem to be unlikely because the
475 stratigraphic correlation highlights a stable pattern of the flood-sediment deposition with the
476 thickest FILs in cores FOR13P2 and FOR13P4 from the depocenter and a thinning of the FIL
477 deposits toward cores FOR13P1 and FOR13P3 located in the slopes (Fig. 2). The alluvial
478 plain may also disturb the record by acting as a sediment trap. Indeed, the meandering river
479 morphology and the gentle slope of the alluvial plain may trigger a decrease of the discharge
480 velocity, resulting in the deposition of the coarser particles on the plain before entering the

481 lake. This may explain the small variability in grain-size in the Foréant sediment record. The
482 grain-size ratio between the base (coarser fraction) and the top (finer fraction) of the FILs is
483 ~ 1.3 , while it usually ranges from 5 to 15 in many different geological and environmental
484 settings (e.g. Oslegger et al., 2009; Giguet-Covex et al., 2012; Simmoneau et al., 2013;
485 Wilhelm et al., 2013 Amman et al., 2015; Wilhelm et al., 2015). However, fine particles (i.e.
486 clays and fine silts that composed the FILs) are transported by suspension in the river (e.g.
487 Passega, 1964). As a result, their trapping and storage in the alluvial plain is unlikely. A
488 negligible storage effect on the fine fraction is also supported by the relatively stable
489 sedimentation rate of the silty sedimentary background (Fig. 6) that suggests an uninterrupted
490 sediment transport to the lake over the studied period.

491 Erosion processes in the catchment may also be modified by land-use that mainly corresponds
492 at this altitude to changes in grazing intensity. An increase of grazing intensity may make
493 soils more vulnerable to erosion during heavy rainfalls and, thereby, may induce an increased
494 sensitivity of the catchment-lake system to record floods, i.e. higher flood frequency and/or
495 flood-sediment accumulation in the sediment record (e.g. Giguet-Covex et al., 2012).
496 Abundance of *Sporormiella* is assumed to reflect local changes of grazing intensity in Lake
497 Foréant catchment (e.g. Etienne et al., 2013). The concentration of *Sporormiella* ascospores
498 measured in core FOR13P3 oscillated from 5 to 43 number.cm⁻³ through the sequence (Fig.
499 2), resulting in accumulation rates varying from 12 to 340 number.cm².yr⁻¹ over time (Fig. 8).
500 This variability in *Sporormiella* abundance has been compared to the variability in flood
501 frequency and flood-sediment accumulation (see Supplementary Material). We do not find
502 significant relationships ($p > 0.05$) between these parameters (Fig. S3), suggesting that
503 variations in pastoralism seemingly have not had a significant impact on erosion processes in
504 the Foréant catchment. However, two samples covering the period AD 1734-1760 show both
505 high *Sporormiella* accumulation rates and flood frequencies (Fig. 8 and S3). This suggests
506 that the flood frequency during this period may be exacerbated by a punctual and very high
507 grazing intensity. Hence, we postulate that erosion processes did not change drastically over
508 the studied period, implying that climate is likely the main factor explaining the recorded
509 flood activity, with exception of the period AD 1734-1760.

510

511 **5.2.3. Palaeoflood activity in the regional climatic setting**

512 Comparison with the historical record shows that the past flood variability is well reproduced
513 by the Foréant record (Fig. 7). The Foréant palaeoflood record is thus interpreted as the
514 recurrence of summer-to-autumn flood events triggered by both local and mesoscale
515 convective phenomena.

516 To discuss the millennium-long flood variability in regard to both Atlantic and Mediterranean
517 climatic influences in the Alpine domain, the Foréant palaeoflood record is compared to the
518 palaeoflood records of Lakes Blanc and Allos (Fig. 7 and 8). Lakes Blanc and Allos have
519 similar characteristics to Lake Foréant such as the high altitude (> 2000 m a.s.l.), the small
520 catchment area (< 3 km²) and the steep catchment slopes, making possible the comparison.
521 Lake Blanc sediments located in the northern French Alps mainly record Atlantic-sourced
522 weather pattern of high altitude, i.e. summer local convective events (Fig. 1; Giguet-Covex et
523 al., 2012; Wilhelm et al., 2012b, 2013). In contrast, Lake Allos sediments located in the
524 southern French Alps mainly record Mediterranean-sourced weather patterns of high altitude,
525 i.e. mesoscale convective events (Fig. 1; Wilhelm et al., 2012a, 2015). The last millennium is
526 usually divided in three climatic periods according to the temperature variations; the warm
527 Medieval Climate Anomaly (MCA, ca. AD 950-1250), the cold Little Ice Age (LIA, ca. AD
528 1300-1900) and the warmer 20th century (e.g. Lamb, 1965; Büntgen et al., 2011; Luterbacher
529 et al., 2012 and references therein). During the MCA, the Foréant flood record shows a low
530 flood frequency with ~ 10 floods per century and, 4 occurrences of thick flood deposits (> 8
531 mm thick) that we interpret as high-intensity flood events. During the LIA, the Foréant record
532 shows a higher flood frequency with ~ 17 floods per century and only 2 high-intensity events.
533 The 20th century is finally characterized by ~ 17 floods per century and absence of high-
534 intensity events. The increased flood frequency during the long and cold period of the LIA,
535 compared to the MCA, was also observed in the Blanc and Allos records (Wilhelm et al.,
536 2012a; 2013; Fig. 8), as well as in many other records from the European Alps (e.g. Arnaud et
537 al., 2012; Glur et al., 2013; Swierczynski et al., 2013; Wirth et al., 2013a, 2013b; Amann et
538 al., 2015; Schulte et al., 2015) and the north-western Mediterranean area (e.g. Jorda and
539 Provansal, 1996; Camuffo and Enzi, 1995; Jorda et al., 2002; Thorndycraft and Benito, 2006;
540 Moreno et al., 2008; Benito et al., 2015; Arnaud-Fassetta et al., 2010). This common pattern
541 in many flood records of southern Europe may be the result of a southward shift and an
542 intensification of the dominant westerly winds during boreal summer related to an increase in
543 the thermal gradient between low (warming) and high (cooling) latitudes (e.g. Bengtsson and

544 Hodges, 2006; Raible et al., 2007). In this scenario, the Alps are likely to experience an
545 increase in precipitation due to an increase in moisture advection from the North Atlantic. In
546 contrary, the occurrence of high-intense floods during both the MCA and the LIA periods is a
547 new feature of Alpine regional patterns, since the most intense floods occurred exclusively
548 during the MCA in the Blanc record (Wilhelm et al., 2013) or during the LIA in the Allos
549 record (Wilhelm et al., 2012a; Fig. 8) and other Mediterranean records of the Alpine region
550 (Jorda and Provansal, 1996; Miramont et al., 1998; Jorda et al., 2002; Arnaud-Fassetta et al.,
551 2010). This suggests that hydro-meteorological processes related to the Atlantic and to the
552 Mediterranean climatic influences may alternatively trigger high intense events in the Foréant
553 area during the MCA and the LIA, respectively. However, the most intense floods at Foréant
554 appear 3 times more frequent during the MCA than during the LIA, a trend that remains true
555 when considering various thickness thresholds (8, 7, 6 or 5 mm) for high-intensity flood
556 events. In addition, the mean sediment accumulation per flood event shows values ~50%
557 higher during the MCA than during the LIA (3.8 vs. 2.4 mm/flood), suggesting an increase of
558 the mean flood-event intensity during the warmer period. These two evidences of increased
559 flood intensity during the warm period may be related to the strengthening of local convective
560 processes due to higher temperatures, as suggested for the north-western flood pattern
561 (Giguet-Covex et al., 2012; Wilhelm et al., 2012b, 2013). In the Foréant area, higher
562 temperatures seem thus to result in a lower flood frequency but in higher flood intensity on
563 the multi-centennial time scale. Flood frequency and intensity during the warmer 20th century,
564 however, do not follow these trends. The frequency is still similar to the LIA one and high-
565 intense events are absent and the mean sediment accumulation per flood event (2.2 mm/flood)
566 is also similar to the LIA. Two hypotheses may be considered to explain this ‘anomaly’. First,
567 this may result from the relatively short period covered by the 20th century (i.e. ~100 years) in
568 comparison with the multi-centennial variability documented for the MCA (i.e. ~300 years)
569 and the LIA (i.e. ~600 years) periods. Thereby, considering stable temperature-flood
570 relationships over time, the 20th century might be a transitional period toward a MCA-like
571 flood pattern with the global warming. This latter possibility would imply an increasing flood
572 hazard in the Foréant region in a near future due to an increased occurrence of high-intensity
573 flood events. Secondly, this may also result from a non-linearity of the flood response to
574 temperature, making the analogy between the MCA and the 20th century more complex, in
575 particular as the current warming is caused by an unprecedented forcing (greenhouse gases).
576 Moreover, the other external forcing such as solar activity, and volcanic eruptions largely
577 varied over the last millennium (e.g. Servonnat et al., 2010; Delaygue et Bard, 2011; Gao et

578 al., 2012; Crowley and Unterman, 2013) and their non-linear combination also with the
579 greenhouse gases may result in different time-space temperature patterns and, thereby, in
580 different flood responses during these two periods. In order to explore forcing-dependent
581 impacts on the climate-flood relationships, deeper analysis utilizing for example advanced
582 statistics or simulations is required.

583

584 **6. Conclusion**

585 High-resolution sedimentological and geochemical analyses of the Lake Foréant sequence
586 revealed 171 event layers. Three of the 171 event layers can be differentiated by characteristic
587 geochemical features (high Ca contents and low Mn contents) and stratigraphic succession.
588 These three event layers are interpreted as mass-movement-induced layers. The other 168
589 event layers show a geochemical pattern similar to the sedimentary background that mainly
590 corresponds to detrital material sourced by the rivers. These event layers are interpreted as
591 flood-induced layers. Only small changes in grain-size variability in the flood-induced layers
592 precluded the use of the grain size as a flood-intensity proxy. However, the relatively
593 homogeneous grain size and deposition pattern within the lake basin made the flood-deposit
594 thickness a suitable proxy for the reconstruction of flood intensity.

595 Comparison with local historical data indicates that Foréant sediments sensitively record past
596 flood events with variability in frequency and intensity related to both Atlantic- and
597 Mediterranean-influenced hydro-meteorological processes, i.e. local and mesoscale
598 convective systems occurring from late spring to fall. As there is no evidence of major
599 changes in erosion processes due to landscape evolution or grazing intensity (except maybe
600 for the period AD 1734-1760), we assume that climate and not land-use changes exerts the
601 dominant control on flood variability in the Foréant-record over the past millennium. The
602 comparison to northern and southern flood records, i.e. to Atlantic- and Mediterranean-
603 influenced records, highlights that the increase of flood frequency during the cold period of
604 the LIA is a common feature of all regional flood patterns from the European Alps. The
605 comparison also revealed that high-intensity events in the Foréant region occurred during both
606 the cold LIA and the warm MCA periods. This specific feature of the Foréant flood record
607 likely results from its sensitivity to both Atlantic and Mediterranean climatic influences.
608 However, high-intensity events are more frequent and the flood intensity is higher during the
609 warm MCA. This suggests that flood hazard may increase in the Foréant region in response to

610 global warming. Surprisingly, the flood variability over the warm 20th century appears still
611 similar to the flood variability of the cold LIA period. This 20th-century flood trend may be
612 interpreted as the result of a transitional period toward a MCA-like flood pattern. This would
613 imply an increasing flood hazard in the Foréant region in a near future due to more frequent
614 high-intensity flood events. However, this may also result from a non-linear temperature-
615 flood relationship. In order to better understand the underlying mechanisms deeper analyses
616 employing advanced statistics or simulations need to be applied.

617

618 **Acknowledgments**

619 B. Wilhelm's post-doctoral fellowship (2013-2014) was supported by a grant from the AXA
620 Research Fund. We would also like to thank Pierre Sabatier for his help with analysis of the
621 ²¹⁰Pb data, Anne-Lise Develle for fruitful discussions, Petra Boltshauser-Kaltenrieder for help
622 with identification of terrestrial-plant remains prior to ¹⁴C dating and Nicolas Thouveny and
623 François Demory for providing access to facilities at the CEREGE palaeomagnetic laboratory
624 (Aix-Marseille University) and for fruitful discussions.

625

626

627

628 **References**

629 Amann, B., Sönke, S., Grosjean, M.: A millennial-long record of warm season precipitation
630 and flood frequency for the North-western Alps inferred from varved lake sediments:
631 implications for the future, *Quaternary Science Reviews*, 115, 89-100, 2015.

632 Appleby, P.G., Richardson, N., Nolan, P.J.: ²⁴¹Am dating of lake sediments, *Hydrobiologia*,
633 214, 35–42, 1991.

634 Arnaud, F., Lignier, V., Revel, M., Desmet, M., Pourchet, M., Beck, C., Charlet, F.,
635 Trentesaux, A., Tribovillard, N.: Flood and earthquake disturbance of ²¹⁰Pb geochronology
636 (Lake Anterne, North French Alps), *Terra Nova*, 14, 225–232, 2002.

637 Arnaud, F., Révillon, S., Debret, M., Revel, M., Chapron, E., Jacob, J., Giguex-Covex, C.,
638 Poulencard, J., Magny, M.: Lake Bourget regional erosion patterns reconstruction reveals

639 Holocene NW European Alps soil evolution and paleohydrology, *Quaternary Science*
640 *Reviews*, 51, 81–92, 2012.

641 Arnaud-Fassetta, G., and Fort, M.: Respective parts of hydroclimatic and anthropic factors in
642 the recent evolution (1956-2000) of the active channel of the Upper Guil, Queyras, Southern
643 French Alps, *Méditerranée*, 102, 143-156, 2004.

644 Arnaud-Fassetta, G., Carcaud, N., Castanet, C. and Salvador, P.G.: Fluvial
645 palaeoenvironments in archaeological context: geographical position, methodological
646 approach and global change—hydrological risk issues, *Quat. Int.*, 216, 93–117, 2010.

647 Avşar, U., Hubert-Ferrari, A., De Batist, M., Lepoint, G., Schmidt S. and Fagel N.:
648 Seismically-triggered organic-rich layers in recent sediments from Göllüköy Lake (North
649 Anatolian Fault, Turkey), *Quat. Sci. Rev.*, 103, 67–80, 2014.

650 Barletta, F., St-Onge, G., Channell, J.E.T. and Rochon, A.: Dating of Holocene western
651 Canadian Arctic sediments by matching paleomagnetic secular variation to a geomagnetic
652 field model, *Quaternary Science Reviews*, 29, 2315–2324, 2010.

653 Bengtsson, L., and Hodges, K.I.: Storm tracks and climate change, *Journal of Climate*, 19,
654 3518–3543, 2006.

655 Beniston, M. and Stephenson, D.B.: Extreme climatic events and their evolution under
656 changing climatic conditions, *Glob Planet Change*, 44, 1–9, 2004.

657 Benito, G., Macklin, M.G., Zielhofer, C., Jones A., and Machado, M.J.: Holocene flooding
658 and climate change in the Mediterranean, *Catena*, 130, 13-33, 2015.

659 Blaauw, M.: Methods and code for ‘classical’ age modelling of radiocarbon sequences, *Quat.*
660 *Geochronol.*, 5, 512–518, 2010.

661 Bloemendal, J., King, J.W., Hall, F.R. and Doh, S.-J.: Rock magnetism of Late Neogene and
662 Pleistocene deep-sea sediments relationship to sediment source, diagenetic processes, and
663 sediment lithology, *Journal of Geophysical Research*, 97: 4361-4375, 1992.

664 Boudevillain, B., Argence, S., Claud, C., Ducrocq, V., Joly, B., Lambert, D., Nuissier, O.,
665 Plu, M., Ricard, D., Arbogast, P., Berne, A., Chaboureaud, J.P., Chapon, B., Crepin, F.,
666 Delrieu, G., Doerflinger, E., Funatsu, B.M., Kirstetter, P.E., Masson, F., Maynard, K.,
667 Richard, E., Sanchez, E., Terray, L. and Walpersdorf, A.: Cyclogenesis et précipitations
668 intenses en région méditerranéenne: origines et caractéristiques, *La Meteorol.*, 66, 18–28,
669 2009.

670 Boroneant, C., Plaut, G., Giorgi, F. and Bi., X.: Extreme precipitation over the Maritime Alps
671 and associated weather regimes simulated by a regional climate model: Present-day and future
672 climate scenarios, *Theor. Appl. Climatol.*, 86, 81–99, 2006.

673 Büntgen, U., Tegel, W., Nicolussi, K., McCormick, M., Frank, D., Trouet, V., Kaplan, J.,
674 Herzig, F., Heussner, U., Wanner, H., Luterbacher, J., Esper, J. : 2500 years of European
675 climate variability and human susceptibility, *Science*, 331, 578-582, 2011

676 Buzzi, A. and Foschini, L.: Mesoscale meteorological features associated with heavy
677 precipitation in the Southern Alpine Region, *Meteorol. Atmos. Phys.*, 72, 131–146, 2000.

678 Campbell, C.: Late Holocene lake sedimentology and climate change in southern Alberta,
679 Canada, *Quatern. Res.*, 49, 96–101, 1998.

680 Camuffo, D. and Enzi, S.: The analysis of two bi-millenary series: Tiber and Po River Floods,
681 In: Jones, P.D. Bradley, R.S. Jouzel, J. (Eds.), *Climatic Variations and Forcing Mechanisms*
682 of the Last 2000 years, NATO ASI Series, Series I: Global Environmental Change, 41
683 Springer, Stuttgart, 433–450, 1995.

684 Corella, J.P., Benito, G., Rodriguez-Lloveras, X., Brauer, A. and Valero-Garcès, B.L.:
685 Annually-resolved lake record of extreme hydro-meteorological events since AD 1347 in NE
686 Iberian Peninsula, *Quat. Sci. Rev.*, 93, 77-90, 2014.

687 Croudace, I.W., Rindby, A. and Rothwell, R.G.: ITRAX: description and evaluation of a new
688 multi-function X-ray core scanner. In: Rothwell, R.G. (Ed.), *New Techniques in Sediment*
689 *Core Analysis*, Geological Society, London, Special Publications, 367, 51-63, 2006.

690 Crowley, T.J. and Untermann, M.B.: Technical details concerning development of a 1200-yr
691 proxy index for global volcanism, *Earth Syst. Sci. Data*, 5, 187-197, 2013.

692 Cuvén, S., Francus, P. and Lamoureux, S.: Estimation of grain size variability with micro X-
693 ray fluorescence in laminated lacustrine sediments, Cape Bounty, Canadian High Arctic,
694 *Journal of Paleolimnology*, 44, 803–817, 2010.

695 Czymzik, M., Brauer, A., Dulski, P., Plessen, B., Naumann, R., von Grafenstein, U. and
696 Scheffler, R.: Orbital and solar forcing of shifts in Mid- to Late Holocene flood intensity from
697 varved sediments of pre-alpine Lake Ammersee (southern Germany), *Quatern. Sci. Rev.*, 61,
698 96–110, 2013.

699 Davis, O.K. and Schafer, D.: *Sporormiella* fungal spores, a palynological means of detecting
700 herbivore density. *Palaeogeol., Palaeoclimatol., Palaeoecol.*, 237, 40–50, 2006.

701 Davison, W.: Iron and manganese in lakes, *Earth-Science Reviews*, 34, 119-163,1993

702 Deflandre, B., Mucci, A., Gagne, J.P., Guignard, C., Sundby, B.: Early diagenetic processes
703 in coastal marine sediments disturbed by a catastrophic sedimentation event, *Geochimica et*
704 *Cosmochimica Acta*, 66(14), 2547–2558, 2002.

705 Delaygue, G. and Bard, E.: An Antarctic view of Beryllium-10 and solar activity for the past
706 millennium, *Climate Dynamics*, 36, 2201–2218, 2011.

707 Donadini, F., M. Korte, Constable C. G.: Geomagnetic field for 0–3 ka: 1. New data sets for
708 global modeling, *Geochemistry Geophysics Geosystems (G3)*, 10, Q06007,
709 doi:10.1029/2008GC002295, 2009.

710 Durant, Y., Laternser, M., Giraud, G., Etchevers, P., Lesaffre, B., Merindol, L.: Reanalysis of
711 44 Yr of Climate in the French Alps (1958–2002): Methodology, Model Validation,
712 Climatology, and Trends for Air Temperature and Precipitation, *J. Applied Meteorology and*
713 *Climatology*, 48(3):429-449, 2009.

714 Etienne, D., Wilhelm, B., Sabatier, P., Reyss, J.L. and Arnaud, F.: Influence of sample
715 location and livestock numbers on *Sporormiella* concentrations and accumulation rates in
716 surface sediments of Lake Allos, French Alps, *J. Paleolimnol.*, 49, 117–127, 2013.

717 Etienne, D. and Jouffroy-Bapicot, I.: Optimal counting limit for fungal spore abundance
718 estimation using *Sporormiella* as a case study. *Veget. Hist. Archaeobot*, 23, 743-749, 2014.

719 Faegri, K. and Iversen, J.: *Textbook of Pollen Analysis*. John Wiley & Sons, New York. 328,
720 1989.

721 Frei, C., Schöll, R., Fukutome, S., Schmidli, J., Vidale, P.L.: Future change of precipitation
722 extremes in Europe: Intercomparison of scenarios from regional climate models, *J. Geophys.*
723 *Res. Atm.*, 111, D06105. doi:10.1029/2005JD005965, 2006.

724 Gao, C., Robock, A. and Ammann, C.: Volcanic forcing of climate over the past 1500 years:
725 An improved ice core-based index for climate models, *J. Geophys. Res.*, 113, D23111, 2008.

726 Giguet-Covex, C., Arnaud, F., Enters, D., Poulénard, J., Millet, L., Francus, P., David, F.,
727 Rey, P.J., Wilhelm, B. and Delannoy, J.J.: Frequency and intensity of high altitude floods
728 over the last 3.5 ka in NW European Alps, *Quatern. Res.*, 77, 12–22, 2012.

729 Gilli, A., Anselmetti, F.S., Glur, L. and Wirth, S.B.: Lake sediments as archives of recurrence
730 rates and intensities of past flood events, In: *Dating Torrential Processes on Fans and Cones –*

731 Methods and Their Application for Hazard and Risk Assessment (Eds M. Schneuwly-
732 Bollschweiler, M. Stoffel and F. Rudolf-Miklau), *Adv. Global Change Res.* 47, 225–242,
733 2012.

734 Girardclos, S., Schmidt, O.T., Sturm, M., Ariztegui, D., Pugin A., and Anselmetti F.S.: The
735 1996 AD delta collapse and large turbidite in Lake Brienz, *Marine Geology*, 24, 137–154,
736 2007.

737 Glur, L., Wirth, S.B., B€untgen, U., Gilli, A., Haug, G.H., Schär, C., Beer, J. and Anselmetti,
738 F.S., Frequent floods in the European Alps coincide with cooler periods of the past 2500
739 years, *Sci. Rep.*, 3, 2770, 2013.

740 Goldberg, E.D.: *Geochronology with 210Pb Radioactive Dating*, IAEA, Vienna, 121–131.
741 1963

742 Gottardi, F., Obled, C., Gailhard, J., Paquet, E.: Statistical reanalysis of precipitation fields
743 based on ground network data and weather patterns: Application over French mountains, *J.*
744 *Hydrology*, 432–433, 154–167, 2010.

745 Howarth, J.D., Fitzsimons, S.J., Norris, R.J. and Jacobsen, G.E.: Lake sediments record high
746 intensity shaking that provides insight into the location and rupture length of large
747 earthquakes on the Alpine Fault, New Zealand, *Earth and Planetary Science Letters*, 403,
748 340–351, 2014.

749 IPCC: *The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment*
750 *Report of the Intergovernmental Panel on Climate Change* (Eds T.F. Stocker, D. Qin, G.-K.
751 Plattner, M. Tignor, S.K. Allen, J. Boschung, A. Nauels, Y. Xia, V. Bex and P.M. Midgley).
752 Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, 2013

753 Jorda, M. and Provansal, M.: Impact de l’anthropisation et du climat sur le detritisme en
754 France du sud-est (Alpes du Sud et Provence), *Bulletin de la Societe Geologique de France*,
755 167, 159–168, 1996.

756 Jorda, M., Miramont, C., Rosique, T. and Sivan, O.: Evolution de l’hydrosysteme durancien
757 (Alpes du Sud, France) depuis la fin du Pleniglaciaire superieur, In: Bravard, J.-P., Magny, M.
758 (Eds.), *Les Fleuves ont une Histoire, Paleo-environnement des Rivieres et des lacs Francais*
759 *depuis 15000 ans. Errance, Paris*, 239–249, 2002.

760 Jenny, J.-P., Wilhelm, B., Arnaud, F., Sabatier, P., Giguet-Covex, C., Mélo, A., Fanget, B.,
761 Malet, E., Ployon, E. and Perga, M.E., A 4D sedimentological approach to reconstructing the

762 flood frequency and intensity of the Rhône River (Lake Bourget, NW European Alps), *J.*
763 *Paleolimnol.*, 51, 469–483, 2014.

764 Korte, M., Donadini, F., Constable, C.G. : Geomagnetic field for 0–3 ka: 2. A new series of
765 time-varying global models, *Geochemistry Geophysics Geosystems (G3)*, 10, Q06008, doi:
766 10.1029/2008GC002297, 2009.

767 Lamb, H. H.: The early medieval warm epoch and its sequel, *Palaeogeogr. Palaeoclimatol.*
768 *Palaeoecol.*, 1, 13–37, 1965.

769 Lambert, J, and Levret-Albaret, A.: Mille ans de séismes en France. Catalogues d'épicentres,
770 paramètres et references, Ouest-Editions, Presses Académiques, Nantes. 1996.

771 Lapointe, F., Francus, P., Lamoureaux, S.F., Saïd, M. and Cuven, S.: 1,750 years of large
772 rainfall events inferred from particle size at East Lake, Cape Bounty, Melville Island, Canada,
773 *Journal of Paleolimnology*, 48(1), 159–173, 2012.

774 Lionello, P., Abrantes, F., Congedi, L., Dulac, F., Gacic, M., Gomis, D., Goodess, C., Hoff,
775 H., Kutiel, H., Luterbacher, J., Planton, S., Reale, M., Schröder, K., Struglia, M.V., Toreti, A.,
776 Tsimplis, M., Ulbrich, U. and Xoplaki, E.: Introduction: Mediterranean Climate: Background
777 Information in Lionello P. (Ed.) *The Climate of the Mediterranean Region. From the Past to*
778 *the Future*, Amsterdam: Elsevier (Netherlands), XXXV-XXXX, 2012.

779 Lurcock, P. C. and Wilson, G. S.: PuffinPlot: A versatile, user-friendly program for
780 paleomagnetic analysis, *Geochemistry, Geophysics, Geosystems*, 13, Q06Z45, 2012.

781 Luterbacher, J., García-Herrera, R., Akcer-On, S., Allan, R., Alvarez-Castro, M. C., Benito,
782 G., Booth, J., Büntgen, U., Cagatay, N., Colombaroli, D., Davis, B., Esper, J., Felis, T.,
783 Fleitmann, D., Frank, D., Gallego, D., Garcia-Bustamante, E., Glaser, R., González-Rouco,
784 J.F., Goosse, H., Kiefer, T., Macklin, M.G., Manning, S., Montagna, P., Newman, L., Power,
785 M.J., Rath, V., Ribera, P., Riemann, D., Roberts, N., Sicre, M., Silenzi, S., Tinner, W.,
786 Valero-Garces, B., van der Schrier, G., Tzedakis, C., Vannièrè, B., Vogt, S., Wanner, H.,
787 Werner, J.P., Willett, G., Williams, M.H., Xoplaki, E., Zerefos, C.S. and Zorita, E.: A review
788 of 2000 years of paleoclimatic evidence in the Mediterranean, In: Lionello, P. (Ed.), *The*
789 *Climate of the Mediterranean region: from the past to the future*, Elsevier, Amsterdam, The
790 Netherlands, 87-185, 2012.

791 Miramont, C., Jorda, M. and Pichard, G.: Évolution historique de la morphogenèse et de la
792 dynamique fluviale d'une rivière méditerranéenne : l'exemple de la moyenne Durance
793 (France du sud-est), *Géographie physique et Quaternaire*, 52(3), 381-392, 1998.

794 Moernaut, J., M. De Batist, F. Charlet, K. Heirman, E. Chapron, M. Pino, R. Brümmer and R.
795 Urrutia, Giant earthquakes in South-Central Chile revealed by Holocene mass-wasting events
796 in Lake Puyehue, *Sediment. Geol.*, 195, 239–256, 2007.

797 Moernaut, J., Van Daele, M., Heirman, K., Fontijn, K., Strasser, M., Pino, M., Urrutia R., and
798 De Batist, M.: Lacustrine turbidites as a tool for quantitative earthquake reconstruction: New
799 evidence for a variable rupture mode in south central Chile, *J. Geophys. Res. Solid Earth*,
800 119, doi:10.1002/2013JB010738, 2014.

801 Monecke, K., Anselmetti, F.S., Becker, A., Sturm, M. and Giardini, D.: The record of historic
802 earthquakes in lake sediments of Central Switzerland, *Tectonophysics*, 394, 21–40, 2004.

803 Munich Re Group: Annual Review: Natural Catastrophes 2002, Munich Re Group, Munich,
804 Germany, 2003.

805 Moreno, A., Valero-Garces, B.L., Gonzalez-Samperiz, P. and Rico, M.: Flood response to
806 rainfall variability during the last 2000 years inferred from the Taravilla Lake record (Central
807 Iberian Range, Spain), *J. Palaeolimnol.*, 40, 943–961, 2008.

808 Noren, A.J., Bierman, P.R., Steig, E.J., Lini, A. and Southon, J.: Millennial-scale storminess
809 variability in the northeastern United States during the Holocene epoch, *Nature*, 419, 821-824,
810 2002.

811 Osleger, D.A., Heyvaert, A.C., Stoner, J.S. and Verosub, K.L.: Lacustrine turbidites as
812 indicators of Holocene storminess and climate: Lake Tahoe, California and Nevada, *J*
813 *Paleolimnol.*, 42:103–122, 2009.

814 Page, M.J., Trustrum, N.A. and DeRose, R.C.: A high resolution record of storm-induced
815 erosion from lake sediments, New Zealand, *Journal of Paleolimnology*, 11, 333-348, 1994.

816 Parajka, J., Kohnová, S., Bálint, G., Barbuc, M., Borga, M., Claps, P., Cheval, S., Dumitrescu,
817 A., Gaume, E., Hlavčová, K., Merz, R., Pfaundler, M., Stancalie, G., Szolgay, J., Blöschl, G.:
818 Seasonal characteristics of flood regimes across the Alpine–Carpathian range, *J. of*
819 *Hydrology*, 394,78–89, 2010.

820 Passega, R.: Grain-size representation by CM patterns as a geological tool, *J. of Sedimentary*
821 *Petrology*, 34(4), 830–847, 1964.

822 Plaut, G., Schuepbach, E. and Doctor, M. Heavy precipitation events over a few Alpine sub-
823 regions and the links with large-scale circulation, 1971–1995, *Climate Research*, 17, 285–302,
824 2009.

825 Raible, C.C., Yoshimori, M., Stocker, T.F. and Casty, C.: Extreme midlatitude cyclones and
826 their implications for precipitation and wind speed extremes in simulations of the Maunder
827 Minimum versus present day conditions. *Clim. Dyn.* 28, 409–423, 2007.

828 Rajczak, J., Pall, P. and Schär, C.: Projections of extreme precipitation events in regional
829 climate simulations for Europe and the Alpine Region, *J. Geophys. Res. Atm.*, 118, 1–17,
830 2013.

831 Reimer, P.J., Bard, E., Bayliss, A., Beck, J.W., Blackwell, P.G., Bronk Ramsey, C., Buck,
832 C.E., Cheng, H., Edwards, R.L., Friedrich, M., Grootes, P.M., Guilderson, T.P., Haflidason,
833 H., Hajdas, I., Hatt, C., Heaton, T.J., Hogg, A.G., Hughen, K.A., Kaiser, K.F., Kromer, B.,
834 Manning, S.W., Niu, M., Reimer, R.W., Richards, D.A., Scott, E.M., Southon, J.R., Turney,
835 C.S.M. and van der Plicht, J.: IntCal13 and MARINE13 radiocarbon age calibration curves 0-
836 50000 years calBP, *Radiocarbon*, 55(4), 1869–1887, 2013.

837 Sauerbrey, M. A., Juschus, O., Gebhardt, A. C., Wennrich, V., Nowaczyk, N. R. and Melles,
838 M.: Mass movement deposits in the 3.6Ma sediment record of Lake Elgygytgyn, Far East
839 Russian Arctic, *Climate of the Past* 9, 1949–1967, 2013.

840 Schiefer, E., Gilbert, R. and Hassan, M.A.: A lake sediment-based proxy of floods in the
841 Rocky Mountain Front Ranges, Canada, *Journal of Paleolimnology*, 45, 137–149, 2011.

842 Schulte, L., Peña, J.C., Carvalho, F., Schmidt, T., Julià, R., Llorca, J. and Veit, H.: A 2600-
843 year history of floods in the Bernese Alps, Switzerland: frequencies, mechanisms and climate
844 forcing, *Hydrol. Earth Syst. Sci.*, 19, 3047–3072, 2015.

845 Schwartz, S., Tricart, P., Lardeaux, J.M., Guillot, S. and Vidal, O.: Late tectonic and
846 metamorphic evolution of the Piedmont accretionary wedge(Queyras Schistes lustrés, western
847 Alps): Evidences for tilting during Alpine collision, *Geol. Soc. Ame. Bull.*, 121, 502-518,
848 2009.

849 Scotti, O., Baumont, D., Quenet, G. and Levret, A.: The French macroseismic database
850 SISFRANCE: objectives, results and perspectives, *Annals of Geophysics*, 47(2), 571-581,
851 2004.

852 Servonnat, J., Yiou, P., Khodri, M., Swingedouw, D., Denvil, S.: Influence of solar
853 variability, CO₂ and orbital forcing during the last millennium in the IPSLCM4 model, *Clim.*
854 *Past*, 6, 445-460, 2010.

855 Simonneau, A., Chapron, E., Vannière, B., Wirth, S.B., Gilli, A., Giovanni, C. Di.,
856 Anselmetti, F. S., Desmet, M. and Magny M.: Mass-movement and flood-induced deposits in
857 Lake Ledro, southern Alps, Italy: implications for Holocene palaeohydrology and natural
858 hazards, *Clim. Past* 9, 825-840, 2013.

859 Sivan, O., Miramont, C., Pichard, G. and Prosper-Laget, V. : Les conditions climatiques de la
860 torrentialité au cours du Petit Age Glaciaire de Provence, *Archeologie du Midi Medieval*, 26,
861 157–168, 2009.

862 Stockmarr, J.: Tablets with spores used in absolute pollen analysis. *Pollen Spores* 13, 615–
863 621, 1971.

864 Støren, E.N., Olaf Dahl, S., Nesje, A. and Paasche Ø.: Identifying the sedimentary imprint of
865 high-frequency Holocene river floods in lake sediments: development and application of a
866 new method, *Quat. Sci. Rev.*, 29, 3021-3033, 2010.

867 Strasser, M., Hilbe, M. and Anselmetti, F.S.: Mapping basin-wide subaquatic slope failure
868 susceptibility as a tool to assess regional seismic and tsunami hazards, *Mar. Geophys. Res.*,
869 32, 331–347, 2011.

870 Sturm, M. and Matter, A.: Turbidites and varves in Lake Brienz (Switzerland): deposition of
871 clastic detritus by density currents, *Int. Assoc. Sedimentol. Spec. Publ.*, 2, 147–168, 1978.

872 Swierczynski, T., Lauterbach, S., Dulski, P., Delgado, J., Merz, B. and Brauer, A.: Mid- to
873 late Holocene flood frequency changes in the northeastern Alps as recorded in varved
874 sediments of Lake Mondsee (Upper Austria), *Quat. Sci. Rev.*, 80, 78–90, 2013.

875 Tachikawa, K., Cartapanis, O., Vidal, L., Beaufort, L., Barlyaeva, T., Bard, E.: The
876 precession phase of hydrological variability in the Western Pacific Warm Pool during the past
877 400 ka, *Quat. Sci. Rev.*, 30, 3716–3727, 2011.

878 Thorndycraft, V.R. and Benito, G.: Late Holocene chronology of Spain: the role of climatic
879 variability and human impact, *Catena*, 66, 34–41, 2006.

880 Torres, N.T., Och, L.M., Hauser, P.C., Furrer, G., Brandl, H., Vologina, E., Sturm, M., and
881 Bürgmann, H. z Müller, B.: Early diagenetic processes generate iron and manganese oxide

882 layers in the sediments of Lake Baikal, Siberia, *Env. Sci.: Processes Impacts*, 16, 879–889,
883 2014.

884 Trigo, I.F. and Davies, T.D.: Decline in Mediterranean rainfall caused by weakening of
885 Mediterranean cyclones, *Geophysical Research Letters*, 27(18), 2913–2916, 2000.

886 Van Daele, M., Versteeg, W., Pino, M., Urrutia, R. and De Batist, M.: Widespread
887 deformation of basin-plain sediments in Aysén fjord [Chile] due to impact by earthquake-
888 triggered, onshore-generated mass movements, *Marine Geol.*, 337, 67–79, 2013.

889 Van Daele, M., Moernaut, J., Doom, L., Boes, E., Fontijn, K., Heirman, K., Vandoorne, W.,
890 Hebbeln, D., Pino, M., Urrutia, R., Brümmer, R. and De Batist, M.: A comparison of the
891 sedimentary records of the 1960 and 2010 great Chilean earthquakes in 17 lakes: Implications
892 for quantitative lacustrine palaeoseismology, *Sedimentology*, 62(5): 1466–1496, 2015.

893 Van Geel, B. and Aptroot, A.: Fossil ascomycetes in Quaternary deposits. *Nova Hedwig* 82,
894 313–329, 2006.

895 Van Geel, B., Buurman, J., Brinkkemper, O., Schelvis, J., Aptroot, A., van Reenen, G. and
896 Hakjbilj, T.: Environmental reconstruction of a Roman period settlement site in Uitgeest (The
897 Netherlands), with special reference to coprophilous fungi. *J. Archaeol. Sci.* 30, 873–883,
898 2003.

899 Wilhelm, B., Arnaud, F., Sabatier, P., Crouzet, C., Brisset, E., Chaumillon, E., Disnar, J.R.,
900 Guiter, F., Malet, E., Reyss, J.L., Tachikawa, K., Bard, E. and Delannoy, J.J.: 1400 years of
901 extreme precipitation patterns over the Mediterranean French Alps and possible forcing
902 mechanisms, *Quat. Res.*, 78, 1–12, 2012a.

903 Wilhelm, B., Arnaud, F., Enters, D., Allignol, F., Legaz, A., Magand, O., Revillon, S.,
904 Giguet-Covex, C. and Malet, E.: Does global warming favour the occurrence of extreme
905 floods in European Alps? First evidences from a NW Alps proglacial lake sediment record,
906 *Clim. Change.*, 113, 563–581, 2012b.

907 Wilhelm, B., Arnaud, F., Sabatier, P., Magand, O., Chapron, E., Courp, T., Tachikawa, K.,
908 Fanget, B., Malet, E., Pignol, C., Bard, E. and Delannoy, J.J.: Palaeoflood activity and climate
909 change over the last 1400 years recorded by lake sediments in the NW European Alps, *J.*
910 *Quat. Sci.*, 28, 189–199, 2013.

911 Wilhelm, B., Sabatier, P. and Arnaud, F.: Is a regional flood signal reproducible from lake
912 sediments? *Sedimentology*, 62(4), 1103–1117. 2015.

913 Wirth S.B., Glur, L., Gilli, A. and Anselmetti, F.S.: Holocene flood frequency across the
914 Central Alps – solar forcing and evidence for variations in North Atlantic atmospheric
915 circulation, *Quatern. Sci. Rev.*, 80, 112–128, 2013a.

916 Wirth, S.B., Gilli, A., Simonneau, A., Ariztegui, D., Vannière, B., Glur, L., Chapron, E.,
917 Magny, M. and Anselmetti, F.S.: A 2000-year long seasonal record of floods in the southern
918 European Alps. *Geophys. Res. Lett.*, 40(15), 4025-4029, 2013b.

919

920 Table 1. Radiocarbon dates of core FOR13P4. We calculated the event-free sedimentary
921 depth by removing the graded beds, which were considered to be instantaneous deposits. See
922 text for explanation, nature of samples and calibration procedures.

923

924 Figure 1. (A) Location of Lake Foréant in the Western Alps, (B) compared to the locations of
925 the previously studied Lakes Blanc (BLB, Wilhelm et al., 2012b; BAR, Wilhelm et al., 2013)
926 and Lake Allos (ALO, Wilhelm et al., 2012a). (C) Location of the Foréant catchment area in
927 the hydrological network flowing to the village of Ristolas. (D) Geological and
928 geomorphological characteristics of the Foréant catchment area. (E) Bathymetric map of Lake
929 Foréant and coring sites.

930

931 Figure 2. Lithological descriptions of cores and stratigraphic correlations based on
932 sedimentary facies. For each core, a photography (left), a X-ray image (center) and a
933 stratigraphic log is shown (right). ^{14}C samples are indicated by red stars. Variability in grain-
934 size distribution and geochemical elements (Fe, K, Ca and Mn) is shown for the core
935 FOR13P2. Mo_{inc} used as a high-resolution proxy of density is shown close to the density
936 measurements performed by gamma-ray attenuation. Variability in *Sporomiella* concentration
937 is shown for core FOR13P3.

938

939 Figure 3. To illustrate the different geochemical characteristics of the sedimentary
940 background and the graded layers, their Ca intensities were plotted against their Fe, K and Mn
941 intensities. FIL refers to flood- and MMIL to mass-movement-induced layers. The different
942 MMILs are labelled according to figure 2.

943

944 Figure 4. (A) ^{226}Ra , ^{210}Pb and ^{137}Cs profiles for core ALO09P12. (B) Application of a CFCS
945 model to the event-free sedimentary profile of ^{210}Pb in excess (without the thick graded beds
946 considered as instantaneous deposits). (C) Resulting age–depth relationship with 1σ
947 uncertainties and locations of the historic ^{137}Cs peaks supporting the ^{210}Pb -based ages. C
948 corresponds to the historic ^{137}Cs peak of Chernobyl (AD 1986) and MP to the maximum ^{137}Cs
949 peak of the nuclear fallout (AD 1963).

950

951 Figure 5. (A) Raw declination and inclination profiles of cores FOR13P1, FOR13P2 and
952 FOR13P4. (B) The same profiles after removal of the thickest graded beds (interpreted as
953 event layers) and adjustment of the different specific-core depths to a common reference
954 depth. (C) Average of profiles shown in (B). (D) Correlation to the ARCH3.4k model
955 reference curve of declination and inclination (Donadini et al., 2009; Korte et al., 2009).
956 ChRM means characteristic remanent magnetization. The well-correlated declination and
957 inclination features are labelled D-x and I-x, respectively.

958

959 Figure 6. Age–depth model for core FOR13P2 calculated using the “clam” R-code package,
960 combining historic ^{137}Cs peaks, ^{210}Pb ages, calibrated ^{14}C ages and magnetic features on the
961 left side. Probability distribution frequencies of mass-movement ages and possible
962 correlations to historical earthquakes on the right side.

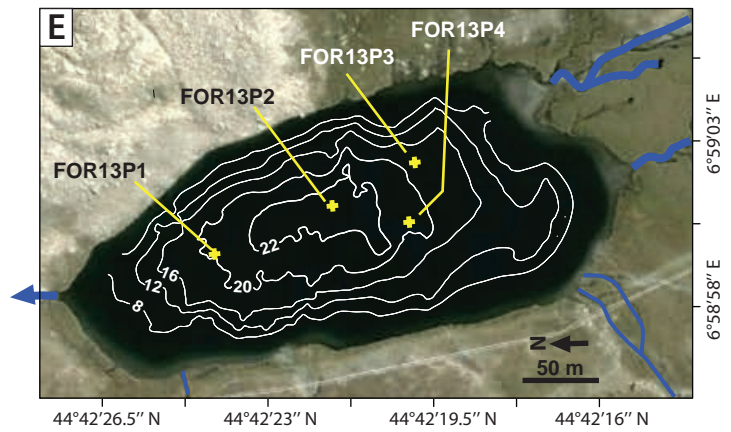
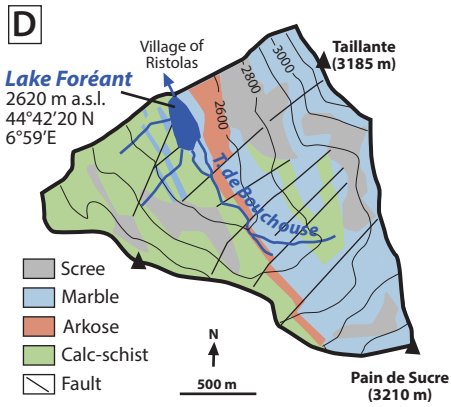
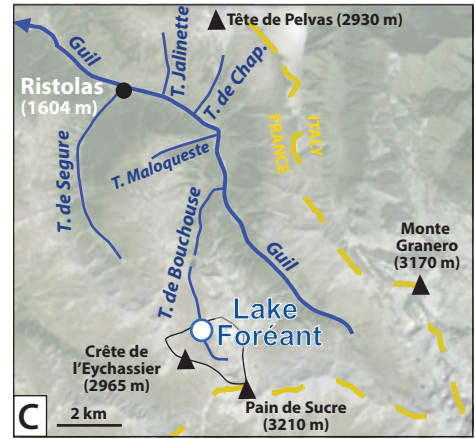
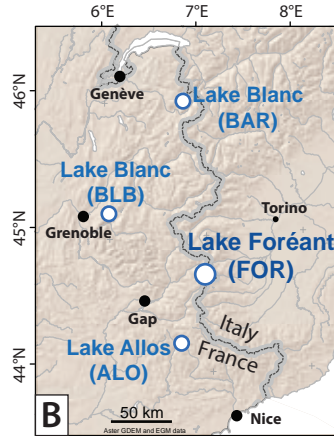
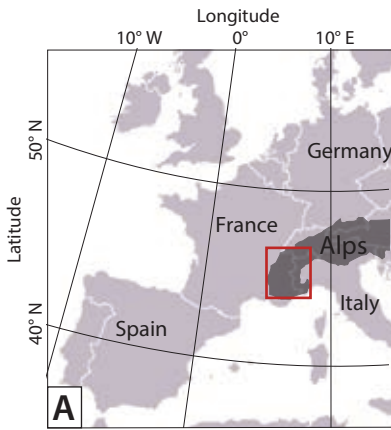
963

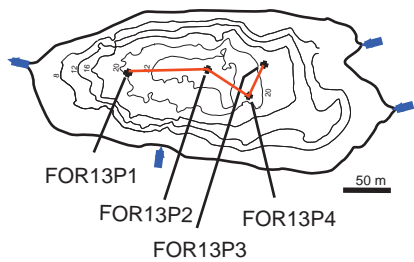
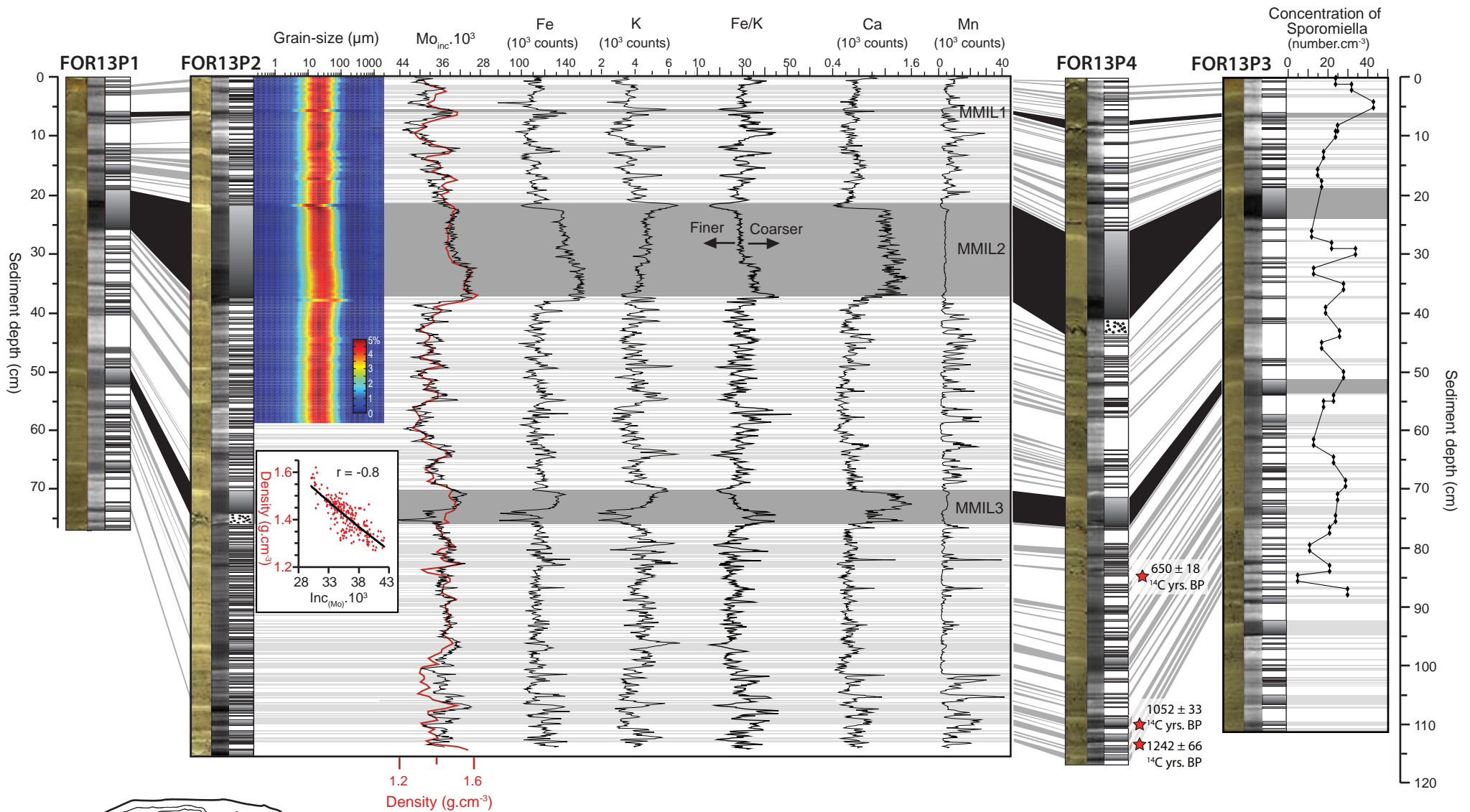
964 Figure 7. Comparison over the last 250 years of the reconstructed Foréant flood frequency
965 (11-yr running sum) and intensity (thickness of flood deposits) with the frequency (11-yr
966 running sum) of historical floods at Ristolas. The question mark refers to a possible gap in the
967 historical data.

968

969 Figure 8. Comparison over the last millennium of (B) the reconstructed Foréant flood
970 frequency (31-yr running average) and intensity (thickness of flood deposits) with (A) the
971 Allos flood record from the southern French Alps (Wilhelm et al., 2012a), (C) the BAR flood
972 record from the northern French Alps (Wilhelm et al., 2013) and (D) the tree-ring-based
973 summer temperature for the European Alps (Büntgen et al., 2011). The reconstructed
974 *Sporomiella*-type flux is also shown next to the Foréant flood record to highlight potential
975 human impacts (i.e. grazing) on the erosion processes that might bias the flood record. The
976 red stars below the Foréant record show the chronological markers with their 2-sigma
977 uncertainty ranges.

BE nr.	Core	Core depth (cm)	Core depth in core FOR13P2 (cm)	Event-free depth in core FOR13P2 (cm)	Material	¹⁴ C yrs. BP	Cal yrs. BP ($\pm 2\sigma$)	Cal yrs. AD ($\pm 2\sigma$)
2094.1.1	FOR13P4	84-85	82-83	33	Terrestrial plant remains	650 \pm 18	561-665	1285-1389
2095.1.1	FOR13P4	109-111	106-108	46 \pm 0.5		1052 \pm 33	923-1052	898-1027
2096.1.1	FOR13P4	113-115	110-112	47 \pm 0.5		1242 \pm 66	1004-1292	658-946





Lithological description

- Sedimentary background
- Graded layer
- Coarse-grain layer

Stratigraphical correlation

- based on graded layers
- based on coarse-grain layers and their overlying graded layers (labelled MMIL)

Dating data

- ★ ^{14}C sample and age

◇ Sed. Background

Graded layers:

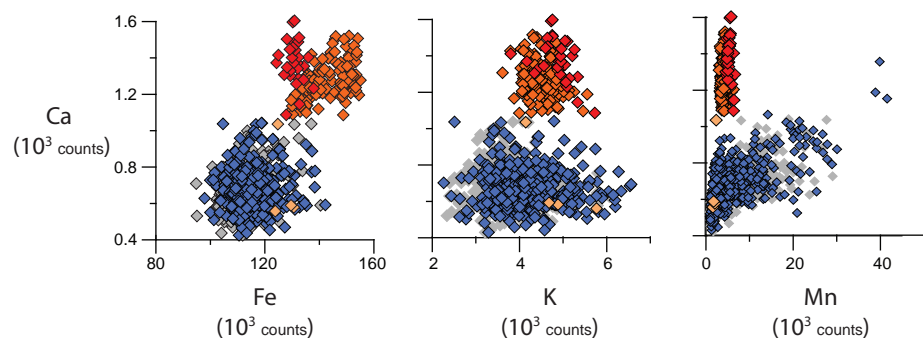
◆ FIL

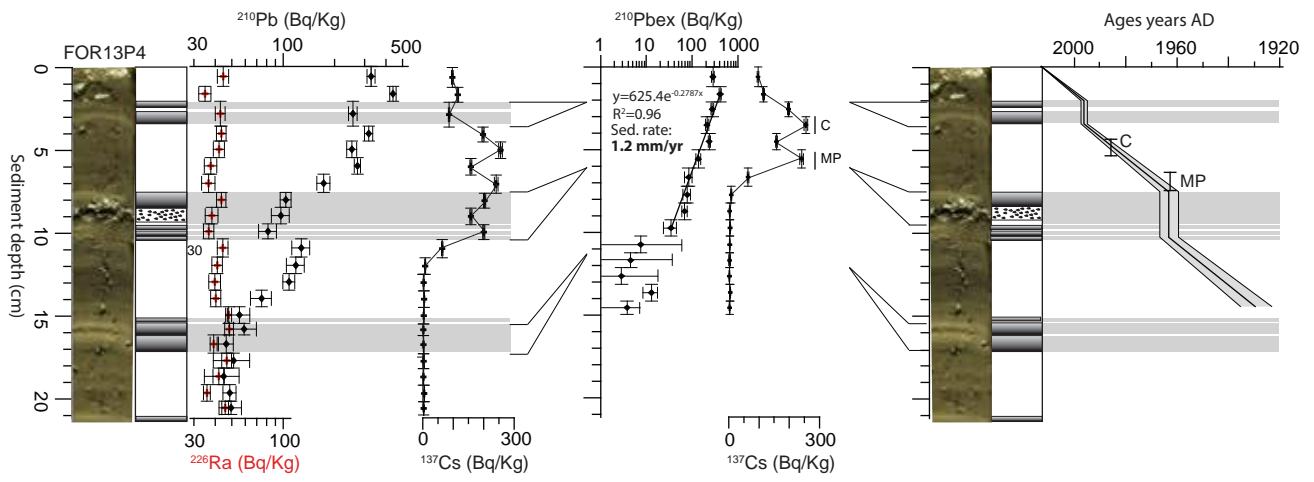
Graded layers that overlie
a coarse-grained layer:

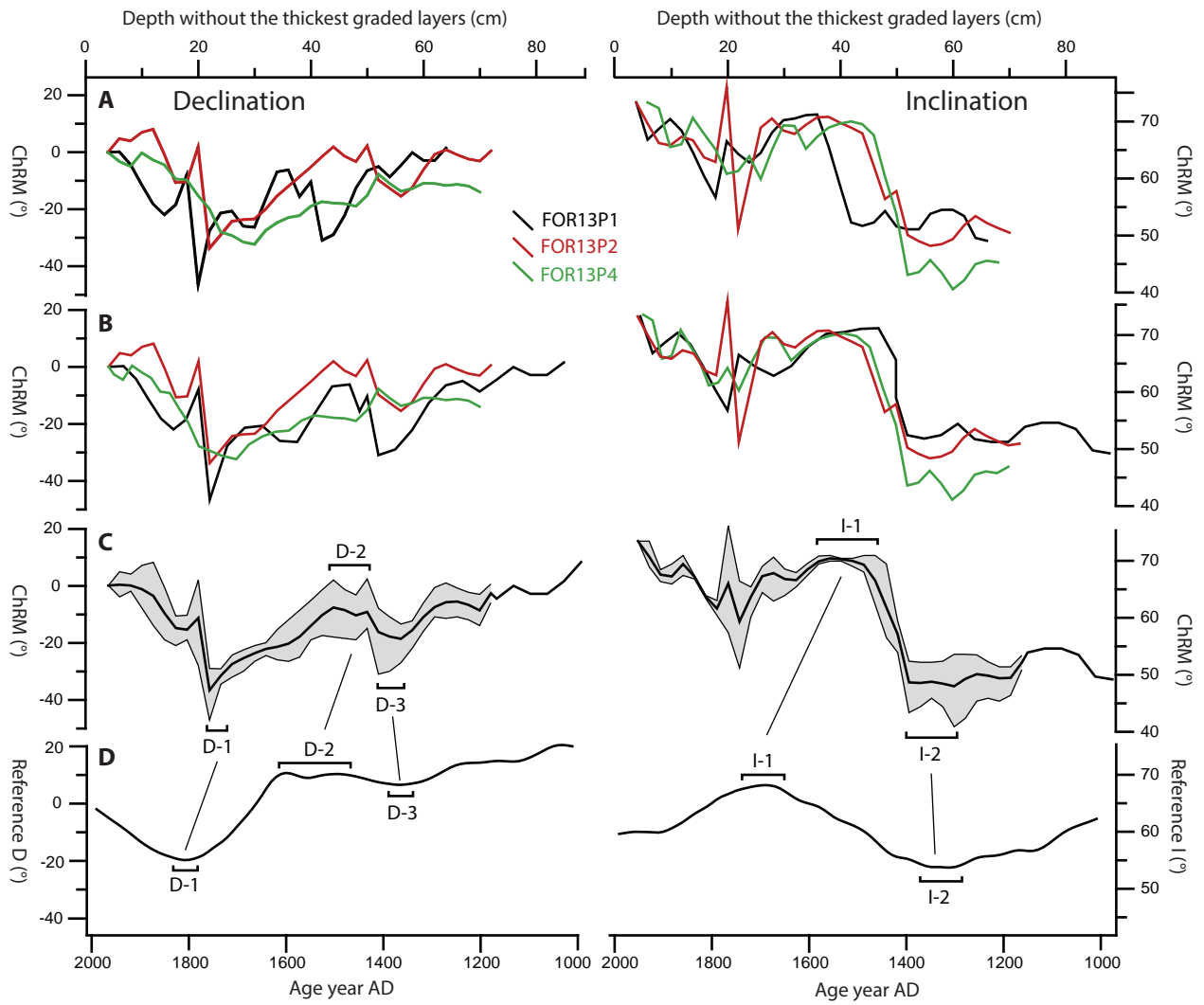
◆ MMIL1

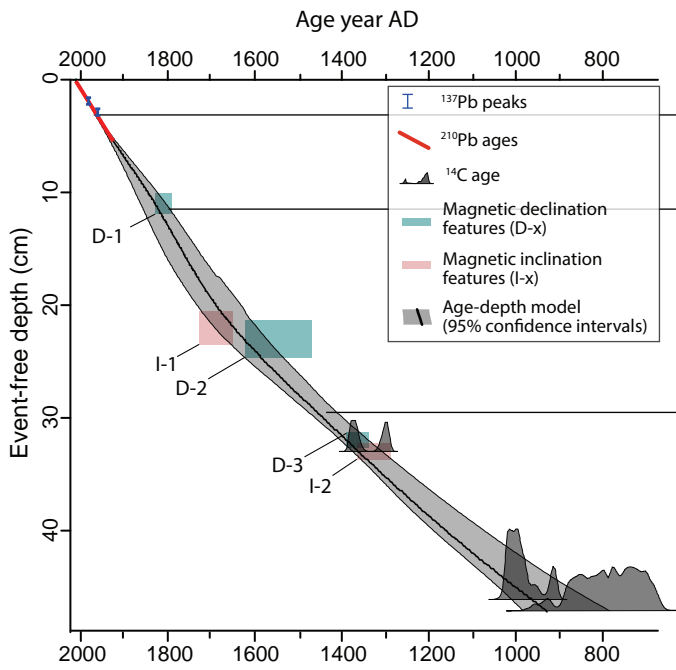
◆ MMIL2

◆ MMIL3



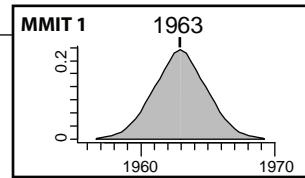






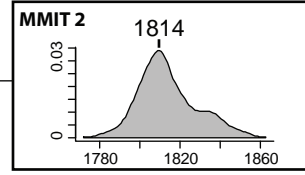
Age probabilities of mass-movement deposits

Possible correlated historical events



Earthquake in AD 1959

$I_0 = \text{VII-VIII}$
 $d = 23 \text{ km}$



Earthquake in AD 1808

$I_0 = \text{VIII}$
 $d = 22.5 \text{ km}$

