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A compilation of Western European terrestrial records 60-8 ka BP: towards an understanding of latitudinal climatic gradients

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36 Walker and Barbara Wohlfarth.

37 **Abstract**

38 Terrestrial records of past climatic conditions, such as lake sediments and speleothems, provide data of
39 great importance for understanding environmental changes. However, unlike marine and ice core records,
40 terrestrial palaeodata are often not available in databases or in a format that is easily accessible to the non-
41 specialist. As a consequence, many excellent terrestrial records are unknown to the broader palaeoclimate
42 community and are not included in compilations, comparisons, or modeling exercises. Here we present a
43 compilation of Western European terrestrial palaeo-records covering, entirely or partially, the 60-8-ka
44 INTIMATE time period. The compilation contains 56 natural archives, including lake records, speleothems,
45 ice cores, and terrestrial proxies in marine records. The compilation is limited to include records of high
46 temporal resolution and/or records that provide climate proxies or quantitative reconstructions of
47 environmental parameters, such as temperature or precipitation, and that are of relevance and interest to
48 a broader community. We briefly review the different types of terrestrial archives, their respective proxies,
49 their interpretation and their application for palaeoclimatic reconstructions. We also discuss the
50 importance of independent chronologies and the issue of record synchronization. The aim of this exercise is
51 to provide the wider palaeo-community with a consistent compilation of high-quality terrestrial records, to
52 facilitate model-data comparisons, and to identify key areas of interest for future investigations. We use
53 the compilation to investigate Western European latitudinal climate gradients during the deglacial period
54 and, despite of poorly constrained chronologies for the older records, we summarize the main results
55 obtained from NW and SW European terrestrial records before the LGM.

56 **Keywords:** Western Europe, terrestrial records, last deglaciation, databases, INTIMATE

57 **1. Introduction**

58 Since the discovery of abrupt climate change at decade-to-century time-scales during the last glacial period,
59 evidence of rapid climate variability has become available from an ever-growing number of palaeo-archives
60 worldwide. In the North Atlantic region, the climate of the last glacial period is characterized by a number

61 of abrupt warming events referred to as Dansgaard-Oeschger events, which are most strongly expressed in
62 the Greenland ice cores (Dansgaard et al., 1993). North Atlantic marine records of the last glacial period are
63 sometimes dominated by ice rafting events, the so-called Heinrich events (Hemming, 2004) while other
64 terrestrial archives from Western Europe do not always show a strong response to such climatic events
65 (Ampel et al., 2010). The abrupt climate variability of the last glacial period differs from the slowly varying
66 orbital parameters that are believed to be the main drivers of climate change during this period.

67 The present work is concerned with the 60-8 ka period, covering the second half of the last glacial,
68 including the Last Glacial Maximum (LGM, ca 22-18 ka BP) and the deglacial period (ca. 14.7-11.7 ka BP),
69 and the early Holocene (ca 11.7-8 ka BP). Furthermore, the period is divided into Marine Isotope Stage 3
70 (MIS3, ca 60-28 ka BP), MIS2 (ca 28-11.7 ka BP), and part of MIS1 (last 11.7 ka). Within the INTegration of
71 Ice-core, MARine, and TERrestrial records (INTIMATE) community, the Greenland ice core stratigraphy
72 (Rasmussen et al., this issue) has been applied as a template for North Atlantic climate variability for the
73 60-8 ka period (Blockley et al., 2012).

74 Compilation of palaeoclimatic evidence from palaeo-archives is a prerequisite for obtaining an overview of
75 past climatic change. Bringing palaeo-records from various archives together on a common chronological
76 framework allows individual records to be placed in a broader context, exploration of both temporal and
77 spatial evolution of climate, and investigation of the drivers and mechanisms behind past climatic changes
78 together with the resulting impacts of climatic change (Birks et al., this issue). Furthermore, data
79 compilations enable model-data comparisons which can help to decipher the mechanisms that drive
80 climatic change.

81 A number of important studies have provided compilations of records of past climatic variability between
82 60-8 ka on both global and regional scales. (Voelker, 2002) provided the first global compilation of marine
83 and terrestrial sites, showing evidence of centennial-scale climatic change during the last glacial period.
84 Since then, this evidence has been supported and extended by a number of high resolution records, such as

85 Antarctic ice cores (EPICA, 2006), Cariaco Basin and Arabian Sea sediment cores (Deplazes et al., 2013), and
86 Asian and American speleothems (Asmerom et al., 2013; Fleitmann et al., 2009; Wang et al., 2007, 2001).
87 The deglacial period provides more records and accurate chronologies than earlier periods, and a
88 comprehensive compilation of global proxy records is provided by (Clark et al., 2012; Shakun et al., 2012).
89 Other studies focused on the climate of the last glacial in the North Atlantic region, which is strongly
90 influenced by the presence of large continental ice sheets (Svendsen et al., 2004). (Björck et al., 1996)
91 linked deglacial records from different North Atlantic archives. Voelker et al. (this issue) provides the most
92 recent compilation of high resolution North Atlantic surface water temperatures for the last 60 ka, and a
93 comprehensive compilation of last glacial millennial-scale European vegetation records was provided by
94 (Fletcher et al., 2010a).
95 This paper presents a new compilation of Western European quantitative terrestrial climate records in the
96 60-8 ka INTIMATE time period (Fig. 1, Table 1) with a focus on the deglacial period. We briefly describe the
97 different terrestrial archives, their respective proxies, interpretation and application for palaeoclimatic
98 reconstructions. Furthermore, we discuss the importance of chronologies and record synchronization. We
99 investigate Western European latitudinal climate gradients during the deglacial period, and discuss the
100 period before the LGM.

101 **2. Data compilation and selection criteria**

102 The present compilation broadly covers Western Europe (with 15°E of longitude as an approximate
103 boundary) and is complemented by parallel compilations covering Eastern Europe (Feurdean et al., this
104 issue), the Austrian and Swiss sectors of the Alpine region (Heiri et al., this issue) and marine North Atlantic
105 records (Voelker et al., this issue). The scarcity of records from Scandinavia for this time period does not
106 require an additional compilation and those sequences are included here. The present compilation includes
107 published records from lakes, speleothems, ice cores, and terrestrial proxies in marine records. The

108 following selection criteria were applied to obtain a dataset with a balanced geographical and temporal
109 distribution:

- 110 - The record must cover a relevant interval of the 60-8-ka INTIMATE time frame, even if it is not
111 continuous.
- 112 - The record must include at least one quantitative or semi-quantitative climate-related variable
113 (proxy or reconstruction). For example, temperature proxy/estimates, precipitation
114 proxy/estimates, vegetation/biome parameters, or lake level estimates. Preferably, all parameters
115 have an error estimate attached to them.
- 116 - An age model must be available. Preferably the age model is independent, or tightly linked to a
117 well-dated archive. All information about how the age model is constructed must be provided and,
118 preferably, uncertainty estimates should be available.
- 119 - Included records must be published or made available for this work.

120 Our compilation includes 56 Western European records of past climatic conditions on land, mostly
121 presenting temperature and precipitation reconstructions, but also records of dust supply and vegetation
122 biomes (Table 1; Fig. 1, Electronic Supplementary Material). The compilation contains 38 records from lake
123 sediments, 9 records from speleothems, one ice core record and 8 records of terrestrial remains (e.g.
124 pollen) in marine sediments. Loess, tree rings, and glacial-related records were also considered for the
125 compilation, but unfortunately we were unable to obtain any of these records that fulfilled all the selection
126 criteria. Data are available in the Electronic Supplementary Material (ESM) attached to this paper. It is
127 beyond the scope of the present work to re-interpret existing records or proxies or to provide revised age
128 models for the included records. The present compilation in a standardized format does, however, permit
129 future revision of age models and synchronization of records whenever new evidence or improved age
130 models become available.

131 The time period covered by each record is indicated in Fig. 2. The scarcity of records available from 60 to 25
132 ka BP is apparent, whereas the last deglaciation is relatively well represented. The lack of quantitative data
133 from MIS3 sequences represents a major limitation of this compilation for understanding abrupt climate
134 variability on land associated with DO cycles. In addition, this data compilation reflects the fragmentary
135 character of some of the records (e.g. speleothem from El Pindal cave) while other records are continuous
136 (e.g. most lake records from SW Europe and marine sediments). However, for this compilation, the quality
137 of quantitative reconstructions has been the main selection criterion. As shown in the location map (Fig. 1),
138 the compiled sites are located from 75°N (Greenland ice core NGRIP) to 36°N (Alboran Sea core MD95-
139 2043), but most (60%) are concentrated in a narrow latitudinal band from 53°N to 45°N, which highlights
140 the scarcity of records from Scandinavia and the Mediterranean region (southern Spain, southern Italy).
141 Scandinavian sites covering the INTIMATE time period (60-8 ka BP) are restricted since that area was ice-
142 covered during most of last glacial cycle (see section 5.2.1). On the other hand, records from the
143 southernmost region of Western Europe are limited due to the aridity of the climate during MIS3 (e.g.
144 Moreno et al., 2012) (see section 5.2.2).

145 Using different approaches and methods, a range of proxies was quantified (Table 1, ESM) to obtain
146 numerical data on temperature (e.g. mean July air temperature from chironomids in lacustrine sequences),
147 precipitation variables (e.g. lake level reconstructed from carbonate in karstic lakes), and vegetation
148 distribution (e.g. biomes from pollen data in lake and marine sediments). Temperature reconstructions are
149 the most common records represented in this database (33 records), followed by records reflecting past
150 variation in precipitation (23 records). Temperature data come mainly from chironomids (19), pollen (8),
151 other organisms (4), sedimentological data (2), ice $\delta^{18}\text{O}$ isotopes (1), $\delta^{18}\text{O}$ isotopes in lake sediment records
152 (1) and pollen data in marine and lake records (8). Precipitation, expressed as the isotopic composition of
153 precipitation or as effective precipitation (water availability), is reconstructed, quantitatively or semi-
154 quantitatively, from stable isotopes in speleothems (8), $\delta^{18}\text{O}$ isotopes in lake sediment records (2), by lake

155 level variations (8) and by pollen data (4). Pollen data are also used to reconstruct vegetation cover (e.g.
156 pollen in marine sediments, 6 records) and biomes (1).

157 **3. Terrestrial archives of past climate changes: advantages and limitations**

158 Each type of archive included in the database is unique and has advantages and limitations regarding its
159 ability to reconstruct climate quantitatively (Table 2). To explore those characteristics, we outline the
160 methodology, type of quantitative information and dating methods used in analysis of lake sediments,
161 speleothems, ice cores, and terrestrial remains in marine records. Methods used to obtain quantitative
162 proxy climate data are summarized in the ESM. Finally, other terrestrial archives (tree rings, loess and
163 glacier deposits), which are not included in this compilation, are briefly discussed.

164 *3.1. Lake sediments*

165 Lacustrine archives potentially record environmental variability (e.g. water chemistry, temperature) at high
166 temporal resolution and high sensitivity to regional climatic changes. When annual sedimentation can be
167 discerned (e.g. varves in maar lakes), records with robust chronologies and high temporal resolution can be
168 obtained, potentially providing detailed climate reconstructions (Brauer et al., 2000; Sirocko et al., 2013).
169 Some lacustrine sediment records are discontinuous (e.g. resulting from sediment erosion, slumping
170 events, or the temporary disappearance of the lake following drought), and identifying and dating those
171 hiatuses can provide important palaeoclimatic information (Vegas et al., 2010). The resolution of a
172 lacustrine environmental reconstruction is influenced by its continuity, sedimentation rate, sampling
173 resolution and chronological accuracy. Lake sediment archives can provide detailed records spanning
174 several hundred thousand years to the recent decades.

175 In spite of the numerous proxies available in lake sediments, only a few proxies such as pollen, diatoms,
176 chironomids and ostracods (e.g. (Heiri et al., 2007; Peyron et al., 2005) can be used to provide quantitative
177 reconstructions. Additionally, lake level variations can be obtained after a detailed sedimentological study
178 (Magny et al., 2013a). Calibration procedures, based on analyses from multiple lakes, describe the modern

179 relationship between individual taxa and one or more environmental variables (e.g. (Battarbee, 2000; Heiri
180 et al., 2011). Although widely used in palaeolimnology, these are not exempt from problems and criticism
181 (Birks et al., 2010; Juggins, 2013). Promising new proxies are becoming available. For instance, $\delta^{18}\text{O}$
182 measured in lacustrine diatoms can be used to infer lake water temperature (Leng and Henderson, 2013),
183 the isotopic composition of *n*-alkyl diols, *n*-C₂₃ or C₂₉ alkanes can be used to reconstruct changes in
184 precipitation, evaporation and/or the origin of air masses (Sachse et al., 2012) and the TEX86 organic
185 geochemical index is becoming an excellent palaeothermometer (Blaga et al., 2013).

186 Lake sediment chronology is mainly based on AMS ¹⁴C dates which may be problematic. Ideally, ¹⁴C dating
187 of lake sediment records is based on terrestrial plant remains. For sediments older than 15 ka BP and
188 records deposited in treeless environments, dating will often have to rely on bulk sediment samples.
189 Furthermore, in some karstic lakes with a limestone catchment, the hardwater and reservoir effects have to
190 be taken into account if the organic content comes from the aquatic environment (Moreno et al., 2010b).
191 Finally, the ¹⁴C method becomes increasingly imprecise and difficult to calibrate beyond 50,000 years
192 (Ramsey et al., 2012). In such cases, other dating techniques are used, such as U-Th disintegration series on
193 lacustrine carbonates (Pèrez-Obiol and Julià, 1994), Optically Stimulated Luminescence (OSL) dating on
194 siliciclastic sediments (Moreno et al., 2012) or palaeomagnetic excursions in particularly long records
195 (Vegas et al., 2010). In addition to these absolute dating techniques, varve counting and tephras are widely
196 used to provide a chronological framework (e.g. (Wulf et al., 2004).

197 3.2. *Speleothems*

198 The major advantage of speleothems (stalagmites and calcite flowstones are the most commonly studied)
199 as palaeoclimatic archives is their dating accuracy. The Uranium-Thorium method allows dating of
200 speleothems up to about 650 ka (Hellstrom, 2003). When geochemistry is favorable, old calcites can be
201 dated to more than 1 million years by applying the U-Pb method (Woodhead et al., 2006). Within the last
202 ca. 100 ka, two sigma (95%) errors can be below 1% of the age, which makes speleothems one of the most

203 accurately datable terrestrial archives. However, as speleothem growth is sensitive to climatic conditions,
204 they may form hiatuses during cold and dry phases (Genty et al., 2010a). On the other hand, precise
205 determination of the durations of these hiatuses may provide valuable information about climatic
206 thresholds that affect regional climatic conditions (Moreno et al., 2010a; Stoll et al., 2013; Zhornyak et al.,
207 2011).

208 The stable isotopic composition of calcite is the most frequently used palaeoclimatic proxy in speleothems
209 (McDermott, 2004). Calcite $\delta^{18}\text{O}$ is linked to many factors, most importantly rainfall origin and trajectory,
210 precipitation amount, and atmospheric temperature. Calcite $\delta^{18}\text{O}$ reflects variations in precipitation sources
211 and amounts, especially for tropical climates (Cruz et al., 2006; Wang et al., 2008, 2001) where $\delta^{18}\text{O}$
212 decreases with increasing monsoon precipitation. Western Mediterranean speleothem records (e.g.
213 Corchia and Chauvet caves) also display a coupling between $\delta^{18}\text{O}$ and climate (Drysedale et al., 2005; Genty
214 et al., 2006; Zanchetta et al., 2007), but the exact causes of $\delta^{18}\text{O}$ variations are not well understood: most
215 likely both rainfall and temperature play a role, along with the sources of precipitation. At specific high
216 altitude sites, such as the Alps (Boch et al., 2011), the calcite $\delta^{18}\text{O}$ has been shown to be controlled mainly
217 by temperature through seasonal effects.

218 Carbon isotopes ($\delta^{13}\text{C}$) are sensitive to climatic changes through the vegetation (C3/C4 photosynthetic
219 pathways) (Dorale et al., 1998) and, more often, to soil microbial activity changes (Dorale et al., 1998;
220 Fleitmann et al., 2009; Genty et al., 2003a; Hellstrom et al., 1998; Moreno et al., 2010a). $\delta^{13}\text{C}$ may also be
221 partly controlled by kinetics during calcite precipitation (Mickler et al., 2006) and responds to millennial-
222 scale climatic variations (e.g. DO events) as well as to centennial-scale changes, such as those during GI-1
223 and GS-1. The possibility that kinetics and/or prior calcite precipitation (PCP) may also control the $\delta^{13}\text{C}$
224 signal does not significantly affect this relationship; it may even amplify it because these effects are usually
225 positively related to amplify climatic effects: for instance, $\delta^{13}\text{C}$ increases as the climate becomes colder and
226 drier (Genty et al., 2006).

227 For the majority of speleothems, especially in the mid-latitudes, it is difficult to distinguish the effects of
228 temperature from hydrological change only using calcite isotopes. Trace element analyses can provide
229 complementary information about humidity through the water/rock interaction, the residence time of
230 seepage water, and through the preferential incorporation of specific elements (e.g. Sr) into the calcite
231 lattice controlled by the growth rate (Borsato et al., 2007; Bourdin et al., 2011; Stoll et al., 2012).

232 Speleothem calcite stable isotopes and trace elements do not provide quantitative reconstructions of past
233 temperature or precipitation. However, recent progress in fluid inclusion (FI) analyses (Dublyansky and
234 Spotl, 2009; Vonhof et al., 2006) and on calcite clumped isotopes (Daeron et al., 2011, 2008; Guo et al.,
235 2008) has demonstrated its potential for estimating climatic parameters. A very promising approach is
236 found in the combination of FI analyses with clumped isotopes data, as this appears to be the only way to
237 correct for kinetics effects (Wainer et al., 2011).

238 *3.3. Ice cores*

239 The only place in the North Atlantic region where ice core records are continuous to the last glacial is
240 Greenland (Johnsen et al., 2001). The Greenland ice cores record climatic change at sub-annual resolution
241 through a variety of proxies for most of the last glacial period (Steffensen et al., 2008) thus providing a
242 consistent record of abrupt climatic change with characteristic millennial-scale DO events (Dansgaard et al.,
243 1993). In Europe, ice cores have been retrieved from glaciers in the Alps, Iceland, Svalbard, and from caves
244 (Perçoiu et al., 2011).

245 The most commonly used proxies from ice cores are 1) the water isotopic composition of the ice, which is
246 used as a temperature proxy, 2) the gas content of the ice, which provides a record of past atmospheric
247 composition and greenhouse gases, and 3) impurities in the ice that record past volcanism, storminess,
248 biomass burning, solar activity, and anthropogenic contributions.

249 Past Greenland ice accumulation rates are estimated from annual layer thickness and ice flow models that
250 simulate the annual layer thinning with depth. Past accumulation rates are known to have changed

251 significantly over the last glacial cycle, with significantly higher accumulation rates during interglacial
252 periods than during the glacial (Alley et al., 1993). This, in turn, has led to changes in Greenland ice sheet
253 volume and thus influenced past sea levels (Dahl-Jensen et al., 2013; Vinther et al., 2009). Within the last
254 glacial period, accumulation patterns have varied strongly in parallel with the abruptly changing climate,
255 with an approximate doubling of precipitation associated with the onset of DO events.

256 The deep Greenland ice cores are independently dated by a combination of annual layer counting for the
257 younger part and ice flow modeling for the deepest part (Wolff et al., 2010). In the Holocene, annual layers
258 can be resolved in the water isotopic signal (Vinther et al., 2006), whereas impurity records can be applied
259 for annual layer counting in the last glacial period (Rasmussen et al., 2006). The glacial part of the
260 Greenland ice cores dates to 60 ka BP, with an absolute uncertainty of about 5% using a multi-proxy
261 impurity dataset from the NorthGRIP ice core (Svensson et al., 2008). The modeled time scale for the oldest
262 part of the ice is based on an empirical relationship between the water isotopic composition and past
263 accumulation patterns (Buchardt et al., 2012). All the major deep Greenland ice cores (Dye-3, GRIP, GISP2,
264 NGRIP, NEEM) have been internally synchronized by volcanic reference horizons (Rasmussen et al., 2013)
265 Seierstad et al., this issue) and linked to other palaeoarchives in the North Atlantic region by tephra layers
266 (Blockley et al., 2012). Greenlandic and Antarctic ice cores are tightly synchronized by atmospheric gas
267 concentration records, records of cosmogenic isotopes, and bipolar volcanic events (Veres et al., 2013).

268 *3.4. Terrestrial tracers in marine records*

269 The study of terrestrial tracers in marine sediment cores from hemipelagic settings provides important
270 information on past climatic changes and their impacts on terrestrial environments. A large amount of data
271 is available from the Atlantic margin of SW Europe and the western Mediterranean Sea for the 60-8 ka
272 period (see Voelker et al., this issue). The particular value of terrestrial tracers in marine cores is the
273 possibility for direct land-sea correlation. The sequences benefit from continuous sedimentation without
274 hiatuses, but their chronologies are derived from either AMS ¹⁴C dating of planktonic foraminifera, the

275 alignment of planktonic $\delta^{18}\text{O}$ or SST curves to a reference record (typically Greenland $\delta^{18}\text{O}$), or a
276 combination of these two approaches. In addition, the reservoir effect of the ocean is variable in time and
277 space and is particularly important for accurate determination of the age of abrupt changes along MIS3 and
278 deglaciation in marine sediments (Austin et al., 2011). Due to the influence of bioturbation at the sediment
279 surface, the maximum achievable temporal resolution of these studies is typically limited to ~2 cm (around
280 200-400 yr for the last glacial period).

281 Key terrestrial tracers studied in marine records include pollen and spores (e.g. (Fletcher and Sánchez Goñi,
282 2008; Lézine and Denèfle, 1997; Naughton et al., 2007; Roucoux et al., 2001; Sánchez-Goñi et al., 2008,
283 2002, 2000), and charcoal (Danialu et al., 2009, 2007). Several sedimentological indicators of changing
284 terrestrial influxes have also been examined, for example grain-size (Moreno et al., 2002), clay mineralogy
285 (Bout-Roumazielles et al., 2007), elemental composition (Hodell et al., 2010) and colour (Rogerson et al.,
286 2006). Quantitative climate reconstructions have been performed using the Modern Analogue Technique
287 on pollen assemblages (e.g. (Combourieu Nebout et al., 2009; Fletcher et al., 2010b; Sánchez-Goñi et al.,
288 2002) (see ESM).

289 *3.5. Other terrestrial archives*

290 Tree-rings, loess, soil-palaeosoil or glacier deposits are other sources of information of climate variability on
291 land for the last glacial cycle and deglaciation (Table 2).

292 Tree-rings can be absolutely dated to an exact calendar year and can provide quantified climatic
293 reconstructions (Friedrich et al., 2004). For instance, recently, cellulose isotope composition studies have
294 been used to reconstruct summer temperature (Etien et al., 2008). However, the value of the tree ring
295 archive in the INTIMATE context is restricted by the temporal coverage and, unfortunately, there are few
296 examples extending back to 15 ka BP (Kaiser et al., 2012). Therefore, the challenge is to find enough
297 preserved wood to permit construction of a reliable cross-dated chronology (e.g. Friedrich et al., 2007). To
298 extend the record beyond the period of trees currently alive in Europe, subfossil wood can be recovered

299 from lakes, sediments, alluvial terraces, brown coal, or glaciers. Different tree-ring chronologies from
300 central Europe (Friedrich et al., 2004; Schaub et al., 2008, 2005) have been combined to form a continuous,
301 absolutely dated tree ring record for the Holocene and the last deglaciation, known as the Central
302 European master series (CELM) (Becker, 1993; Friedrich et al., 2004; Kaiser et al., 2012). While this provides
303 a reference for dendrochronological dating, it cannot yet be used for palaeoclimatic reconstruction because
304 the series has not been calibrated against any climate parameters (Friedrich, personal communication).

305 Another source of palaeoclimatic information is Western European loess. Loess deposits frequently develop
306 during glacial periods, and are associated with large floodplains close to ice sheets subject to eolian
307 deflation (Antoine et al., 2001). Their main advantage is the high sedimentation rate (up to 1 cm in 10 yr)
308 and the use of several independent dating methods (^{14}C , OSL, etc.) to provide a record of high temporal
309 resolution. Comparison of loess sequences from Northern France, Belgium and Germany indicates similar
310 stratigraphic patterns during the last glacial cycle (Antoine et al., 2001), clearly resulting from a common
311 climatic control. The most detailed investigations are from Nussloch, Germany (Hatte et al., 2001; Lang et
312 al., 2003; Rousseau et al., 2002), where the loess deposits of the last glacial cycle are exceptionally well
313 developed compared to other European sites. In fact, variations in grain size, magnetic susceptibility and
314 malacofaunal composition of the Nussloch loess sequence have been correlated with Greenland ice
315 records, indicating a striking similarity during cold periods in the last glacial cycle (Antoine et al., 2009;
316 Moine et al., 2008; Rousseau et al., 2007). However, new dating of the Nussloch sequence has provided a
317 different chronology for some horizons (Kadereit et al., 2013; Tissoux et al., 2010) that requires more
318 investigation. Loess sequences from the Austrian and Swiss sectors of the Alpine region are presented by
319 Heiri et al. (this issue).

320 Glacial deposits, mostly moraines, but also glaciolacustrine deposits, are good markers of glacial advances
321 and retreats during last glacial cycle in Western European mountains, such as the Pyrenees (García-Ruiz et
322 al., 2013) and the Alps (Ivy-Ochs et al., 2009). A variety of methods have been used to identify the different
323 phases of glacial retreat, such as the relative topographic and stratigraphic position of the moraines and

324 equilibrium line altitude (ELA) depression relative to the Little Ice Age (LIA). More recently, a combination
325 of dating techniques (cosmogenic radionuclides, ^{14}C , OSL, U-series) has been applied to glacial sediments.
326 Although these archives can provide a useful source of climatic information, their lack of continuity requires
327 special data treatment that precludes their inclusion in this compilation. However, examples from Austria
328 and Switzerland are presented in the Alpine compilation (see Heiri et al. this issue).

329 **4. Linking of records**

330 Good chronological control is critical for comparisons of palaeoclimate records to explore the timing,
331 duration and synchronicity of major climatic events and to evaluate the processes and mechanisms
332 involved in the spatial transference (rapid or gradual) of the climate signal. Therefore, only records with
333 independent age models were included in this compilation, and the dates are included as part of the
334 collected information (see ESM). Although ^{14}C dating is used for most age models in the compiled records,
335 varve counting and U-Th are also employed in many sequences (e.g. maar lakes or speleothems) (Table 1).
336 For some records, the identification of “climate events” was considered part of the chronology (e.g. Sokli
337 record(Engels et al., 2008a). In addition to absolute dating methods, the identification of tephra (volcanic
338 ash) layers in the sediments has become an important way to link distant records in Western Europe
339 chronologically (e.g. (Blockley et al., 2012). In Western Europe, tephra layers between 60-8 ka BP originated
340 from multiple volcanic sources, e.g. Iceland, the Eifel region in western Germany, the French Massif Central
341 and central-southern Italy. Fig. 3 shows the occurrence of the most significant tephra layers encountered in
342 the records compiled for this study.

343 Icelandic volcanoes produced a large number of basaltic, rhyolitic and minor intermediate tephra deposits
344 during the Late Quaternary, some of which are widely dispersed in the North Atlantic and on the European
345 mainland. For example, dating and synchronization of last deglaciation chironomid records in Scotland (e.g.
346 Abernethy Forest, (Brooks et al., 2012) and Norway (Kråkenes, (Brooks and Birks, 2000), used the Vedde
347 Ash ($12,120 \pm 114$ years BP GICC05) and the Saksunarvatn Tephra ($10,297 \pm 90$ years BP GICC05)

348 (Rasmussen et al., 2006) which are also recorded in the NGRIP ice core (e.g., (Abbott and Davies, 2012;
349 Mortensen et al., 2005) (Fig. 3). The GS-1 Vedde Ash is the most widespread Icelandic tephra marker and
350 has been recently detected south of the Alps in the sediment sequences of Lago Piccolo di Avigliana (Lane
351 et al., 2012) (Fig. 3).

352 The Vedde Ash occurs in several terrestrial archives in Central Europe (e.g. Meerfelder Maar; (Lane et al.,
353 2013, 2012) that also include the Laacher See Tephra (LST; $12,880 \pm 40$ varve years BP; Brauer et al., 1999a)
354 from the Eifel volcanic field in western Germany. The phonolitic LST forms a reliable anchoring point,
355 particularly in annually laminated sediments for the transition from GI-1a to GS-1 (e.g., Brauer et al., 1999b;
356 Lane et al., 2011a; Wulf et al., 2013). The LST is widespread in lacustrine sediments in Central Europe (see
357 overview in Riede et al., 2011). Furthermore, a basanitic tephra of possible Eifel provenance was found in
358 the Les Echets palaeolake sequence ca. 500 km southwest of its source (Fig. 3) indicating great potential for
359 synchronizing other long French pollen sequences at ca. 42-45 ka BP (Veres et al., 2008).

360 SW European records (Spain, Portugal) so far lack tephras due to both the predominately easterly dispersal
361 of widespread ashes from Holocene Massif Central and Late Quaternary Italian volcanoes, and the local
362 restriction of pyroclastic deposits deriving from alkaline volcanic provinces in NE and SE Spain (e.g. Garrotxa
363 Volcanic Field; (Cebria et al., 2000). On the contrary, the Central Mediterranean region has a high potential
364 of tephrostratigraphical linking, as continuous Late Quaternary explosive activities of Central and Southern
365 Italian volcanoes produced numerous tephra layers. As an example, more than 120 tephra events are
366 recorded during the last 60 ka BP in the tephra reference profile and pollen record of Lago Grande di
367 Monticchio (Wulf et al., 2004), including the Pomici Principali (ca. 12 ka BP) and the Neapolitan Yellow Tuff
368 (14.1 ka BP) (Fig. 3).

369 **5. Discussion**

370 The time interval considered in INTIMATE network (60-8 ka BP) was characterized by large environmental
371 changes across Western Europe as a response, among other mechanisms (Clement and Peterson, 2008), to

372 (1) advances and retreats of the Scandinavian Ice Sheet (SIS) (Svendsen et al., 2004) and (2) rapid
373 oceanographic changes associated to the different intensity modes of the thermohaline circulation
374 (Rahmstorf, 2002). In general, cool and dry climates were identified in Western European sequences during
375 Greenland Stadials (GS) while the climate was warmer and more humid during Greenland Interstadials (GI)
376 (Fletcher et al., 2010a). Our compilation of quantitative reconstructions of climate variability allows a
377 comparison of records from northern and southwestern Europe during the last deglaciation interval (GS-
378 2.1-a, GI-1, GS-1) (Fig. 4) and before the LGM (GS-2.1b) (Fig. 5).

379 *5.1. Last deglaciation: exploring latitudinal gradients*

380 The last deglaciation is characterized by the decay of ice sheets, a global mean sea-level rise of
381 approximately 80 m, large disturbances in terrestrial and marine ecosystems and a net release of the
382 greenhouse gases CO₂ and CH₄ into the atmosphere (Clark et al., 2012). During this period, changes in
383 atmospheric and ocean circulation affected the global distribution and fluxes of water and heat, resulting in
384 a series of abrupt climatic changes (GI-1a to GI-1e, GS-1), which have broadly similar trends in
385 palaeoclimate records from many sites throughout the North Atlantic region (Björck et al., 1998). The last
386 deglaciation is well represented in most of the records compiled in our database (Fig. 2), thus allowing us to
387 explore (1) latitudinal gradients in the response time to the onset of the last deglaciation (ca. 14.7 ka BP in
388 NGRIP ice core) and (2) the characterization of GI-1 in terms of temperature change and precipitation
389 variability.

390 Fig. 4 shows a latitudinal transect of high-resolution terrestrial records during 16-8 ka cal BP. Most of these
391 records are summer temperature reconstructions, although some are isotopic records from cave
392 speleothems which may be influenced by both temperature and precipitation (e.g. (Genty et al., 2006). The
393 Savi $\delta^{18}\text{O}$ record is interpreted as representative of temperature changes (Frisia et al., 2005) while $\delta^{13}\text{C}$
394 records from Chauvet and El Pindal caves are the expression of changes in soil microbial activity and
395 degassing in relation to changes in humidity (Genty et al., 2006; Moreno et al., 2010a). Neglecting

396 asynchronous responses at the onset and termination of GI-1 that may be due to age model uncertainties,
397 in general, a similar pattern is apparent between the records, which consistently indicate a warmer/wetter
398 climate during GI-1 and colder/drier conditions during GS-1. The absolute summer temperature difference
399 between the two intervals varies around 5-6 °C (with a minimum around 3°C in Lago Piccolo di Avigliana
400 and a maximum around 8°C in Egelsee), well above the error for temperature reconstructions based on
401 chironomids (about 1-1.5°C). No consistent latitudinal differences are observed in the magnitude of
402 temperature change from GI-1 to GS-1. However, northern records (e.g. Jansvatnet, Krakenes, Lough
403 Nadourcan) indicate a larger temperature increase at the onset of the Holocene than at the onset of GI-1,
404 similar to the NGRIP ice core record (Fig. 4). In contrast, alpine records (e.g. Savi, Egelsee, Lago Piccolo di
405 Avigliana) and southern European reconstructions (e.g. La Roya) show similar temperature variations for
406 the onset of both GI-1 and the Holocene. More temperature reconstructions, especially from southern and
407 Mediterranean regions of Western Europe, are necessary to confirm this preliminary observation.

408 Another latitudinal difference is related to the character of the onset of GI-1, which is generally more
409 abrupt in northern records and more gradual in southern ones, including speleothem records (e.g. El Pindal
410 cave, Fig. 4). This observation is also apparent when several Iberian records (see Moreno et al., 2012) and
411 references therein) and speleothem records from southern France and northern Africa (Genty et al., 2006)
412 are compared to the NGRIP ice core. This different character of the onset of GI-1 cannot be explained solely
413 by differences in the accuracy of the age models. In fact, the chronology for speleothems from southern
414 France and northern Spain is sufficiently well-constrained to observe this gradual onset (see dates in ESM).
415 Additionally, a North–South gradient through GI-1 is apparent in which GI-1a is warmer than GI-1e in the
416 south but cooler in the north (black lines in Fig. 4). The warming trend during GI-1 was previously observed
417 in sea surface temperature reconstructions from Western Mediterranean marine cores (Cacho et al., 2001)
418 and in western Mediterranean temperate forest development (Combourieu Nebout et al., 2009). In spite of
419 this compelling evidence, there is still no clear explanation for the causes of these different patterns.

420 However, all hypotheses suggest a close link to the North Atlantic ocean, atmospheric circulation patterns,
421 Scandinavian ice sheet retreat and sea ice cover (Genty et al., 2006; Lane et al., 2013).

422 Model results of the 14.7 ka BP transition (GS-2.1-a/GI-1e, see Rasmussen et al., this issue) indicate that the
423 main temperature changes occurred in winter (Renssen and Isarin, 2001). More importantly, the model
424 simulations demonstrate that the largest change in atmospheric variability occurred in NW Europe where
425 the daily temperature variability decreased by 50%, whereas storm activity also decreased substantially
426 (Renssen and Bogaart, 2003). These changes are related to a northward shift of the Atlantic sea-ice margin
427 at 14.7 ka BP, leading to a northward relocation of Atlantic storm tracks. In such a situation we should
428 expect a more marked and abrupt temperature change in northern European records, which were more
429 directly affected by the position of the Atlantic storm tracks. Southernmost European records were located
430 south of the winter sea-ice margin throughout the deglacial interval (which was located at ca. 45°N in GS-
431 2.1-a, (Renssen and Isarin, 2001), and did not experience such an abrupt change. In fact, the latitude ca. 40-
432 43°N has been considered a “frontier” between two different responses in terrestrial climate to climate
433 change during the Holocene (Magny et al., 2013b). These modelling results explain the difference in abrupt
434 or gradual onset of GI-1 on a N-S transect in Western Europe, whereas the observed differences in the
435 pattern within the last deglaciation remain unexplained.

436 Besides the long-term trends within GI-1, some of the compiled records also indicate synchronous
437 temperature or precipitation changes within the short events GI-1a to GI-1e. Temperature and
438 precipitation decreases are indicated in lake records such as Ammersee, Gerzensee, Hijkermeer and in
439 speleothems from Chauvet and El Pindal caves during the cold GI-1b, GI-1c-2 and GI-1d intervals (Fig. 4).
440 Furthermore, recent analyses of pollen records in southern Iberian marine cores indicate short-lived
441 intervals of forest decline consistent with cooling and drying during GI-1d and GI-1b (Combourieu Nebout
442 et al., 2009; Fletcher et al., 2010b). Sometimes limited sampling resolution and the lack of a precise
443 chronology have prevented the identification of short-term events. Therefore, a comprehensive
444 comparison of terrestrial responses to these short climate oscillations is not yet possible.

445 *5.2. Western Europe climate reconstruction before the LGM*

446 Palaeoclimatic reconstructions predating the LGM (ca. 23 – 19 ka BP, (Mix et al., 2001) offer challenges that
447 are less often encountered when studying younger material. For instance, with increasing age,
448 chronological control on sedimentary sequences (or other records) tends to worsen. Second, non-analogue
449 situations between present and past conditions are more likely to occur, hampering quantitative
450 reconstruction of key climatic parameters.

451 Sites that occur north of ~50°N latitude were either covered by the Scandinavian Ice Sheet (SIS) during (part
452 of) MIS3 and MIS2, or they were situated in a periglacial environment, which prevented formation of
453 continuous records (Engels et al., 2008a). South of the sub- and periglacial region, many sites provide
454 records throughout large parts of MIS2 and 3 (Fletcher et al., 2010a; Moreno et al., 2012). Below, we
455 discuss both regions of W Europe in more detail.

456 *5.2.1 NW Europe*

457 Dating is often problematic in sedimentary records from Scandinavia and other parts of the (peri-) glacial
458 environment, for the period predating the LGM. Subglacial erosion removed large sections of sedimentary
459 records in Scandinavia, and only discontinuous or fragmentary records are preserved in scattered locations
460 in NW Europe. The periglacial environment was probably too dry for the (continuous) accumulation of
461 organic matter, and only during warm periods (interstadials) did the SIS retreat, allowing lake or peat
462 deposits to accumulate. Therefore, lithostratigraphic correlation and palynostratigraphy are often not
463 possible in this region. Second, the older part of MIS-3 is at the age limit of radiocarbon dating. Until the
464 recent publication of the age-depth model for the Lake Suigetsu 06 core sequence (SG06; (Ramsey et al.,
465 2012) it was not possible to calibrate radiocarbon ages older than ~26 cal ka BP. Although the cumulative
466 counting error of the SG06 sequence still reaches ~6%, the availability of a calibration curve for the period
467 prior to 26 ka BP allows us for the first time to calibrate terrestrial radiocarbon samples and to correlate
468 key climate records from MIS2 and 3 directly. Currently, the chronologies of almost all NW European sites

469 dated to MIS3 show major differences in both the applied dating methodology and in the number of dates
470 available per site, and it is often not clear during which time interval (i.e. which Greenland Stadial or
471 Interstadial) the deposit was formed. Combined, this hampers a detailed comparison of the available
472 paleoclimatological data for MIS2 and 3 in NW and central Europe.

473 The influence of the SIS is dominant in the formation and preservation of natural archives in the northern
474 parts of W Europe. Following the last major expansions of the SIS, only a few palaeoclimate records remain
475 in Scandinavia that predate the LGM time-interval. Furthermore, there is a bias towards preservation of
476 sediments formed during the (long) MIS3 interstadials. The SIS decreased in size during the prolonged
477 warm intervals of early MIS3, and the increase in temperature and moisture availability led to lake and
478 peatland development.

479 Most deposits studied from NW Europe were formed in shallow lacustrine environment. Dating for classic
480 sites such as Oerel (Behre and van der Plicht, 1992), Upton Warren (Coope et al., 1961) or Denekamp,
481 Hengelo and Moershoofd (van Der Hammen and Wijmstra, 1971) mostly relied on a few radiocarbon dates
482 as well as on palynostratigraphy. Based on the differences in the radiocarbon ages as well as in the pollen-
483 assemblages, the names of these sites were assigned to different interstadials of the Middle Pleniglacial.
484 These interstadials were later tentatively correlated with the Greenland Interstadials (Bos et al., 2001).
485 Proxies that were studied for the NW European sites mostly include pollen, non-pollen palynomorphs and
486 botanical and zoological macroremains. Using the plant-indicator approach, mean temperatures for the
487 warmest month have been reconstructed for some of these sites, resulting in reconstructed temperatures
488 that are close to present-day temperatures (see Huijzer and Vandenberghe, 1998, for a review of 268 NW
489 European Pleniglacial sites). Coleoptera-based temperature reconstructions of 15-18 °C (Coope, 2002)
490 showed that temperatures during early MIS3 might have been (sub-) equal to present-day conditions
491 (Huijzer and Vandenberghe, 1998; Behre and van der Plicht, 1992; Behre et al., 2005).

492 Sokli (N Finland) has been the subject of one of the most detailed MIS3 multi-proxy projects in N Europe,
493 and some of the results of this research project are incorporated in the database presented in this paper
494 (ESM). The entire sediment column has been dated using a combination of radiocarbon dating, OSL dating,
495 lithostratigraphic correlation and palynostratigraphy (Alexanderson et al., 2008; Helmens et al., 2007a,
496 2007b). The MIS3 sediments of the Sokli sequence have been analysed for pollen (Bos et al., 2009),
497 chironomids (Engels et al., 2008a), diatoms (Helmens et al., 2009), botanical macrofossils (Bos et al., 2009)
498 and sedimentological properties (Alexanderson et al., 2008; Helmens et al., 2007a, 2007b). Mean July air
499 temperatures have been derived from pollen and chironomids (Bos et al., 2009; Engels et al., 2008a;
500 Helmens et al., 2007a) while minimum mean July air temperatures have been reconstructed from aquatic
501 plant remains using the indicator species approach. The combined results show previously unknown rapid
502 climate warming to present-day temperatures in the early MIS-3.

503 Other recently studied MIS3 sites from NW Europe include Pilgrimstad (Wohlfarth et al., 2011) and
504 Riipiharju (Hättestrand and Robertsson, 2010) in Sweden and Hitura (Salonen et al., 2008), Ruunaa (Lunkka
505 et al., 2008) in Finland and Nochten (Engels et al., 2008b) in Germany. A combination of different dating
506 techniques (OSL, radiocarbon) and a multi-proxy approach to palaeoclimatic reconstruction were used. The
507 combined evidence of open-water conditions over different areas of the Fennoscandian region during early
508 MIS3 suggests a significantly decreased size of the SIS, tentatively correlated to the longest Greenland
509 Interstadial (GIS14) by (Helmens and Engels, 2010) and placed at ~49.6-54.2 ka BP (Rasmussen et al., this
510 issue). Other records from the Austrian and Swiss regions in the Alps covering the pre-LGM are reviewed in
511 Heiri et al. (this issue).

512 *5.2.2. SW Europe*

513 Pioneering palynological research into lake basins and deep bog sequences (Beaulieu and Reille, 1984;
514 Florschütz et al., 1971; Follieri et al., 1989) established that southern European archives can provide
515 continuous records of terrestrial environmental and climatic change spanning the last glacial-interglacial

516 cycle, thus providing insights into regional conditions and climatic variability for the intervals equivalent to
517 MIS3 and 2. In general, the proxies examined in SW European terrestrial archives have highlighted the
518 dynamic nature of glacial climate over southern Europe. The early view of a small number of southern
519 European glacial interstadials has now evolved into the recognition of pervasive millennial-scale variability
520 as temporal resolution, chronological control and range of proxies for records has improved (e.g. (Allen et
521 al., 1999; Follieri et al., 1998; Sánchez-Goñi et al., 2002; Wohlfarth et al., 2008) (Fig. 5). The record of both
522 terrestrial and marine climate tracers in marine sediments has provided important insights into rapid glacial
523 changes in SW Europe, notably the confirmation of in-phase changes between terrestrial and marine
524 paleoenvironments in response to Dansgaard-Oeschger and Heinrich variability (e.g. (Combourieu Nebout
525 et al., 2009; Roucoux et al., 2001; Sánchez-Goñi et al., 2000; Turon et al., 2003). However, the timing of key
526 changes in lake and marine archives is typically constrained by radiocarbon dating, with the inherent
527 challenge of increased uncertainty beyond the LGM and towards the limits of this dating technique. In
528 some cases, chronologies were based on event stratigraphy and the alignment of sub-orbital and
529 millennial-scale events to master ice-core chronologies, provoking debate within the Quaternary
530 community (Blaauw, 2011). Tephrochronology appears as a highly effective chronostratigraphic tool for
531 correlating long sedimentary sequences within and beyond the Mediterranean region (Lane et al., 2011b)
532 (see section 3), excluding the Iberian Peninsula and Atlantic façade.

533 Speleothem records from SW Europe provide excellent information about past climate conditions beyond
534 the LGM, although few records are available (Villars and Chauvet in France and El Pindal in Spain) (Genty et
535 al., 2010b, 2003a; Moreno et al., 2010a; Wainer et al., 2009) (see also Heiri et al., this volume for Alpine
536 speleothems) and the fact that they do not provide quantitative reconstructions. The Chauvet and El Pindal
537 records overlap by just a few millennia (25-22 ka, Fig. 5) and, despite their different resolution, both
538 indicate a synchronous warming trend after Heinrich event 2, coinciding with GI-2.1 and 2.2. The main
539 feature of the Villars record is a general cooling trend between GIS12 at ~45.5 ka and the synchronous
540 ending of three stalagmites at 30 ± 1 ka (Fig. 5), with an initial well-marked climatic threshold at ~41 ka, after

541 which the growth rate of each stalagmite slows down significantly (Genty et al., 2010b). Due to difficulties
542 interpreting $\delta^{18}\text{O}$ records from temperate latitudes, $\delta^{13}\text{C}$ profiles are more often employed to infer past
543 changes in moisture and soil microbial activity. High speleothem $\delta^{13}\text{C}$ values attained during GS indicate dry
544 climates on land (Fig. 5). Similarly, biome reconstruction from Lago Grande di Monticchio (Allen et al.,
545 1999) and AP:NAP (arboreal pollen to non-arboreal pollen ratios) from a marine core in the Alboran Sea
546 (Fletcher and Sánchez Goñi, 2008; Sánchez-Goñi et al., 2002) also indicate dry conditions (Fig.5).

547 Finally, the combined evidence from multiple studies indicates a characteristic pattern of climatic impacts
548 associated with Dansgaard-Oeschger variability during MIS3, with dry and cool (warm and humid)
549 conditions in southern Europe associated with cold stadial (warm interstadial) conditions at higher
550 latitudes. A few quantitative climate reconstructions have been performed, for example, using pollen
551 climate response surfaces at Lago Grande di Monticchio, Italy (Allen et al., 1999), or MAT reconstructions of
552 pollen records from marine sediments (e.g. Sánchez-Goñi et al., 2002). Clearly, there is a need for further
553 quantitative climate reconstructions incorporating a range of reconstruction techniques and proxies.

554 **7. Suggestions for future studies**

555 Considering the large number of Western European terrestrial palaeo-studies available in the literature, the
556 relatively low number of records included in the present study reflects the difficulties related to obtaining
557 quantitative climate proxies from terrestrial archives that are chronologically well-constrained (see Brauer
558 et al., this issue). Based on the experience of making the present compilation, we have several suggestions
559 on how future studies can provide the wider palaeo-community with more high-quality terrestrial records.

560 *Geographical key areas of interest for future investigations within the terrestrial palaeo-community.* The
561 majority of existing Western European palaeoclimate records providing quantitative climate information
562 are constrained to a relatively narrow latitudinal band from 53 to 45°N and records from Scandinavia and
563 the Mediterranean are relatively scarce (Fig. 1). In addition, a high proportion of the investigated sites is
564 located in mountainous areas. This uneven coverage of sites is not only geographical but also temporal, as

565 many high latitude records are associated with long interstadials, such as GI-14. There is therefore a need
566 for more palaeoclimate reconstructions from the high latitudes (e.g. lacustrine records in Sweden) and
567 from lowland areas in SW Europe (e.g. lowlands in France, Spain and Italy). Quantitative reconstructions
568 covering the time interval before the LGM are very scarce in W Europe, limiting our knowledge of climate
569 variability associated with DO changes.

570 *Inclusion of additional archives.* The present compilation consists of records from lakes, speleothems, ice
571 cores, and terrestrial proxies in marine records. However, records of loess sequences, tree rings, and glacier
572 geomorphology also constitute important contributions to the quantitative reconstruction of past climate
573 conditions. Unfortunately, the chronological framework in many of these records is not well-developed or
574 they are difficult to interpret in terms of climate variability. There is, however, potential for these archives
575 to produce important climate reconstructions in the future.

576 *Proxies: towards better quantifying past climate.* This compilation has shown an unbalanced use of proxies
577 to calibrate past climate conditions. Climate inferences from chironomid and pollen data follow consistent
578 approaches and methods, whereas these approaches are more difficult in other proxies due to (1) the lack
579 of adequate calibration data and numerical inference models and (2) the paucity of information on how
580 climate influences the distribution and abundance of the proxies. More work is needed to increase the
581 number of regional training sets (chironomids, pollen, diatoms) and to understand taphonomic processes
582 and the response of proxies to climatic change. In addition, the recent development of new proxies (and
583 methods) to produce quantitative reconstructions of past climatic conditions (e.g. combination of fluid
584 inclusions analyses with clumped isotope data on speleothems or the application of the TEX86 index in lake
585 sediments) provide exciting opportunities to improve our quantitative understanding of past climatic
586 changes.

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592

593 **Figures:**

594 **Fig. 1:** Location map showing the records compiled (pink: lake sediments; white: speleothems; blue:
595 terrestrial proxies on marine sediments). The Alpine region is enlarged to better indicate the included
596 records. See text and Table 1 for more information on the archives and proxies compiled in this study.

597 **Fig. 2:** Temporal coverage (base and top) of all records included in this compilation. Gray bands indicate
598 Greenland Interstadials (GIs) and pink bands the position of marine expression of Heinrich events.

599 **Fig. 3:** Map of main tephra occurrences in the Western European terrestrial palaeoclimate records
600 mentioned in the text.

601 **Fig. 4:** Selected high-resolution terrestrial records from Western Europe from 16 to 8 ka BP. From top to
602 bottom: NGRIP ice core (Rasmussen et al., 2006); Jansvatnet (Birks et al., 2012), Krakenes (Brooks and Birks,
603 2001), Nadourcan (Watson et al., 2010), Fiddaun (van Asch and Hoek, 2012), Hijkermeer (Heiri et al., 2011),
604 Klein Ven (van Asch et al., 2013), Ammersee (von Grafenstein et al., 1999), Egelsee (Larocque-Tobler, 2010),
605 Gerzensee (Lotter et al., 2000, 2012) (including temperature reconstruction from chironomids and pollen)
606 and Lago Picolo di Avigliana (Larocque and Finsinger, 2008) lakes; Savi (Frisia et al., 2005), Chauvet (Genty
607 et al., 2006), and El Pindal (Moreno et al., 2010a) caves and La Roya (Muñoz Sobrino et al., 2013) and
608 Monticchio lakes (Allen et al., 1999) (biome reconstruction derived from pollen data, indicating higher
609 temperature and/or precipitation as the number increases). Latitudinal position of the selected sequences
610 and the type of climate variable reconstructed are indicated. Shaded bands mark GS-1 and GS-2.1a
611 intervals, based on the NGRIP ice core (Rasmussen et al., this issue). Black lines along GI-1 show the
612 warming or cooling character of the different records.

613 **Fig 5:** Selected terrestrial records from Southwestern Europe from 60 to 20 ka BP compared to the NGRIP
614 ice core record. From top to bottom: $\delta^{18}\text{O}$ (‰) from the NGRIP ice core (Rasmussen et al., 2006); $\delta^{13}\text{C}$ (‰)
615 from Chauvet (Genty et al., 2006), Villars (Genty et al., 2010b) and El Pindal (Moreno et al., 2010a) caves,
616 biome reconstruction from Monticchio lake derived from pollen data (Allen et al., 1999) and arboreal

617 pollen to non arboreal pollen ratios from marine core MD95-2043 (Fletcher and Sánchez Goñi, 2008;
618 Sánchez-Goñi et al., 2002). Latitudinal position of the selected sequences is indicated. Shaded bands mark
619 warm GIs, based on the NGRIP record (Rasmussen et al., this issue).

620 **Tables:**

621 **Table 1.** List of compiled records indicating the type of dating, the climate variable that was reconstructed
622 and the proxy data that were quantified. Reference to published data is indicated.

623 **Table 2.** Advantages and limitations in the palaeoclimate reconstruction from the most common terrestrial
624 archives.

625 **Electronic Supplementary Material:**

626 Information about methods to calibrate proxies in order to obtain quantitative data about temperature,
627 precipitation or biomes. All 56 records are included in a standardized format.

628

Table 1. List of compiled records from Western Europe in the INTIMATE chronological framework, specifying the type of dating, the climate variable that was reconstructed and the proxy data that were quantified. Reference to published data is indicated for more information.

| Nº | Archive | Country | Site | Altitude (m above or below sea level) | Type of dating | Reconstructed variable | Quantified proxy | Reference |
|----|----------------|---------|-----------------------------|---|-------------------------------------|----------------------------|---------------------------|--|
| 1 | Lake sediments | Finland | Sokli | 220 | C-14, palynostratigraphy, OSL | Temperature | Chironomids | (Engels et al., 2008a) |
| 2 | | France | Lautrey | 788 | C-14, tephra | Temperature, lake level | Lithology, chironomids | (Heiri and Millet, 2005; Magny et al., 2006) |
| 3 | | | Les Echets | 267 | C-14, tephra | Temperature | Diatoms | (Ampel et al., 2010, 2009, 2008; Veres et al., 2009; Wohlfarth et al., 2008) |
| 4 | | Germany | Eifel | 565 | C-14 | Dust supply | Lightness, dust | (Seelos et al., 2009; Sirocko et al., 2013) |
| 5 | | | Friedländer Grosse Wiese | 10 | C-14, events, tephra | Temperature | Chironomids | (van Asch et al., 2012) |
| 6 | | | Holzmaar | 425 | Varve counting, | Temperature, | Pollen | (Litt et al., 2009) |

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| | | | | tephras, C-14 | precipitation | | |
| 7 | | Meerfelder Maar | 336.5 | Varve counting, tephras, C-14 | Temperature, precipitation | Pollen | (Litt et al., 2009) |
| 8 | | Rehwiese | 28 | Varve counting, tephra | Temperature | Calcite layer thickness | (Neugebauer et al., 2012) |
| 9 | | Ammersee | 533 | C-14, tephra | Isotopic composition of precipitation | Ostracods $\delta^{18}\text{O}$ | (von Grafenstein et al., 1999) |
| 10 | Ireland | Fiddaun | 30 | C-14, events | Temperature | Chironomids | (van Asch and Hoek, 2012) |
| 11 | | Lough Nadourcan | 70 | C-14 | Temperature | Chironomids | (Watson et al., 2010) |
| 12 | Italy | Accesa | 157 | C-14, tephra | Temperature, precipitation, lake level | Lithology, pollen | (Magny, 2006; Magny et al., 2007) |
| 13 | | Frassino | 78 | C-14 | Precipitation | Bivalves $\delta^{18}\text{O}$ | (Baroni et al., 2006) |
| 14 | | Preola | 4 | C-14 | Lake level | Lithology | (Magny et al., 2011) |
| 15 | | Lavarone | 1100 | C-14, palynostratigraphy | Temperature | Chironomids | (Heiri et al., 2007) |
| 16 | | Piccolo di | 365 | C-14, tephra | Temperature | Chironomids | (Larocque and Finsinger, |

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| | | | Avigliana | | | | | 2008) |
| 17 | | | Ledro | 652 | C-14 | Temperature, precipitation, lake level | Lithology, pollen | (Joannin et al., 2013; Magny et al., 2009; Vanni re et al., 2013) |
| 18 | | | Grande di Monticchio | 656 | varve dating, C-14, tephras | Biomes | Pollen | (Allen et al., 1999) |
| 19 | | Netherlands | Hijkermeer | 14 | Palynostratigraphy | Temperature | Chironomids | (Heiri et al., 2011) |
| 20 | Klein Ven | | 28.5 | C-14, palynostratigraphy | Temperature | Chironomids | (van Asch et al., 2013) | |
| 21 | Usselo | | 33 | C-14, palynostratigraphy | Temperature | Coleoptera | (van Geel et al., 1989) | |
| 22 | | Norway | Jansvatnet | 53 | C-14 | Temperature | Chironomids | (Birks et al., 2012) |
| 23 | | | Kr kenes | 38 | C-14 | Temperature | Chironomids | (Brooks and Birks, 2001) |
| 24 | | Spain | Banyoles | 173 | C-14, U-Th | Vegetation cover, Precipitation | Calcite $\delta^{18}\text{O}$, $\delta^{13}\text{C}$ | (P rez-Obiol and Juli , 1994; Valero-Garc s et al., 1998) |
| 25 | | | Enol | 1070 | C-14 | Temperature | Calcium content | (Moreno et al., 2010b) |
| 26 | | | Estanya | 670 | C-14 | Lake level | Lithology | (Morell n et al., 2009) |
| 27 | | | Fuenteillejo | 635 | C-14, | Lake level | Lithology | (Vegas et al., 2010) |

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| | | | | | palaeomagnetism | | | | |
| 28 | | | Portalet | 1608 | C-14 | Lake level | Lithology | (González-Sampériz et al., 2006) | |
| 29 | | | La Roya | 1802 | C-14 | Temperature | Chironomids | (Muñoz Sobrino et al., 2013) | |
| 30 | | Switzerland | Egelsee | 770 | C-14 | Temperature | Chironomids | (Larocque-Tobler, 2010; Wehrli et al., 2007) | |
| 31 | | | Foppe | 1470 | C-14 | Temperature | Chironomids | (Samartin et al., 2012b) | |
| 32 | | | Gerzensee | 603 | isotope stratigraphy, palynostratigraphy, tephra | Temperature, Precipitation, lake level | Chironomids, pollen, lithology, cladocera | (Lotter et al., 2000, 2012; Magny, 2012; van Raden et al., 2012) | |
| 33 | | | Hinterburgsee | 1515 | C-14 | Temperature | Chironomids | (Heiri et al., 2003) | |
| 34 | | | Origlio | 416 | C-14 | Temperature | Chironomids | (Samartin et al., 2012a) | |
| 35 | | | Maloja | 1815 | C-14 | Temperature | Chironomids | (Ilyashuk et al., 2009) | |
| 36 | | | UK | Abernethy forest | 340 | C-14 | Temperature | Chironomids | (Brooks et al., 2012) |
| 37 | | | | Gransmoor | 8 | C-14 | Temperature | Coleoptera | (Blockley et al., 2004; Elias |

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| 38 | | | Llanilid | 60 | C-14 | Temperature | Coleoptera | (Blockley et al., 2004; Elias and Matthews, 2013; Walker et al., 2003) |
| 39 | Speleothems | France | Chauvet- Chastm6 | 205 | U-Th | Soil and vegetation, precipitation | Calcite $\delta^{18}\text{O}$, $\delta^{13}\text{C}$ | (Genty et al., 2006) |
| 40 | | | Villars-11 | 175 | U-Th | Soil and vegetation, precipitation | Calcite $\delta^{18}\text{O}$, $\delta^{13}\text{C}$ | (Genty et al., 2006) |
| 41 | | | Villars-14 | 175 | U-Th | Soil and vegetation, precipitation | Calcite $\delta^{18}\text{O}$, $\delta^{13}\text{C}$ | (Genty et al., 2010a; Wainer et al., 2009) |
| 42 | | | Villars-27 | 175 | U-Th | Soil and vegetation, precipitation | Calcite $\delta^{18}\text{O}$, $\delta^{13}\text{C}$ | (Genty et al., 2010a) |
| 43 | | | Villars-9 | 175 | U-Th | Soil and vegetation, precipitation | Calcite $\delta^{18}\text{O}$, $\delta^{13}\text{C}$ | (Genty et al., 2003b) |
| 44 | | | Ireland | Cragg | 80 | U-Th | Soil and vegetation, temperature, precipitation | Calcite $\delta^{18}\text{O}$, $\delta^{13}\text{C}$ |
| 45 | | Italy | Corchia | 840 | U-Th | Precipitation | Calcite $\delta^{18}\text{O}$, | (Zanchetta et al., 2007) |

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| | | | | | | | $\delta^{13}\text{C}$ | |
| 46 | | | Grotta Savi | 441 | U-Th | Temperature | Calcite $\delta^{18}\text{O}$, $\delta^{13}\text{C}$ | (Frisia et al., 2005) |
| 47 | | Spain | El Pindal cave | 24 | U-Th | Soil and vegetation, precipitation | Calcite $\delta^{18}\text{O}$, $\delta^{13}\text{C}$ | (Moreno et al., 2010a) |
| 48 | Ice core | Greenland | NGRIP | 2917 | layer counting | Temperature | $\delta^{18}\text{O}$ isotopes | (Andersen et al., 2004) |
| 49 | Marine (pollen) | Alboran Sea | ODP-906 | -1841 | C-14, events | Temperature, precipitation | Pollen data | (Combourieu Nebout et al., 2009; Fletcher et al., 2010a; Genty et al., 2010b) |
| 50 | | | MD95-2043 | -1108 | C-14, events | Vegetation | Pollen data | (Fletcher and Sánchez Goñi, 2008; Sánchez-Goñi et al., 2002) |
| 51 | | Iberian margin | MD03-2697 | -2164 | C-14 | Vegetation | Pollen data | (Naughton et al., 2007) |
| 52 | | | MD95-2039 | -3381 | Events | Vegetation | Pollen data | (Roucoux et al., 2001) |
| 53 | | | MD95-2042 | -3148 | C-14, events | Vegetation | Pollen data | (Sánchez-Goñi et al., 2008, 2000, 1999, 2009) |
| 54 | | | MD99-2331 | -2110 | C-14, events | Vegetation | Pollen data | (Naughton et al., 2007) |
| 55 | | Western France | MD04-2845 | -4100 | C-14, events | Vegetation | Pollen data | (Sánchez-Goñi et al., 2008) |

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| 56 | | Adriatic Sea | MD90-917 | -1010 | C-14, events | Temperature, precipitation | Pollen data | (Combourieu-Nebout et al., 2013) |
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Table 2. Advantages and limitations in the palaeoclimate reconstruction from the most common terrestrial archives.

| Archive | Advantages | Limitations |
|-----------------------|---|--|
| Lake sediments | <p>High resolution, high sedimentation rates, even laminated or varved sediments</p> <p>Quantifying past temperatures is possible from different lacustrine organisms groups</p> <p>Reconstructing past lake levels can be a way to explore water availability in the past</p> <p>Some lake sediments contain tephra layers</p> | <p>Lack of continuity in dry or cold climates</p> <p>Dating is difficult beyond C-14 range or when sediments have low organic content</p> <p>Lacustrine sedimentation is tightly linked to local factors – not always a good representation of past regional climate</p> |
| Speleothems | <p>High resolution, annual bands</p> <p>Precise chronological control (U-Th) under most common conditions</p> | <p>Lack of continuity in dry and cold climates</p> <p>Some speleothems can not be dated due to low U in the samples or high detrital content</p> <p>Accurate interpretation of stable isotopes usually requires other proxies</p> |
| Ice cores | <p>High resolution, annual bands</p> <p>Precise chronological control (varve counting)</p> | <p>Only obtainable from ice sheets</p> <p>Difficult to obtain quantitative temperature reconstructions</p> |
| Loess | <p>Proxy for atmospheric intensity in the past</p> <p>Dating comes from the combination of ^{14}C and OSL techniques</p> | <p>Difficult to obtain continuous records</p> <p>Complicated stratigraphy</p> <p>Limited spatial coverage</p> |
| Tree rings | <p>High resolution, annual bands</p> <p>Precise chronological control (tree-ring</p> | <p>Quantification of climate from tree rings can be problematic</p> |

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| | counting and ^{14}C) | Scarce records covering INTIMATE chronological framework |
| Glacier deposits | Dating comes from the combination of different techniques (^{14}C , OSL, ^{10}Be , etc) | Difficult to obtain continuous records Difficult to obtain climate parameters in a quantitative way |
| Terrestrial tracers in marine sediments | Long, continuous archives Terrestrial information can be easily combined (under the same chronology) with marine data (SST, productivity, etc) | Source areas for terrestrial tracers may be poorly defined, and transport pathways may change over time and in response to climate changes |

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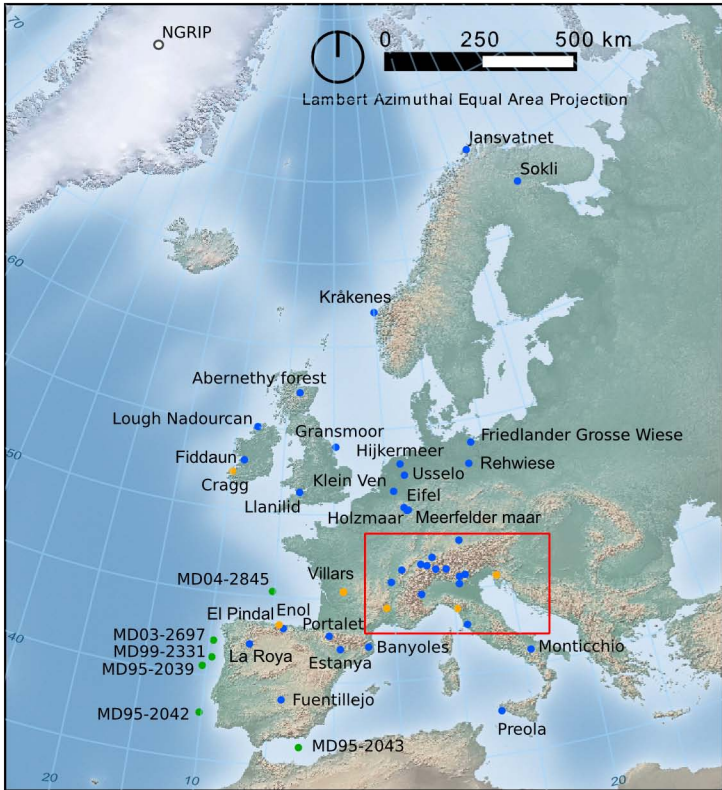
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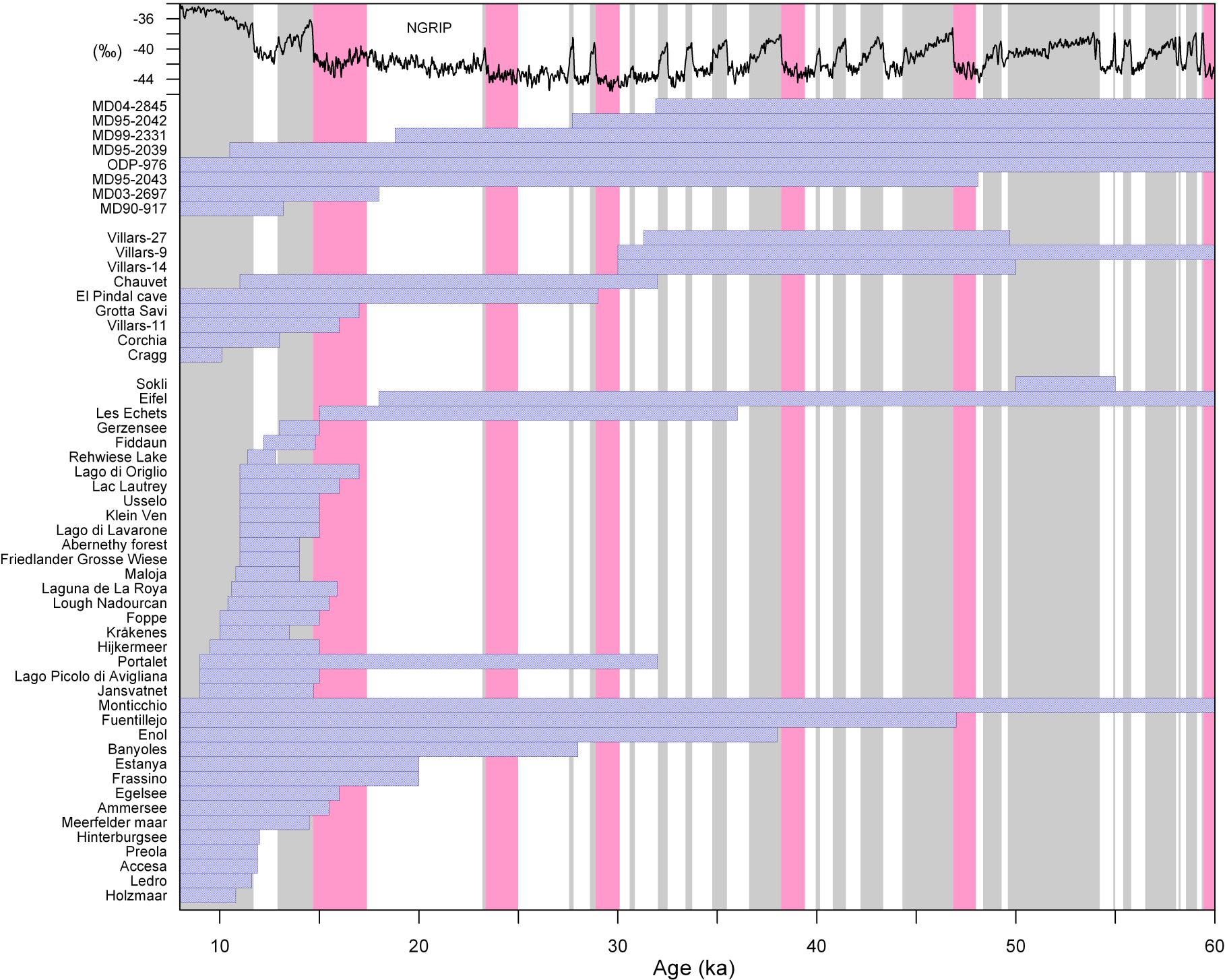
Legend

Sites (type)

- ice core
- lake
- marine (pollen)
- speleothem



Source base map:
NaturalEarthData.com





Legend:

Icelandic tephras:

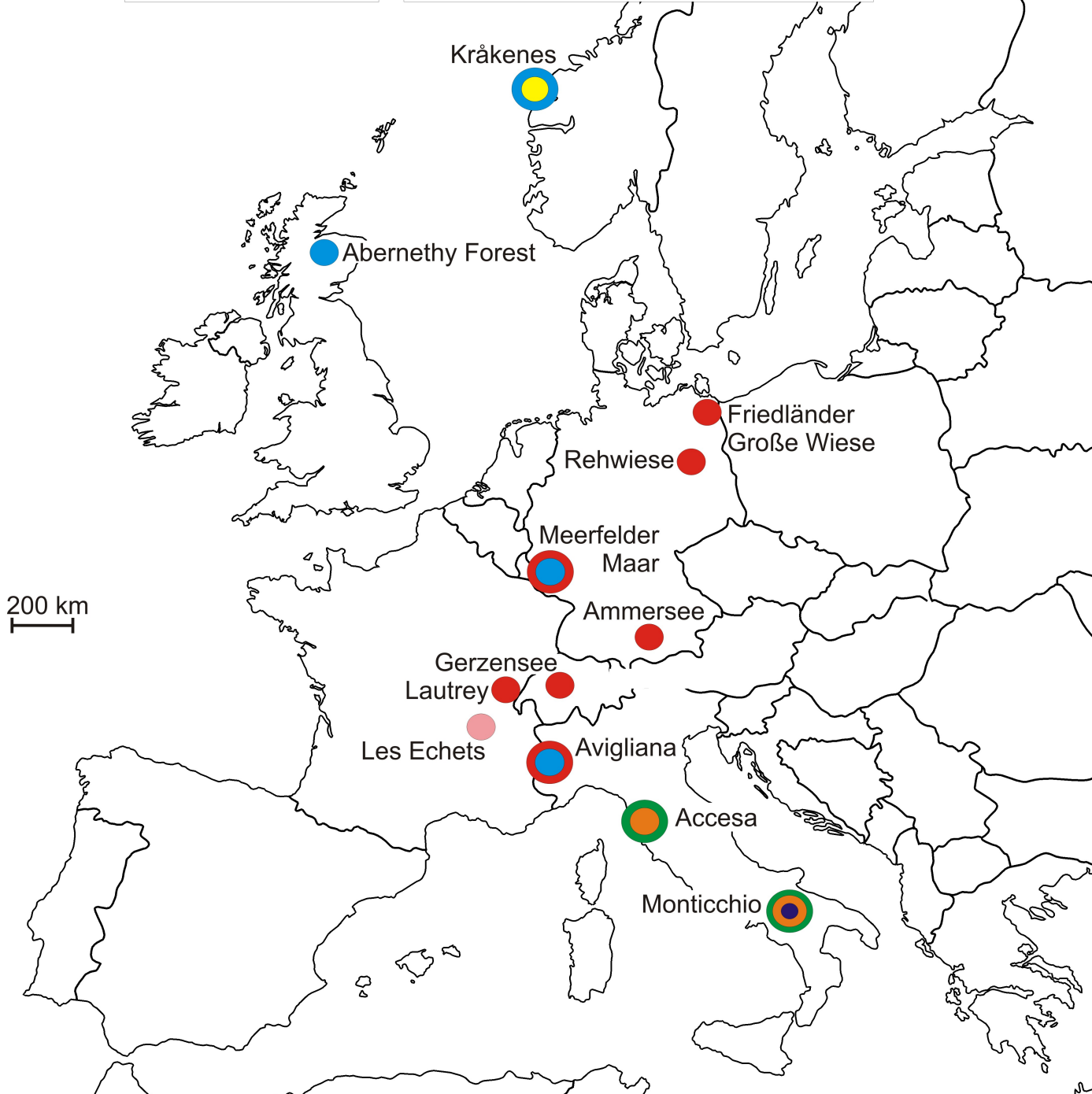
- Yellow circle: Saksurnarvatn ($10,296 \pm 45$ GICC05 yrs BP)
- Blue circle: Vedde Ash ($12,120 \pm 57$ GICC05 yrs BP)

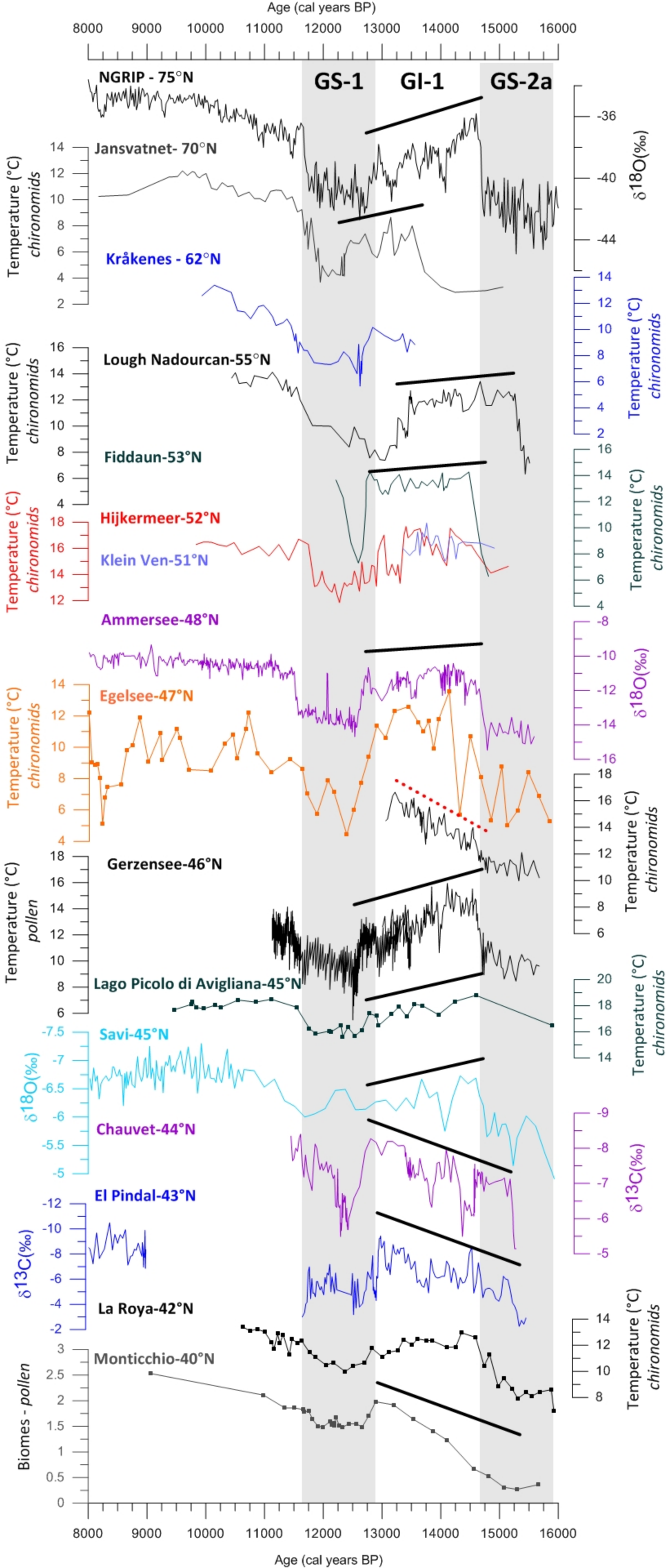
Eifel tephras:

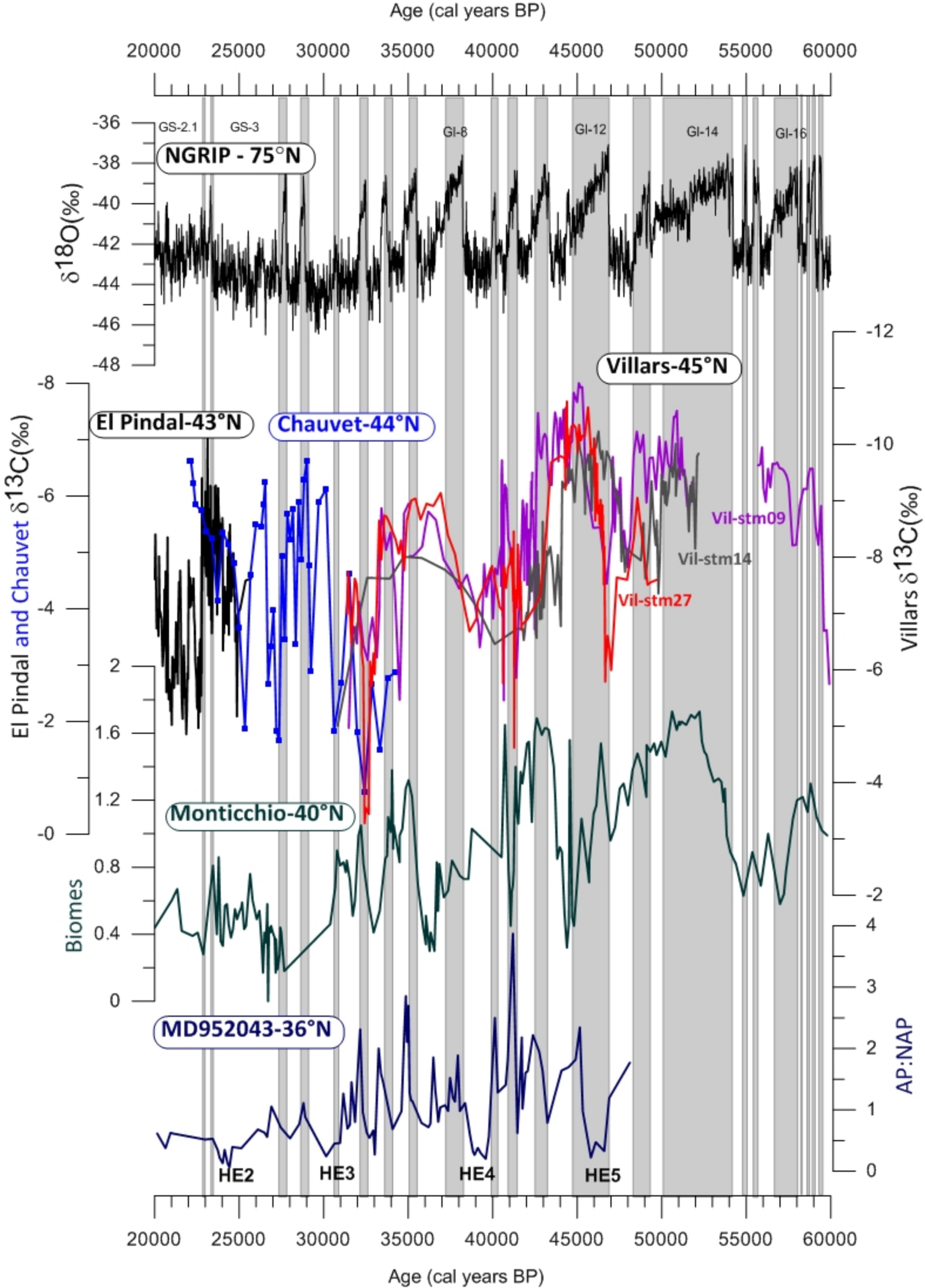
- Red circle: Laacher See Tephra ($12,880 \pm 40$ varve yrs BP)
- Pink circle: Les Echets Tephra (44 ± 2 ka BP)

Italian tephras:

- Orange circle: Avellino ($3,950 \pm 40$ cal yrs BP)
- Green circle: Pomici Principali ($12,180 \pm 210$ cal yrs BP)
- Purple circle: Neapolitan Yellow Tuff ($14,110 \pm 210$ cal yrs BP)







SUPPLEMENTARY MATERIAL

METHODS FOR PROXY CALIBRATION

1. Temperature

Most Western Europe temperature reconstructions in this compilation are based on chironomids (19 records from a total of 33 temperature reconstructions, Table 1). Chironomidae is a family of two-winged flies (Insecta: Diptera), also known as non-biting midges. The larvae of most species are aquatic and chironomids can be found in almost all types of freshwater. The head capsules of chironomid larvae are abundant and well-preserved in most lake sediments, although are less diverse and may be absent from the bottom of lakes that have long periods of anoxia. Individual fossils can usually be identified to genus or species-morphotype (Brooks et al., 2007). The winged adult chironomids are poor fliers but are widely and rapidly dispersed across the landscape by wind. Therefore the response of chironomids to climatic and environmental change is rapid, often within the sampling resolution of most sediment core analyses, so no lag is evident. The development rate of the immature stages of most chironomid species is primarily driven by summer temperature and so the distribution and abundance of each taxon is closely related to mean July air temperature (Eggermont and Heiri, 2012). This relationship is the basis for chironomid-based temperature inference models. These inference models typically have an error of about 1-1.5°C and an r^2 of around 0.8-9. Chironomid-based temperature inference models and modern calibration sets suitable for use on European lake sediment sequences are currently available from Scandinavia (e.g. Larocque et al., 2001), Switzerland (Lotter et al., 1997), Iceland (Langdon et al., 2008) western Russia (Self et al., 2011) and western Europe (Heiri et al., 2011). These models have been successfully applied to many sequences since the LGM to produce high resolution temperature inferences. The reliability of these temperature reconstructions has been confirmed by comparison to independently derived reconstructions based on other proxies, particularly oxygen isotopes

(Brooks and Birks, 2000), and for recent sequences, with meteorological records (Larocque et al., 2009). Because of the sensitivity of chironomids to climate change, chironomid-base reconstructions apparently reflect even the short-duration, low amplitude, climate oscillations seen in the Greenland ice core records (e.g. GI-1d or GI-1b, events) with little or no lag in the response (Brooks and Birks, 2001).

Three other groups of organisms have been used to reconstruct temperature in this compilation: diatoms, Coleoptera and Cladocera. Diatoms are a major group of algae, and are among the most common types of phytoplankton. Diatoms have played an important role in the development of quantitative methods for environmental reconstruction, not only for temperature reconstruction (e.g. Lotter et al., 1997) but also for pH variability (e.g. Birks et al., 1990) and salinity (e.g. Fritz et al., 1991). Many studies on latitudinal climatic gradients indicate the importance of a temperature relationship in diatoms (see references in Smol and Stoermer, 2010). In this compilation, the Les Echets temperature record (Table 1, Fig. 1) is based on diatoms and shows a clear differentiation among DO interstadials and stadials (Ampel et al., 2010; Wohlfarth et al., 2008). Coleoptera, or beetles, are one of the most species-rich groups of organisms. They are excellent proxies for climate change because of their good preservation, because their morphology and ecology does not appear to have changed through time, and because they respond quickly to climatic changes. Since the work by Atkinson et al., (1987), many other researchers have used Coleoptera to determine quantitative environmental changes (Coope et al., 1998; Ponel et al., 2007). However, due to the poor chronological control of most Western European beetle records, only three coleopteran-based reconstructions are included in this compilation: the Usselo record in the Netherlands (Van Geel et al., 1989) and Gransmoor and Llanilid records in UK (Elias and Matthews, 2013) (Table 1, Fig. 1). Cladocera is an order of small crustaceans that has been used by several authors to infer temperature from lake sediments (Smol et al., 2001, chap. 2). Lotter et al., (1997) made the first attempt to quantify the relationship between climate and

Cladocera assemblages and his model was later applied to obtain the Gerzensee temperature reconstruction (Lotter et al., 2000) (Table 1, Fig. 1).

Temperatures can be also reconstructed from Greenland ice core records. Thus, Greenland temperatures from the last glacial period are estimated by a combination of 1) the water isotopic composition of the ice ($\delta^{18}\text{O}$) which provides a high resolution temperature proxy (Johnsen et al., 2001), 2) accurate borehole temperature records that provide long term absolute temperatures of past precipitation (Dahl-Jensen et al., 1998), and 3) measurement of nitrogen isotopes in the gas phase ($\delta^{15}\text{N}$) to provide an estimate of abrupt climatic changes (Huber et al., 2006). During the coldest phases of the last glacial period temperatures in central Greenland were about 25°C colder than the present, and temperature increases associated with the onset of DO events are between 5-15 °C of amplitude (Wolff et al., 2010). Recently, an absolute temperature calibration for the entire NGRIP ice core has become available (Kindler et al., 2013).

Finally, although inferring qualitative temperature changes from sedimentological and geochemical data in lake sediments is common (e.g. González-Sampériz et al., 2006), quantification of those proxies is still far from complete. The main difficulty is the necessity to calibrate the sedimentological and geochemical properties for each lake sediment record. In this compilation, and in a semi-quantitative way, the Calcium record of Enol lake, in the Cantabrian mountains (Spain), is presented as a temperature proxy as carbonate is only preserved in that environment under warm climates (Moreno et al., 2010b). Similarly, the thickness of calcite layers in Rehwielse palaeolake (Germany) is a temperature indicator during GS-1 as those varves are formed during the warm season (Neugebauer et al., 2012). Some temperature reconstructions based on pollen records are also presented in this compilation (Table 1) and are discussed below (section 3).

2 Precipitation/effective moisture

2.1 Lake levels

Lakes are one of the best archives to record past changes in water availability, such a reconstruction is particularly reliable in closed lakes where lake level is controlled by the precipitation – evaporation balance and other inputs and outputs are negligible. However, quantifying the processes influencing lake level is not an easy task and depends on sedimentary facies that can be related to particular lake levels. At least two approaches have been used to describe variations in lake levels using sedimentological information. One approach, which is quantitative but not calibrated against present-day water level, is based on the assignation of a number to the sedimentary facies, indicating the lowest lake level by the lowest number. In some lakes with high lake level variation, the observed sedimentary facies can give interesting information about the deposition environment and allow estimates of relative lake level change. In the example from Lake Estanya, in the Spanish Pre-Pyrenees, deposition of yellowish, massive, coarse-grained gypsum is associated to an ephemeral saline lake–mud flat – lake level 1 – while black, massive to faintly laminated silty clay sediments were characteristics of a deep, dimictic, freshwater lake – lake level 10 (Morellón et al., 2009). The second approach was first applied to marls accumulated in the lakes of the Jura and Swiss Plateau and then applied in other places, such as the Italian Peninsula (Magny, 2012; Magny et al., 2007). This technique is based on counting and classifying the different carbonate concretions, morphotypes of biochemical origin that compose the coarser fraction (Magny, 2001), together with an assessment of the lithofacies changes. Modern analogues based on surface samples taken along transects perpendicular to the shore have revealed that each concretion morphotype shows a specific spatial distribution from the shore to the deep zone in relation to the hydrodynamics and the aquatic vegetation belts (e.g. *Oncolites* characterize nearshore areas; cauliflower-like forms dominate the littoral platform, etc.). Therefore, characterizing the different sub-environments in several cores taken on a depth transect, dating the observed transitions in all the cores and correlating them, allows the main changes

in lake level to be distinguished (e.g. Accesa lake, Lac Lautrey – see a recent compilation in Magny et al., 2013).

2.2. Stable isotopes as indicators of precipitation and hydrological changes

Determining the isotopic composition of past precipitation is usually a challenging task, with the notable exception of ice cores, where the solid precipitation (snow, then transformed into ice) can be measured almost directly from the ice and is often seen as a basic proxy for reconstructing past surface temperatures (see above). For other natural substances we have to consider that the primary isotopic composition of meteoric precipitant can be quite different from “environmental water” from which the proxy precipitates and the composition of the proxy is also influenced by other biological, physical and/or chemical processes. For instance, a body of water can be influenced by evaporation, which tends to enrich it in heavy isotopes compared to rainfall. In addition, the oxygen isotope composition of a carbonate precipitated from a water body is also influenced by the temperature and kinetic isotopic fractionations (e.g. Leng and Marshall, 2004).

2.2.1 Lake stable isotopes

Lake sediments supply different kinds of authigenic minerals like mollusc shells, ostracods, diatoms or biogenic-induced carbonate, which can be analyzed for stable isotopes for the reconstruction of past environmental conditions (e.g. Leng, 2006; Sagnotti et al., 2001). More recently, the isotopic composition of chironomids has been proposed as a potential tool to reconstruct past lake water isotopic compositions (Verbruggen et al., 2010). Unfortunately, the climatic information which can be obtained from isotopes is strongly skewed by the period of formation of the different substances and the characteristics of the water body such as its recharge system and residence time. However, careful selection of the appropriate lacustrine system together with the analysis of a specific carbonate fraction (i.e. ostracods), allows the

reconstruction of past variation in isotopic composition of lake waters which can be related to rainfall isotopic composition (von Grafenstein et al., 1999).

Most of the $\delta^{18}\text{O}$ records from Northwestern and Central European lakes are from the last deglaciation and (early) Holocene (e.g. Ahlberg et al., 1996; Böttger et al., 1998; Lotter et al., 1992; von Grafenstein et al., 1999; Hammarlund and Buchardt, 1996) and, unfortunately, many of them were obtained from bulk carbonate with the consequent risk of detrital contamination (Leng et al., 2010). The evolution of those $\delta^{18}\text{O}$ records during the lateglacial indicates an increasing trend at the beginning of GI-1 interstadial and, in some cases, a decreasing trend with a clear phase of lower values linked to climatic deterioration during GS-1. Lotter et al. (1992) have shown that not only major shifts in the oxygen-isotope profiles from central European lakes but also minor shifts are nearly synchronous with those recognized in the Greenland oxygen-isotope record. This is mostly related to the strong link between surface temperature and $\delta^{18}\text{O}$ of precipitation in these regions (e.g. Rozanski et al., 1993) as dominant factors in controlling the final isotopic composition of lacustrine carbonate. Additionally, in order to improve interpretation of $\delta^{18}\text{O}$ variability in lacustrine carbonates, the sensitivity of lakes to hydrological budget and evaporation must be taken into account as these effects can be dominant compared to the effects of temperature and amount of precipitation on the isotopic composition of rainfall. This is particularly evident from the water isotopic composition of Mediterranean lake water (Roberts et al., 2010, 2008) where last deglaciation $\delta^{18}\text{O}$ records are linked to changes in the hydrological balance (Baroni et al., 2006; Zanchetta et al., 2007a).

2.2.2 Speleothem isotopic composition

The majority of speleothem-based palaeoclimate studies have focused on $\delta^{18}\text{O}$ records as proxies of variability in the isotopic composition of rainfall (Lachniet et al., 2009). Compared to lakes, speleothem $\delta^{18}\text{O}$ records are less affected by evaporation in the soil-epikarst zone (at

least for the European records), thus facilitating their interpretation. However, the presence of non-equilibrium isotopic effects during calcite precipitation (Mickler et al., 2006) may mask part of the relationship with local rainfall, which is usually averaged in the isotopic composition of dripwaters. On the other hand, different time periods of calcite precipitation or water infiltration can skew the isotopic signal towards a specific season (Tremaine et al., 2011). The interpretation of speleothem $\delta^{18}\text{O}$ records differs regionally. Thus, in the Southern Europe-Western Mediterranean area the signal is often considered to be dominated by the amount of precipitation (Zanchetta et al., 2007 and references therein) impinging in the cave waters. In northern regions, such as the Alps, temperature is believed to play a dominant role (e.g. Frisia et al., 2005; Mangini et al., 2005), although changes in storm trajectories could also be important (Mangini et al., 2007; Spötl et al., 2010). Recently, McDermott et al., (2011) presented the first compilation of European speleothems $\delta^{18}\text{O}$ records of the Holocene, observing interesting latitudinal patterns with a tight isotope-latitude correlation which indicates that speleothems are reliable recorders of combined rainfall oxygen isotope composition and temperature. Unfortunately, a similar treatment of data for the last deglaciation is not still possible due to the scarcity of long records.

Besides $\delta^{18}\text{O}$ values, other proxies can shed light on regional climate variability reconstructed from speleothems. These proxies include $\delta^{13}\text{C}$ variability, which was shown to be very sensitive to climate changes and likely controlled by the vegetation and soil activity, thus a combination of moisture and temperature (Genty et al., 2003; Moreno et al., 2010a), trace element concentrations (Fairchild and Treble, 2009), and growth rates (Domínguez-Villar et al., 2009). Despite the fact that quantitative interpretation of these proxies is still problematic, isotopic measurements of fluid inclusion and on the clumped isotope measurements that provided the first temperature reconstructions combined with fluid inclusions data show great promise (Vanhof et al., 2006; Wainer et al., 2011).

3. Biome and climate reconstruction from pollen data

Biomes are the basic ecological units used to describe the major ecosystems of the Earth. Biomes are primarily distributed according to climatic factors (warmth of the growing season, winter temperature and available moisture) and are defined by the characteristic life-forms that inhabit each one. Biome reconstruction is essentially a means of classifying information about past vegetation, usually pollen data, into biomes. It recognises that vegetation in the past may have differed in species composition relative to present-day plant communities. Underpinning the approach is the assumption that biomes contain characteristic suites of plant functional types (PFTs). For example, desert vegetation usually includes many drought-tolerant forbs and very few broad-leafed trees, even though the species composition of desert vegetation varies enormously from region to region as a consequence of historical biogeography.

Biomisation was originally developed to validate biome maps based on climatic modelling experiments (Prentice et al., 1996; Prentice and Webb III, 1998). Its wide applicability in diverse environments has seen biome reconstruction employed in almost all parts of the world where pollen records exist (Elenga et al., 2000; Jolly et al., 1998; Pickett et al., 2004; Prentice et al., 1996; Tarasov et al., 2000, 1998). The resulting maps are currently the most spatially and temporally accurate depictions of past vegetation on a global scale. PFT-based approaches to pollen data (Peyron et al., 1998) may also be used in conjunction with climate–vegetation models to produce quantitative palaeoclimatic estimates (Guiot et al., 2000), although the broad tolerances of individual PFTs mean that the errors inherent in this procedure are large.

In European and Mediterranean regions, palaeoclimatic reconstructions have been produced on a continental scale at a coarse temporal resolution (e.g. Cheddadi and Bar-Hen, 2009; Cheddadi et al., 1997; Davis and Brewer, 2009; Davis et al., 2003; Huntley and Prentice, 1988; Peyron et al., 1998), while high-resolution pollen records from individual sites have been used to provide continuous palaeoclimatic records (e.g. Brewer et al., 2012; Guiot et al., 1989; Peyron et al., 2013, 2011). Quantitative palaeoclimatic estimates from pollen data are

produced using different approaches: since the early 20th century, there has been a development of methods for the estimation of climate from pollen assemblages, from the simple climate overlap methods of Iversen (1944) to more sophisticated probability-based approaches (Juggins and Birks, 2012).

Most standard methods are based on different ecological concepts. We can distinguish “assemblage-based” methods such as the Modern Analogues Technique (MAT), “species-based” methods such as Weighted Averaging (WA) or Weighted Averaging Partial Least Squares regression (WA/PLS), which are transfer functions, and probabilistic indicator taxa approaches such as Probability Density Functions (PDF). MAT reconstructs past climate parameters by considering the similarity of pollen assemblages (Guiot, 1990; Overpeck et al., 1985). The MAT uses squared-chord distance, to determine the degree of similarity between samples with known climate parameters (modern pollen samples) to a sample for which climate parameters are to be estimated (fossil pollen sample). The smaller the chord distance is, the greater the degree of analogy between the two samples. The palaeoclimate is reconstructed by weighted averaging. While the MAT does not require real calibration and is based on a comparison of past assemblages to modern pollen assemblages, regression-based transfer functions include on a calibration between environmental variables and modern pollen assemblages. WAPLS, an extension of simple Weighted Averaging, is a “classic” regression method that assumes that the relationships between pollen percentages and climate are unimodal: each taxon tends to be most abundant at a particular optimal value of an environmental variable and cannot survive where the value of that variable is too low or too high. Assuming that taxon’s ecological preferences remain constant in time, even if the overall pollen assemblage has no modern analogue, estimation for past environmental conditions can be provided.

Despite the broad agreement of pollen-based palaeoclimate reconstructions with other proxies, there are important issues surrounding their use (Salonen et al., 2012). Bartlein et al.

(2010) provide an assessment of statistical reconstruction methods and their application to fossil pollen data, highlighting the problems associated with 'no analogue' or "multiple analogue" situations (Birks et al., 2010) and spatial auto-correlation effects. Some problems are also noted with WA and WAPLS when the effective number of species in the fossil data-set is low (Telford and Birks, 2011). Recognition of these uncertainties had recently led to further analyse the potential advantages and limitations of the different methods (Guiot et al., 2009; Juggins, 2013; Telford and Birks, 2011, 2009, 2005). Recently, multi-method approaches are adopted which allows to better assess the error of reconstruction inherent in pollen-based climate prediction. Such approaches have been successfully tested for different time periods in Europe and Mediterranean areas and enable us to provide more precise and robust climate estimates than those based on only one method (Brewer et al., 2008; Klotz et al., 2003; Kühl et al., 2010; Peyron et al., 2013, 2011, 2005). Moreover, we find new methods such as Boosted Regression Trees (Salonen et al., 2012) or the recent improvement of the inverse modelling approach (Garreta et al., 2012) highly promising and well-performing tools for pollen-inferred palaeoclimatic reconstructions. Reliable palaeoclimate estimates are more and more needed as benchmarks for model simulations (Morrill et al., 2013).

4. Other climate variables

Besides temperature and precipitation, few other climate variables have been quantified based on terrestrial proxies from Western Europe sequences. It is worth highlighting the work carried out on some sequences to reconstruct past atmospheric circulation intensity from the dust content in the sediments. Thus, particles transported by wind were identified in several lakes from the Eifel region in northern Germany based on the grain-size, colour and geochemical composition (Sirocko et al., 2013) producing a summary curve that reflects the intensity of atmospheric circulation and dust availability for the last glacial cycle (electronic Supplementary material). Similarly, the content of dust particles in ice cores has served to discriminate periods with more material supplied from world deserts that have been

interpreted as arid periods but with an enhanced intensity of atmospheric winds (Fuhrer et al., 1999; Mayewski et al., 1997). A more intense atmospheric circulation linked to dry periods of the last glacial cycle (DO stadials) was also apparent in some studies carried out in marine sediments from Western Mediterranean where an increase in grain-size of detrital particles was observed (Bout-Roumazielles et al., 2007; Combourieu Nebout et al., 2002; Moreno et al., 2002). In any case, we are still far from being able to estimate the necessary wind intensity to account for the observed changes in the sediments in order to provide quantified data.

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