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1 **Dynamics and sources of last glacial aeolian deposition in southwest France derived**
2 **from dune pattern, grain-size gradients and geochemistry, and reconstruction of**
3 **efficient wind directions.**

4

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22

23 **Abstract**

24 Dune pattern, grain-size gradients and geochemistry were used to investigate the sources and
25 dynamics of aeolian deposition during the last glacial in southwest France. The coversands
26 form widespread fields of low-amplitude ridges (zibars), whereas Younger Dryas parabolic
27 dunes mainly concentrate in corridors and along rivers. Spatial modelling of grain-size
28 gradients combined with geochemical analysis points to a genetic relationship between
29 coversands and loess, the latter resulting primarily from dust produced by aeolian abrasion of
30 the coversands. The alluvium of the Garonne river provided also significant amounts of dust
31 at a more local scale. The geochemical composition of loess shows much lower scattering
32 than that of coversands, due to stronger homogenisation during transport in the atmosphere.
33 Overall, sandy loess and loess deposits decrease in thickness away from the coversands. Dune
34 orientation and grain-size gradients suggest that the efficient winds blew respectively from the

35 W to the NW during the glacial, and the W-SW during the Younger Dryas. A comparison
36 between the wind directions derived from the proxy data and those provided by
37 palaeoclimatic simulations suggests a change of the main transport season. Ground surface
38 conditions and their evolution throughout the year, i.e. the length of the season with snow and
39 frozen or moist topsoil, and the seasonal distribution of wind speeds able to cause deflation
40 are thought to have been the main factors that controlled the transport season in the study
41 area.

42

43 Key words: coversand, loess, Last Glacial, dunes, grain-size modelling, geochemistry, wind
44 direction, southwest France

45

46 1. Introduction

47

48 In the Aquitaine basin (Southwest France), Pleistocene coversands spread over a surface of
49 approximately 12,000 km², and are bordered on their eastern and southern margins by a large
50 loess belt. The stratigraphy of aeolian deposits has been the focus of recent studies and a large
51 set of ¹⁴C, ESR and OSL ages is available both for coversands and loess (Bertran et al., 2009,
52 2011; Hernandez et al., 2012; Sitzia et al., 2015). Mapping of aeolian deposits has also been
53 the subject of many attempts. The conventional field approach allowed precise delineation of
54 the coversands since the 60's (Enjalbert, 1960; Legigan, 1979), but the loess deposits were
55 always omitted in the maps, and are still lacking on the current 1/50,000 geological map of
56 France (*infoterre.brgm.fr*). Bertran et al. (2011) showed the potential of coupling field
57 mapping and the use of soil databases designed for agricultural purposes, in which physico-
58 chemical analyses of topsoil samples are listed. A first map of loess and transitional facies
59 was produced using the online BDAT database (*www.gissol.fr*) where the sand, silt and clay
60 content of topsoil samples are averaged at the scale of administrative units of an area of tens
61 of square kilometers. Since then, the Land Use and Cover Area frame Statistical survey
62 (LUCAS) database on topsoil properties in Europe has been made available and was used for
63 further mapping of aeolian deposits (Bertran et al., 2016). The points which satisfy the grain-
64 size criteria of coversands and loess were extracted from the rasters of predicted soil texture
65 established by kriging of the LUCAS data by Balabio et al. (2016). These rasters have a
66 resolution of 500 m. A comparison with previous maps showed a good fit in most of the
67 tested areas in Europe (Bertran et al., 2016). Improvement of the French database (Donesol,

68 <http://acklins.ortleans.inra.fr/outil/donesol>) and access to raw data makes it now possible to
69 investigate more in detail the aeolian deposits at a regional scale, since more refined
70 information on grain-size composition and on loess thickness is available.

71

72 Pioneer mineralogical studies have been performed by Klingebiel (1966) and Legigan (1979).
73 They have suggested that the sources of the coversands were mostly Mio-Pliocene alluvial
74 and deltaic sands that crop out on the continental plateau. However, these studies remain
75 crude, and the possible contribution of wind-blown particles derived from the alluvium fed by
76 the surrounding ranges (Massif Central, Pyrenees) was hardly detectable using heavy mineral
77 analyses alone. The provenance of the particles and the transport paths remain, therefore, to
78 be documented in detail. Grain-size gradients (Lautridou, 1985; Ruegg, 1983; Liu, 1988),
79 loess thickness (Frazee et al., 1970; Mason, 2001), and the mineralogical and chemical
80 composition of the deposits (Muhs et al., 2008; Stevens et al., 2010; Rousseau et al. 2014)
81 have proved to be efficient tools for reconstructing the transport paths at a regional scale and
82 for identifying the main factors involved in deposition. Together with dune orientation, these
83 data give insight into past atmospheric circulation and have been widely used for
84 reconstructing wind directions during the Last Glacial in Europe (Poser, 1950; Maarleveld,
85 1960; Léger, 1990; Isarin et al., 1997; Zeeberg, 1998; Van Huissteden et al., 2001; Renssen et
86 al., 2007).

87

88 The aim of this paper is to combined geomorphological, sedimentological and geochemical
89 approaches to better understand the origin and dynamics of the aeolian deposition in
90 southwest France. Then, the reconstructed wind regimes during the Last Glacial in the study
91 area are compared with previous climatic simulations (Sima et al., 2009, 2013), in an attempt
92 to identify the period of the year favourable to dust activity for different climate patterns.

93

94 **2. Geological setting**

95

96 Since the beginning of the Middle Pleistocene, the centre of the Aquitaine basin was quite
97 similar to its current configuration and corresponded to a sand plain almost lacking relief and
98 gently sloping toward the Atlantic Ocean. Such a context, unparalleled along the French
99 coast, favoured the accumulation and preservation of aeolian deposits throughout the glacial
100 phases of the Pleistocene (Bertran et al., 2011; Sitzia et al., 2015).

101

102 At the basin scale, the following geomorphological units have been distinguished (Fig.1B) :
103 (1) ventifact pavements, present mainly in the northern part of the basin at the surface of the
104 plateau alluvium on both sides of the Garonne River (Plateau Girondin and Blayais) ; (2) the
105 coversands (“Sable des Landes Formation” *sensu* Sitzia et al., 2015), limited by the Garonne
106 river to the east and the north, the Adour river to the south, and the Atlantic Ocean to the
107 west; (3) a large loess belt mainly developed to the east and the south of the coversands.

108

109 Recent works allowed reappraisal of the regional chronostratigraphy of sand and loess
110 deposits (Bertran *et al.*, 2009, 2011; Sitzia et al., 2015). Three main phases of coversand
111 accumulation were identified throughout the Last Glacial (Sitzia et al. 2015): (1) Marine
112 Isotopic Stage (MIS) 4-3 (64-32 ka), characterized by predominant wet sandsheet
113 accumulation; (2) MIS 2 and the beginning of MIS 1 (Greenland Interstadial 1e) (25-14 ka),
114 associated with the development of dry sandsheets and coincident with the maximum
115 expansion of the Landes desert; (3) Younger Dryas (YD, Greenland Stadial 1), marked by the
116 development of parabolic dune fields. Parabolic and dome dunes dating back to the 14th-19th
117 centuries have also been documented and probably reflect local dune reactivation due to
118 increased human impact on the vegetal cover during the Little Ice Age (Bertran et al., 2011).

119

120 Sitzia et al. (2015), through the analysis of the chronological data of near-surface coversands,
121 revealed an uneven geographical distribution of the ages. They are dominantly Middle
122 Pleistocene in the northernmost part, MIS 4-3 in the Plateau Girondin, and MIS 2 to YD in the
123 Plateau Landais to the south. According to the authors, this pattern might have resulted from a
124 deficit in sand in the north during the Last Glacial Maximum (LGM) since the emerged
125 continental plateau was much wider (and as a consequence, the coastline was more distant)
126 than in the south. Available ages for loess suggest that they accumulated mainly during the
127 two last glacial periods (Bertran et al., 2011; Hernandez et al., 2012).

128

129 **3. Modern wind regime.**

130

131 Prevailing winds along the Atlantic coast of mid-latitude Western Europe are handled by two
132 major pressure areas, the Azores High, located in the Atlantic Ocean between latitude 25 and
133 40°N, and the Icelandic Low, located between latitude 60 and 75°N. Seasonally, the position
134 of these two pressure centres shifts and the wind regime changes. During the winter, dominant
135 winds come from the southeast, whereas westerly to northwesterly winds are the most

136 prominent during the summer (Fig. 2A). The map of mean wind speeds at 10 m available at
137 the European Joint Research Centre (<http://agri4cast.jrc.ec.europa.eu/DataPortal>) shows the
138 following (Fig.2B):

139

140 (1) The highest wind speeds are observed along the coast and in the Garonne estuary
141 where channelling of the airflows leads to wind acceleration;

142

143 (2) Overall, the wind speeds decrease the further the altitude increases;

144

145 (3) The area with the lowest mean wind speed is located at the southern edge of the
146 Pleistocene coversands.

147

148 **4. Data and methods**

149

150 *4.1. Analysis of dune patterns*

151

152 The pattern of the dune ridges has been analysed using (1) digitized black-and-white aerial
153 photographs from the 1950 missions (1/25,000) of the French Institut Géographique National
154 (IGN) and (2) IGN aerial colour photographs available on Google Earth. The main dune
155 morphologies were described according to McKee (1979). To establish the crest direction we
156 followed the approach proposed by Ewing et al. (2006). For barchanoid ridges, a line parallel
157 to the general direction of the dune crest was used.

158

159 Most of the parabolic dunes have been mapped during the 1/50,000 geological survey of the
160 Aquitaine basin by the French Bureau des Recherches Géologiques et Minières (BRGM)
161 (Capdeville and Dubreuilh, 1994). The dune morphologies were described according to the
162 terminology of Pye and Tsoar (2009), and the growth direction was measured using the
163 bisector of the angle between the two arms of the dune.

164

165 Where available, the Digital Elevation Model (DEM) 5 m of the IGN (RGE Alti® 2.0, ©IGN
166 2014) made it possible the observation of the detailed morphology of the parabolic dunes.
167 Since the small aeolian ridges remain difficult to identify from the DEM, a photogrammetric
168 survey has been performed with a drone in 3 areas of 1 to 8 km², which provided more precise
169 elevation data. The pixel resolution is 0.3 m.

170

171 *4.2. Grain size, thickness and geochemistry of the aeolian deposits*

172

173 The grain size and thickness dataset consists of (1) data collected in this work or previously
174 published (Bertran et al., 2009, 2011; Dubreuilh, 1976; Thibault, 1970), and (2) data available
175 from the Donesol database (INRA, 2011) for the Gironde, Landes, Dordogne and Lot-et-
176 Garonne districts. The datasets were cleaned to select only the outcrops and the stratigraphical
177 units of aeolian origin. The selected data were finally treated for the purpose of geostatistical
178 analysis.

179

180 The samples for grain-size analysis were processed in the PACEA laboratory (Université de
181 Bordeaux) using a Horiba LA-950 laser particle size analyser. The sample pre-treatment
182 includes suspension in sodium hexametaphosphate (5 g/L) and hydrogen peroxide (35%) for
183 12 hours, and 60 seconds of ultrasonification in the analyser to achieve optimal dispersion.
184 The Mie solution to Maxwell's equations provided the basis for calculating the particle size as
185 is recommended by the ISO committee (Jones, 2003; ISO, 2009), using a refractive index of
186 1,333 for water and 1.55i - 0.01i for the particles (see Supplementary Information 2). In the
187 Donesol database, the grain-size limits are those used by the French pedologists (Duchaufour,
188 1997), i.e. 2000–200 μm (coarse sand), 200–50 μm (fine sand); 50–20 μm (coarse silt); 20–2
189 μm (fine silt); <2 μm (clay). The analyses used in the database have been done by wet sieving
190 of the particles >50 μm and with the pipette method for the particles <50 μm . The results
191 obtained with this method cannot be directly compared with those obtained with the laser
192 particle size analyser (Konert and Vandenberghe, 1997; Buurman et al., 2001; Antoine et al.,
193 2001). Here, after comparing 20 samples treated using both methods, the following limit
194 classes have been chosen: <4.6 μm (clays), 4.6–26 μm (fine silt), 26–50 μm (coarse silt),
195 >50 μm (sand). These limits are similar to those used by Antoine et al. (2003). Polymodal
196 grain size distributions expressed in phi units have been deconvoluted in populations of
197 particles characterized by a Gaussian distribution using the software Fityk 0.9.8 (Wojdyr,
198 2010) to identify precisely the main modes and to follow their spatial evolution.

199

200 The samples selected for the geochemical analysis (n=77) come from outcrops studied during
201 this work or previously published (Bertran et al., 2009, 2011; Sitzia et al. 2015). The outcrops
202 have been selected in order to characterize the main types of aeolian deposits and their
203 potential sources: (1) Last Glacial coversands n=22) ; (2) YD parabolic dunes (n=13) ; (3)

204 loess from the Landes area (n=17) ; (4) loess from the Entre-Deux-Mers area (n=4) ; (5)
205 recent alluvium of the Leyre (n=4), the Ciron (n=3), the Garonne (n=3) and the Adour (n=3)
206 rivers; (6) Plio-Pleistocene plateau alluvial deposits (n=7). The samples from aeolian sands
207 have been taken below the B horizon of the surface podzol. All the loess samples come from
208 the argillic Bt horizon and were taken between 0.6 and 0.9 m deep. After removing the grains
209 >2 mm, the sand fraction was separated from the component <63 μm by wet sieving and
210 crushed. The sand and the fine fractions have been analyzed apart by Energy Dispersive X-
211 Ray Fluorescence (ED-XRF) by the spectrometry division of Ametek (Elancourt, France).
212 The measurements were made with a XEPOS instrument and do not allow quantification of
213 elements lighter than aluminum. Quantification was based on fundamental parameters.

214

215 Some representative samples of aeolian deposits and possible sources have been studied by X-
216 Ray Diffraction (XRD). As for ED-XRF, the sand and silt/clay fractions were studied
217 separately. The analysis was carried out by the Institut de Chimie de la Matière Condensée in
218 Bordeaux (ICMCB) using a PANalytical X'Pert diffractometer with a Bragg - Brentano theta
219 - theta configuration. The measurements were made with a measuring time of 30 mn,
220 between 8 and 80° and with a minimum step angle of 0.02°. Mineralogical identification was
221 performed using the EVA software coupled with a JCPDS-ICDD PDF2 database.

222

223 The grain-size and the geochemical data are expressed as percentages; therefore, they have to
224 be considered as compositional data (Aitchison, 1982). The analysis of compositional data by
225 classic multivariate statistics provides inappropriate results (Aitchison, 1982). Aitchison
226 (1986) proposed a robust solution to this problem, which used the logarithms of component
227 ratios instead of the raw percentages. In this work, because of the nature of the data, we used
228 non parametric replacement of the null values according to the method proposed by Martín-
229 Fernández et al. (2003) for below-detection values. Here, all the compositional data were
230 analysed using the CoDaPack software (Comas and Thió-Henestrosa, 2011).

231

232 For the purpose of geostatistical data analysis, a variographic analysis (Webster and Oliver,
233 2007) has been carried out with the VarioWin software (Pannatier, 1996). For compositional
234 data, in agreement with Pawlowsky and Burger (1992), each variable was transformed before
235 variographic analysis. Here, the Isometric Logratio Transformation has been applied
236 (Egozcue et al., 2003). Finally, we used ordinary kriging (Webster and Oliver, 2007) for the
237 interpolation through the geostatistical toolbox of ArcGis 9.3.

238

239 The models obtained have been tested by a double cross validation. In ArcGis, the value of a
240 variable is estimated in each point of the space by kriging and this operation is repeated every
241 time by removing one of the points for which the true value was originally known. Following
242 this procedure, the software calculates the mean error (ME) and the mean square deviation
243 ratio (MSDR). The mean error has to be close to 0, while the mean squared deviation ratio has
244 be as close as possible to 1 (Webster and Oliver, 2007). The model validity has been also
245 measured through an independent data set. The original database (grain size, thickness)
246 selected for the geostatistical analysis has been split before the variographic analysis using the
247 random selection spatial tool of ArcGis. We chose to randomly select 10% of the total number
248 of samples. The remaining 90% have then been used for the variographic analysis and the
249 calculation of the ME and MSDR statistics. Doing so, it was possible to plot the observed
250 values (independent data set) against the estimated values and to calculate a linear regression
251 between the two sets. Following Pawlowsky-Glahn and Olea (2004), the fitness of the
252 regression for grain-size data was evaluated by calculating the standardized residual sum of
253 squares (STRESS). The closer the value of the STRESS index to zero, the less is the
254 difference between observed and estimated values, and thus the best is the tested modeling.

255

256 *4.3. High-resolution glacial climate simulations*

257

258 We dispose of two numerical palaeoclimatic simulations at high-resolution over the studied
259 area that we can compare to our proxy data. The first simulation, designed for the Last Glacial
260 Maximum (LGM), was used by Banks et al. (2009) for niche modelling. In this simulation,
261 the orbital parameters and the atmospheric CO₂ concentration (Berger, 1978; Raynaud et al.,
262 1993) were set to their values for 21 ka. The LGM ice-sheet configuration (height and extent)
263 reconstructed by Peltier (2004, ICE-5G) was prescribed, and the land-sea mask was adjusted
264 to a sea-level drop of about 120 m. The GLAMAP2000 reconstruction of the LGM sea-
265 surface temperatures (SSTs) and sea-ice extent was imposed (Paul and Schäfer-Neth, 2003;
266 Sarnthein et al., 2003). The second simulation is one of the numerical experiments used by
267 Sima et al. (2009, 2013) to study the European climate (and dust) sensitivity to changes in
268 North-Atlantic SSTs as those associated with the MIS 3 millennial-timescale variability. The
269 orbital parameters and CO₂ concentration are set to 39 ka values, the ice-sheet configuration,
270 also selected from the ICE-5G reconstructions, corresponds to a sea level about 60 m lower
271 than today (i.e. half of the LGM value), and the land-sea mask is adapted to the sea-level

272 drop. Because no reconstruction or modelling result was available at that time for MIS 3 sea-
273 surface conditions, the GLAMAP2000 dataset was imposed in the base-line MIS 3
274 simulation. A cold and a warm perturbation were obtained by applying zonal anomalies of up
275 to 2°C to the GLAMAP North-Atlantic SSTs in the latitudinal band between 30°N and 63°N.
276 Considering the differences of North-Atlantic SSTs between the LGM and an average MIS 3
277 state suggested by experiments with an earth system model of intermediate complexity (Van
278 Meerbeeck et al., 2009), here we use the warm perturbation (hereafter “MIS 3” simulation) to
279 characterize the average MIS 3 conditions in the comparison with the LGM.

280

281 In both MIS 3 and LGM experiments, the modern vegetation cover was prescribed in most
282 regions, following the PMIP recommendations (e.g. Braconnot, 2004). Thus, in the area of
283 interest here, trees occupy 60-80% of every grid-point in the coversands (in yellow on Fig.
284 1B), and about 15- 40% elsewhere. The soil is bare all year long on 5-15% of the grid-points’
285 surface, and the rest up to 100% of each grid-point is occupied by grass. This is a drawback
286 because the LGM vegetation was much scarcer. The use of modern vegetation does not affect
287 the simulated wind direction, but only the wind speed, which is strongly reduced due to
288 overestimating the roughness length. As the wind speed is a key element in determining the
289 season(s) when dust emissions occur, we also present sensitivity tests of this variable to more
290 appropriate values of roughness length. Using prescribed present-day vegetation also leads to
291 unrealistically overestimating its effect of protecting the ground from aeolian erosion. Thus,
292 for the impact of the surface conditions on the dusty season, we will only be able to discuss
293 soil moisture and snow cover.

294

295 The simulations were performed with the LMDZ atmospheric general circulation model
296 (version 3.3), coupled with the SECHIBA land surface model (Krinner et al., 2005), on a
297 stretched grid with resolution down to 60 km on Western Europe. For each simulation there
298 are 21 years of run, one year of spin-up followed by 20 years for analyses.

299

300 **5. Results**

301

302 *5.1. Dunes*

303

304 5.1.1. Low amplitude dunes

305

306 Among the low amplitude dunes, four main morphological types have been identified from
307 aerial photographs: transverse ridges, barchanoid ridges, isolated barchanoid ridges and dome
308 dunes/sand patches. The ridges appear in cultivated areas as bright lineaments separated by
309 darker interdune areas due to variations in soil organic and water content. Most of the ridge
310 fields were observed in the Plateau Girondin area, where the groundwater table is close to the
311 ground surface and the YD parabolic dunes are scarce. According to available chronological
312 data, these ridges date back to MIS 4-3, or possibly to older phases of sand deposition in the
313 northernmost area. The DEM provided by photogrammetry shows that the ridge pattern
314 corresponds to regular to irregular low amplitude reliefs, 0.3 to 0.7 m high and 70 to 150 m in
315 wavelength (Fig. 3). The reliefs are aligned along the bright lineaments visible in aerial
316 photographs. In contrast, no significant ridge areas were found in aerial photographs to the
317 south on the Plateau Landais where the MIS 2 coversands are thicker. Field survey, however,
318 showed that ridges also were present there in small areas forming well-preserved linear reliefs
319 approximately 1 m in height (Fig. 4). In cross-sections, all these dune-like forms are
320 associated to a sandsheet facies; high-angle stratification is almost absent (Sitzia et al., 2015).

321

322 In the Plateau Girondin, the following observations can be made (Fig. 5): (1) the different
323 types of ridge morphologies form homogeneous fields elongated parallel to the coast; (2) the
324 dome dunes/sand patches are located on the margins of the barchanoid and transverse ridge
325 fields; (3) the dune fields are replaced toward the east by an area characterized by thin
326 sandsheet deposits almost lacking dune-like forms; (4) the distribution of the different types
327 of morphologies seems to be mainly controlled by topography.

328

329 A total of 425 measurements of crest direction have been made from barchanoid and
330 transverse ridges. The rose diagrams show that predominant crest directions are consistently
331 N-S (i.e. the vector perpendicular to dune crests is W-E, Fig. 5A). Local departure from this
332 direction was found to be mainly controlled by topography. This is particularly obvious in
333 Figure 5C for the transverse ridges whose crest direction, roughly parallel to the contour lines,
334 vary between N-S and NE-SW within a few kilometres.

335

336 5.1.2. Parabolic dunes

337

338 The parabolic dunes are 10 to 15 m high and exceptionally exceed 25 m (Dune de Cazalis,
339 Legigan, 1979). The total number of digitized dunes is 2636. Following the typology

340 proposed by Pye and Tsoar (2009), simple (hairpin, lunate), complex (hemicyclic, digitated,
341 nested, rake-like, en-echelon) and compound (long-walled ridge with secondary transverse
342 arms) forms were observed.

343

344 The parabolic dunes exhibit a non-random distribution over the landscape. They are mainly
345 clustered around the rivers, and form corridors up to 80 km in length along shallow valleys
346 oriented W-E and over the plateau (Fig. 6). The dune morphology changes according to
347 topography. Along the Leyre and the Ciron valleys, as well as near the coast, the parabolic
348 dunes are mostly simple; on the contrary, complex and compound forms are mainly observed
349 in the corridors (Fig. 7A, C). Complex forms several kilometres in length have been observed
350 in the main dune corridor (Fig. 7B).

351

352 A total of 394 measurements of dune orientation have been made. The rose diagrams reveal a
353 systematic orientation towards the ENE and a low dispersion of the values (Fig. 6).

354

355 *5.2. Grain-size gradients at the basin scale*

356

357 In the Donesol grain-size database, 513 out of the 3683 sites, considered to be pure (i.e. non
358 colluviated) aeolian deposits, were retained (Fig. 8). Among the selected sites, 462 were used
359 in the geostatistical analysis and 51 for cross-validation. The results of cross-validation (see
360 Supplementary Information) indicate that both the variographic analysis and the ordinary
361 kriging were satisfactory. The interpolated maps of the content in coarse sand, fine sand,
362 coarse silt and fine silt are shown in Figure 8. The following observations can be made: (1)
363 every grain-size class shows a gradient oriented NW-SE; (2) the coarse fractions are mainly
364 located in the north-western part of the basin and extend inland from Arcachon to the SE; (3)
365 the Garonne and Adour valleys (and their tributaries) serve as limits for the coversands; the
366 silty sand facies are restricted to a narrow strip on the left bank of the Garonne river, whereas
367 they form a broader band on the right bank and in the southern part of the basin; (4) in the
368 loess belt, the ratio coarse/fine silt varies only slightly in the study area. When increasing the
369 number of grain-size classes, it can be observed that, both in the Entre-Deux-Mers area and in
370 the southern part of the basin, the coarse silt maximum is located northwest of the fine silt
371 maximum (Fig. 9). Comparison between these maps and that derived from the LUCAS
372 database (Fig. 1) shows that loess have a more limited extent in the latter, where the mapped
373 aeolian units have, therefore, to be considered as minimal.

374

375 The sand deposits extend up to the border of the Landes plateau and show a main mode
376 ranging between 340 μm for the northernmost site (Retjons) and 270 μm for the southernmost
377 one (Pujo-le-Plan) (Fig. 10B). In the Adour valley and its tributaries, the distribution is
378 polymodal. From the north to the south, the main mode is first in medium sands (270 μm) and
379 then decreases rapidly and gives way to a mode on silts (22 μm). In this area it can be noticed
380 (1) a rapid decrease in the size of the sand particles (Fig. 10B); (2) a decrease in the
381 contribution of the coarse modes in favour of the finer modes (Fig. 10C). Finally, south to the
382 Adour valley, a mode around 17 μm becomes predominant (Fig. 11A). This mode decreases
383 from 17 to 14 μm quite evenly as a function of the distance to the coversands. For all the sites
384 located between the Adour valley and the Pyrenean foothills, significant amounts of clay are
385 present. The mode, close to 3 μm , remains steady regardless of the distance to the coversands
386 (Fig. 11A), whereas the total clay content slowly increases with distance (Fig. 10D).

387

388 In the northern part of the basin, the grain-size distributions show that (1) the sands of the
389 Blayais area are relatively coarse (mode between 300 and 600 μm); (2) the silts of the Entre-
390 Deux-Mers area have a mode between 30 and 34 μm ; this is appreciably coarser than the
391 mode of the loess deposits in the southern part of the basin; (3) the silt mode decreases
392 significantly toward the SE. The most distant sample from the sources (Barbas) has a mode
393 around 12 μm .

394

395 *5.3 Thickness variability at the basin scale*

396

397 According to Karnay et al. (2010), the coversands (Sable des Landes Formation) form a thin
398 sheet, usually less than 2-3 m thick. Nevertheless, the underlying Castet Formation (Dubreuilh
399 et al., 1995), which is now considered to be part of the Sable des Landes Formation (Sitzia et
400 al., 2015) and reaches locally up to 40 m in thickness, was not considered by Karnay et al.
401 (2010). As a whole, the coversand thickness decreases from the west to the east according to
402 the distance to the coast and with altitude.

403

404 The loess data were partitioned in two sets, one for the north (n= 292), and the other for the
405 south (n=434). The geostatistical analysis points out that (1) the ordinary kriging decreases
406 systematically the standard deviation of the thickness around the mean; (2) the deviation
407 between measured and estimated values is larger for the northern data set. The Anderson-

408 Darling test reveals that the deviation distribution fits a normal (Gaussian) distribution. For
409 the southern loess, the accuracy of thickness estimation is 57.4 cm and the mean error is 81.8
410 cm. The most significant errors are due to extreme values. For the northern loess, the accuracy
411 of estimation and the mean error cannot be evaluated because the deviation distribution does
412 not follow a Gaussian function.

413

414 The interpolated maps show that the loess thickness decreases rapidly with increasing
415 distance to the coversands, in a gradual way in the southern area, more steeply in the northern
416 area (Fig. 12). The maximum loess thickness reaches 3.5 m in the south and 5 m in the north.
417 Because of the uneven distribution of the selected Donesol data and the loose geographical
418 coverage, the interpolated maps do not make it possible to identify if deposition preferentially
419 occurred on particular slopes according to their orientation.

420

421 *5.4. Geochemistry and potential sources*

422

423 The distribution maps of elemental compositions (Fig. 13) and the scattergrams of the
424 elements versus Si (Fig. 14) (mainly as quartz) point out the following:

425

426 (1) Zr, the principal component of zircon, and Ti, probably as oxides, are mainly concentrated
427 in the fine fraction (silt, fine sand). Therefore, these elements are mostly abundant in loess
428 and in the fine fraction of the coversands. In a Si-Ti diagram, the samples of aeolian deposits
429 are aligned along a line with a negative slope (Fig. 14, line a), except for the fine fraction of
430 some coversand samples. This strongly suggests that the most of the aeolian samples result
431 from the size fractionation of a common source of intermediate composition between that of
432 sand and that of fines (Fralick and Kronberg, 1997). The samples from the Adour and
433 Garonne rivers also show a linear trend (line b), but plot on a different line from that of the
434 aeolian deposits, which indicates different sources. On the contrary, the samples from the
435 Leyre and the Ciron rivers, which drain both the coversands and old alluvial deposits, plot
436 among the aeolian samples or between the Adour-Garonne and loess samples. Zr has a
437 behaviour roughly similar to Ti. The loess composition appears to be homogeneous when
438 compared with the whole data set. Such an observation is also valid for all the elements.

439

440 (2) Ca, as a component of carbonate or silicate minerals (feldspar, amphibole), is lacking in
441 the aeolian sand fraction, whereas it has been found in small amount in the alluvial deposits.
442 Ca has been detected as traces in the fine fraction of a few coversand samples where it is
443 probably associated with Al in silicate minerals.

444

445 (3) K is present both in the aeolian sand fraction as part of silicate minerals (feldspar,
446 muscovite) and in slightly larger amounts in the fine fraction (illite, mixed-layer illitic clay).
447 Mean concentrations are significantly greater in the alluvial deposits. In the Si-K diagram, the
448 K content of the aeolian sand fraction appears to be uncorrelated to Si and the samples are
449 aligned on a horizontal line. On average, the sand fraction of loess is depleted in K in
450 comparison with the coversands. The samples of fine fraction do not plot on the same line as
451 the sands, but below it, suggesting a loss of K due to alteration and soil leaching (Fralick and
452 Kronberg, 1997).

453

454 (4) On the Si-Al diagram (Fig. 14), the alignment of the aeolian samples along the same line
455 suggests a partition of Al between sand and fines due to sedimentary sorting, with Al mainly
456 concentrated in clays. High Al contents plotting above this line have been found only in the
457 fine fraction of the coversands. Such enrichment in Al can be related to organic-mineral
458 complexes in the B horizon of the surface podzol (Righi and De Coninck, 1974; Melkerud et
459 al., 2000).

460

461 The principal component analysis (“compositional biplot”, Aitchison and Greenacre, 2002)
462 for the sand fraction (Fig. 15) explains 72.3% of the total variance. The first axis
463 discriminates the group Ca+K+Rb+Sr from the group Fe+Ti+Zr. K and Rb are mainly related
464 to K-silicates, mostly feldspar and micas (De Vos et al., 2006; Wedepohl, 1978). Sr has a size
465 intermediate between K and Ca and can replace these two elements both in silicates and in
466 carbonate minerals (De Vos *et al.*, 2006). Because of the lack of carbonates in the studied
467 samples, the group K+Rb+Sr reflects the pole of potassium silicates, while Ca corresponds to
468 calcium silicates. Fe (+ Cr as a substitute) is present in silicate minerals and in oxides and
469 hydroxides (De Vos *et al.*, 2006). The group Fe, Ti and Zr thus represents the pole of oxides
470 and the minerals resistant to alteration.

471

472 The position of the samples in the biplot allows making the following remarks: (1) the
473 samples of aeolian sands (Landes, Blayais, parabolic_dunes), the old alluvial terraces of the

474 Garonne river (Plateau_alluvium), the recent alluvial deposits (Fz) of the Midou and Leyre
475 rivers (which essentially drain the Sable des Landes Formation) are distributed between the
476 barycentre of the biplot and the pole quartz + K-silicates; (2) the sand fraction of the southern
477 loess (Loess_Landes) is associated with the pole of resistant minerals and oxides; (3) the
478 recent alluvial deposits (Fz) of the Adour and Garonne rivers, which drain the Pyrenees, plot
479 apart from the other samples, and are opposed to the pole quartz + resistant minerals; they
480 include a high proportion of easily alterable Ca-minerals; (4) the sand fraction of the loess
481 from the Entre-Deux-Mers area is located between the southern loess and the recent alluvial
482 deposits.

483

484 The biplot for the fine fraction (Fig. 16), which takes into account a larger set of elements
485 because of values above the detection limit, produces results substantially different from those
486 for the sand fraction. It explains 63.7% of the total variance. Several poles can be
487 distinguished. The first correspond to calcium-rich minerals (Ca); the second to resistant
488 minerals, quartz and zircon (Si+Zr+Hf); the third to potassium silicates (K+Rb, and Ba,
489 strongly adsorbed by soil clays, Reimann et al., 2014); and finally, the poles of Fe-oxides
490 (Fe+Cr+As) and of Al-complexes (Al).

491

492 In this biplot, the loess samples appear more homogeneous and plot in the proximity of the
493 pole oxides and K-silicates (probably dominated by clays). The loess composition partially
494 covers that of coversands and is located apart of the recent alluvial deposits, which are close
495 to the pole of Ca-minerals.

496

497 *5.5. Results of climate simulations*

498

499 Figure 17 shows the results of the LGM and MIS 3 simulations for wind direction and for the
500 variables that constrain the period of the year when dust emissions can occur, i.e. daily
501 maximum wind speed (more informative for dust mobilization than daily average), snow-free
502 surface fraction, and dry soil depth. The annual cycle of these variables is constructed on daily
503 data for winds, and on monthly data for surface conditions, averaged over 20 years of run for
504 each climate state and over the domain between longitude 2°W-1°E and latitude 43°-45.5°N
505 (delimited by the black rectangle in Fig. 1A and zoomed in Fig. 1B).

506

507 The annual cycle of wind direction (Fig. 17A, B) is similar in the two simulations and not
508 much different from the present-day one (Fig. 2). In both cases, the winds blow from W-SW
509 in winter, turn from W in springtime, then from W-NW in summer. A reverse evolution
510 occurs from the summer to the winter, through westerly winds in autumn. The daily
511 maximum wind speed, averaged over the domain and over 20 years of run (black curves in
512 Fig. 17C, D), is highest in winter, lowest in summer, and springtime values are higher than
513 autumn ones. The LGM daily maximum winds are generally stronger than for MIS 3,
514 especially from mid-May to October. The differences are small, and the values themselves are
515 relatively low, less than 7 m.s^{-1} . When looking at individual grid cells and individual days of
516 the 7200-point time series (i.e., 20 years x 360 days/year of run on a 360-day calendar), the
517 highest (winter) daily maximum winds in the MIS 3 run often exceed 12 m.s^{-1} , with peaks at
518 about 16 m.s^{-1} , while in the LGM simulation the winter values are frequently above 16 m.s^{-1} ,
519 with extremes above 20 m.s^{-1} . The simulated winds are thus too weak considering the strong
520 dust activity suggested by the field data. This is due to the high values of the roughness length
521 (z_0) calculated by the model on the basis of the present-day vegetation cover that was
522 prescribed in both simulations. For comparison, z_0 is about 37 cm on average on the studied
523 domain, while values of 1 cm are associated in the model to bare soil. The vegetation was
524 considerably scarcer than today for both LGM and average MIS 3 climates, so we recalculate
525 and plot the daily maximum winds for two values of z_0 more adapted to the investigated area
526 and periods. The blue curves in Fig. 17C, D correspond to $z_0 = 1.5 \text{ cm}$, and the green ones to
527 $z_0 = 4 \text{ cm}$. The first z_0 value is the one that the model would calculate for a surface covered
528 by about 50% of grass, and the rest of bare soil. The second one corresponds to a majority of
529 grass, with less than 10% bare soil and less than 10% trees (that can be assimilated to shrubs,
530 which are not represented in the model). With these lower z_0 , the averages of daily maximum
531 wind increase considerably, they almost double for $z_0 = 1.5 \text{ cm}$. Looking at individual days
532 and grid cells, even for the relatively less windy MIS 3 and the lower test value $z_0 = 4 \text{ cm}$, the
533 maximum daily values regularly exceed 25 m.s^{-1} in winter, with a few peaks around 40 m.s^{-1} .
534

535 Park and In (2003) have statistically determined a wind-speed threshold of 7.5 m.s^{-1} for dust
536 mobilization in sandy areas from Asian deserts. If we consider this limit to also apply to the
537 Landes coversands, the LGM average maximum winds practically never drop below it in the
538 test with $z_0=1.5 \text{ cm}$, so that only the surface conditions would determine the dusty part of the
539 year. For the MIS 3 run, the average daily maximum winds drop below the threshold in July
540 for $z_0 = 1.5 \text{ cm}$, at mid-May for $z_0 = 4 \text{ cm}$ respectively, and strengthen again in November in

541 both tests. The tests with these z_0 values can be interpreted in two ways. When comparing
542 LGM to MIS 3, using the spatial averages of variables over the entire domain, the blue curve
543 for the lower z_0 (for about 50% grass and 50% bare soil) is more appropriate to characterize
544 the LGM, and the green curve for the higher z_0 (majority of grass, with some shrubs and bare
545 soil) is more appropriate for a MIS 3 situation. For each of the LGM and MIS 3 simulations
546 separately, the blue curve can be seen as representing the parts of the domain with scarce
547 vegetation (mainly the coversands) and the green one, the more vegetated ones – mainly the
548 loess areas, where part of the dust was trapped.

549

550 As a general evolution, the snow-free fraction of the surface and the depth of dry topsoil are
551 low in the cold season and high in the warm one (Fig. 17E, F), the important question being to
552 determine the interval when both variables are high enough to allow significant dust
553 emissions.

554

555 In both simulations, most of the Landes surface is snow-free all year long. From October to
556 February the depth of dry topsoil averaged over the domain is too small to allow a significant
557 dust activity (a 5 mm threshold value was used in Sima et al. 2009). The dusty season could
558 thus start in the beginning of March. At least for the LGM this is questionable, due to a warm
559 winter bias causing an underestimation of both snow cover and soil moisture. Indeed, as
560 mentioned in Banks et al. (2009), the simulated LGM climate over Western Europe is in
561 agreement with pollen-based reconstructions (Wu et al., 2007) for summer and annual mean
562 temperatures, and for mean annual precipitation, but winter cooling is underestimated by a
563 few degrees. Thus, a comparison with the reconstructed directions for dust-efficient winds is
564 needed to better constrain the beginning of the dusty season. Dust activity is strong until May
565 or June, decreases to a minimum in July-August (it can even stop in more vegetated parts of
566 the domain), and may briefly intensify again in September-October before ceasing from
567 November to February due to unfavourable surface conditions (here soil moisture, in reality
568 probably also snow).

569

570 In the MIS 3 simulation, the average dry soil depth becomes high enough to allow significant
571 dust emissions in the beginning of March. Dust activity stops in May due to wind weakening,
572 which actually may be accentuated by the seasonal development of vegetation, not captured in
573 the simulation. According to the model, areas with scarce vegetation might continue to emit

574 dust in the beginning of the summer. The simulated winds start to intensify again in autumn,
575 but the increasing soil moisture plays against a restart of the emissions.

576

577 **6. Discussion**

578

579 *6.1. Dunes*

580

581 Low amplitude ridges

582

583 The ridges observed in the Gironde district, regardless of their morphology (transverse,
584 barchanoid), are interpreted as zibars based on their low amplitude and the lack of slipface
585 (Holm, 1960; Warren, 1972; Tsoar, 1978; Kocurek and Nielson, 1986; Nielson and Kocurek,
586 1986; Mountney and Russell, 2004). Zibars are often composed of coarse sand and the grain
587 size is thought to be the main factor involved in preventing the development of a slipface
588 (Bagnold, 1941; Cooke and Warren, 1973; Nielson and Kocurek, 1986).

589

590 The interpretation of the transverse and barchanoid ridges in the study area as zibars seems to
591 be the most likely, even if they do not meet all the criteria mentioned above. Indeed, the
592 zibars of the study area do not show a mode in coarse sands but in medium sands, similar to
593 that found in parabolic dunes. With regards to dune morphology, we did not find in the
594 literature zibar examples with a similar range of pattern (transverse, barchanoid, isolated
595 ridges). Transverse and barchanoid ridges reflect unidirectional winds and point to sand
596 transport perpendicular to the wind (Pye and Tsoar, 2009), in an environment almost devoid
597 of vegetation.

598

599 The interpretation of the variations in dune pattern across the coversands is not
600 straightforward, since the chronological data indicate that the ridge fields developed at
601 different times: the barchanoid and transverse ridges in the Barp-Cestas area formed during
602 MIS 4-3, whereas further north the barchanoid ridges date back to the Middle Pleistocene
603 (Sitzia et al., 2015). Therefore, it remains unclear whether the morphological differences
604 reflect changes in aeolian dynamics (especially in the availability of sand) during the same
605 depositional phase or whether they testify to different periods of aeolian deposition.
606 Nonetheless, it appears that topography, and particularly the slope gradient of the surface on
607 which the ridges progressed, largely controlled their morphology.

608

609 Parabolic dunes

610

611 Parabolic dunes form in partially vegetated landscapes mainly under unidirectional winds, and
612 are observed in all latitudes (Lancaster, 1995; Pye and Tsoar, 2009), along sandy coasts (Pye,
613 1982), but also inland (Sun and Muhs, 2007). They develop mainly from blowouts (Pye and
614 Tsoar, 2009; Hesp, 2002; Barchyn and Hugenholtz, 2013). The morphology of parabolic
615 dunes is controlled by the wind strength and direction, the sand supply and the nature of
616 vegetated ground (Pye, 1982; Pye and Tsoar, 2009; Wolfe et al., 2008). The parabolic dunes
617 gradually evolve into elongated forms. In some cases, the arms give rise to linear dunes when
618 the sand supply is low.

619

620 According to available OSL and ^{14}C ages (Sitzia et al., 2015), the formation of the parabolic
621 dunes in the Landes area took place mainly during the YD. The clusters of dunes along the
622 valleys may suggest that the sand was partly supplied by the alluvium. However, the presence
623 of dune corridors expanding over the plateau and of dunes located on the windward side of
624 the valleys oriented N-S (e.g. Leyre, Ciron and Avance), testifies unambiguously that the
625 local fluvial deposits were not the main sand sources. A coastal origin for the sand transported
626 along the corridors parallel to the wind direction appears to be the most likely.

627

628 The development of dunes along the rivers resulted from the deceleration of the winds
629 crossing the valleys. Wiggs et al. (2002) proposed the following conceptual model: (1) on the
630 plateau area above the valley, erosion prevails and sediment accumulation is patchy; (2) on
631 the windward side of the valley, the wind loses its carrying capacity and sediment tends to
632 accumulate; (3) on the opposite side of the valley, the wind accelerates and the erosive
633 potential increases; (4) on the plateau, the wind decelerates again over a short distance
634 promoting sand deposition. This model correctly explains the dune distribution in the study
635 area.

636

637 *6.2. Loess sources and transport dynamics*

638

639 The interpolated maps of sand and silt contents suggest a genetic relationship between the
640 aeolian facies, as previously proposed by Bertran *et al.* (2011, 2016). The distribution of the
641 contour lines shows a mean NW-SE gradient at the basin scale.

642

643 In the Landes and the Blayais areas, a gradual transition between coarse and fine sand
644 deposits can be observed, which is not controlled by the nature of the substrate but reflects
645 wind sorting and/or decreasing carrying capacity as suggested by the distribution of current
646 mean wind speeds (Fig. 2). The sand-silt transition is mainly controlled by the topographic
647 change that occurs between the plateau areas (Plateau Girondin, Plateau Landais, Blayais) and
648 the main valleys (Garonne, Adour and Dordogne). Several authors have already outlined the
649 role played by topography in the coversand-loess transition, among others Kasse (1997),
650 Appelman (1956) and Mason et al. (1999). In the southern part of the basin, the transitional
651 facies (silty sands and sandy silts) form a strip on the right bank of the Adour river, showing
652 that deposition of the suspended particles was primarily triggered by the expansion of airflows
653 above the valley. In the northern part of the basin, the transitional facies form a narrower belt
654 on the left (windward) side of the Garonne valley (Fig. 8), in a context broadly comparable to
655 the Adour area. In the estuary area to the north, the maps show an almost lack of fine-grained
656 accumulation on both sides of the valley. The sand-loess transition shifts by approximately 40
657 km to the east. The abundant ventifacts found on the alluvial terraces point to a deflation-
658 dominated area.

659

660 The geochemical analysis also strongly suggests a genetic relationship between coversand and
661 loess. The sand fraction of the southern loess is depleted in K-silicates in comparison with the
662 coversands and correlatively enriched in quartz and resistant minerals. The enrichment is
663 especially obvious for Zr and Ti, which concentrate mainly in the fine fraction and tend
664 broadly to increase as a function of the distance to the coversands. Such a preferential
665 accumulation of Zr and Ti in loess deposits has been observed worldwide (Gallet et al., 1998;
666 Scheib et al., 2014) and is interpreted as resulting mainly from grain-size sorting. The recent
667 alluvial deposits of the Adour river, which are rich in Ca-minerals, did not contribute
668 significantly to dust production, which derive mainly from the coversands. In the Entre-Deux-
669 Mers area, the alluvium of the Garonne river provided higher amounts of dust, as suggested
670 by the larger content of easily alterable minerals in the sand fraction of the loess and, to a
671 lesser extent, in the fine fraction. Overall, loess appears to be much more homogeneous than
672 coversand, whose composition is closer to the alluvial sources. We assume that this
673 homogeneity results from the mixing in the atmosphere of the different dust sources (Gallet et
674 al., 1998).

675

676 The maps of the thickness of aeolian deposits, which cumulate several sedimentary phases,
677 provide complementary information. In the northern part of the basin, the thickest loess
678 deposits are located close to the Garonne valley in the Entre-Deux-Mers area and show rapid
679 thinning toward the east (Fig. 12A). In the southern area (Fig. 12B), the loess deposits are
680 mainly located on the left bank of the Adour river and decrease in thickness more gradually
681 toward both the south and the east. Distinct areas of thick loess accumulation are visible on
682 the maps but can be, at least in part, an artefact due to kriging using a limited number of data.
683 The observed pattern of loess thickness can be explained by the following factors:

684

685 (1) The distance to the dust sources. Close to the sources, the decrease in thickness of the
686 deposits is initially very rapid because the coarser particles, which occupy a large volume,
687 accumulate at short distances. Farther, the decrease is more gradual and due to lower
688 concentration of suspended particles in the dust clouds (Simonson and Hutton, 1954;
689 Waggoner and Bingham, 1961; Frazee et al., 1970). The observed pattern points clearly to the
690 Sable des Landes Formation and the Blayais sands as the main sources of dust. The rivers
691 (Dordogne, Lot) flowing from the Massif Central mountains did not provide significant
692 amounts of dust, as suggested by the lack of variation in loess thickness around the valleys. A
693 plausible explanation is that armouring developed quickly in the dominantly coarse-grained
694 alluvial deposits during the Glacial, preventing significant dust entrainment by deflation.

695

696 (2) A change of surface roughness related to the vegetation cover (Tsoar and Pye, 1987). It is
697 likely that the transition between coversand and loess was associated with a change in plant
698 cover, which favourable for trapping the particles.

699

700 *6.3. Pattern of dust production and accumulation*

701

702 The geochemical analysis shows that northern and southern loess differ substantially in origin.
703 For the latter, dust derived mainly from aeolian abrasion of the coversands (cf. Wright et al.,
704 1998; Crouvi et al., 2010). Disconnection of the Landes plateau from the main drainage axes
705 since at least the early Middle Pleistocene (Sitzia et al. 2015) due to tectonics resulted in
706 intense redistribution of older alluvial sands by deflation and in the development of
707 widespread coversands during the successive glacial periods. Repeated remobilization of the
708 same aeolian sand stock together with fresh inputs from the continental shelf led to
709 progressive enrichment of the sand material in minerals resistant to alteration (Legigan,

710 1979). The particle transport was predominantly NW-SE and the contribution of the alluvial
711 plains, mainly oriented perpendicular to the wind, was low. The sand-loess transition is
712 controlled by topography and coincides with the Adour valley (or the valleys of the
713 tributaries).

714

715 In the northern part of the basin, dust came from aeolian abrasion of the coversands in the
716 Blayais and the Plateau Girondin areas, and by deflation of alluvial material from the Garonne
717 river. Channelling of the winds by the wide Garonne corridor and higher wind speeds
718 favoured large dust production from the plain. Dust accumulated in the Entre-Deux-Mers area
719 to form thick deposits mostly restricted to a narrow band on the valley side.

720

721 The loess thickness, which reaches at best 3.5 m in the south and 5 m in the Entre-Deux-Mers
722 area, is low in comparison with values observed in other European regions (Lautridou, 1985;
723 Haase et al., 2007). Several factors have been proposed to explain the low rate of
724 accumulation (Bertran *et al.*, 2011): (1) a limited production of dust by aeolian abrasion of the
725 coversands; (2) a narrow and steep continental shelf that does not allow storage of large
726 amounts of fine alluvial sediment in the Garonne valley during sea lowstands; (3) a high soil
727 moisture and, therefore, a low ground susceptibility to wind erosion due to an oceanic climate.
728 The palaeoclimatic simulations both for LGM and MIS 3 agree to show that the average
729 rainfall was significantly higher near the Atlantic coast than in Eastern Europe (Laîné et al.,
730 2009; Kjellstrom et al., 2010; Strandberg *et al.*, 2011; Sima et al., 2013) where much thicker
731 loess deposits accumulated. However, this role remains questionable insofar as the
732 reconstructed maximum in rainfall occurred in the winter season, which did not necessarily
733 correspond to the most favourable period of aeolian transport.

734

735 *6.4. Last-glacial wind regime from data and model constraints*

736

737 The grain-size gradient together with the orientation of low-amplitude ridges and parabolic
738 dunes allow reconstructing the prevailing wind directions of aeolian transport during the Last
739 Glacial and the Lateglacial (Fig. 17A). The transverse ridges of the Plateau Girondin area
740 dated to MIS 4-3 indicate efficient winds from the W. Some variability exists but appears to
741 result from local topographical factors and from channelling of the air masses along the
742 Garonne corridor. Unfortunately, the almost lack of MIS 2 ridge fields visible in aerial
743 photographs does not allow getting data for this period. The crests of the MIS 2 ridges

744 illustrated in Figure 4 are oriented W-E, and indicate local efficient winds blowing from the
745 north. At the basin scale the contour lines of the grain size point to winds originating from the
746 W or the NW, and are assumed to reflect the mean direction of efficient winds during the
747 glacial. Finally, the parabolic dunes attest for winds coming from the W-SW during the YD.
748 The following considerations must be taken into account with regard to the reconstruction of
749 wind directions:

750

751 (1) The accuracy of the geological proxies used may vary. The dunes are considered as the
752 best marker for the reconstruction of past wind circulation patterns (Lancaster, 1990),
753 especially when dunes formed under unidirectional winds (Marrs and Kolm, 1982). With
754 regard to grain-size gradients, Renssen et al. (2007) suggested that a significant uncertainty is
755 associated with this type of feature. The analysis made here across the Aquitaine basin,
756 however, shows a good consistency in the regional distribution of the contour lines.

757

758 (2) The reconstructed directions of aeolian transport are indicative of the wind regimes at the
759 regional scale during the periods of accumulation. In most of the modern sand seas, the axis
760 of unidirectional dunes is oriented perpendicular (e.g. barchan, transverse ridges, parabolic
761 dunes) to the dominant wind direction, i.e. the direction with the highest transport rate of sand
762 (Fryberger and Dean, 1979). Seasonal fluctuations in wind strength are assumed to be the
763 main factor involved in modulating the intensity of deflation. In periglacial environments,
764 however, wind intensity is not the only factor that controls the season of aeolian transport.
765 Other factors, particularly the snow cover and soil moisture, play a significant role on the
766 susceptibility of the ground to deflation (Koster, 1988; Seppälä, 2004). Depending on the
767 temperature and precipitation regimes (also determining the vegetation), different situations
768 exist. In cold and arid environments, based on observations in the Søndre Strømfjord region
769 (West Greenland), Dijkmans and Törnqvist (1991) have shown that aeolian transport occurred
770 mainly during part of the winter season, whereas summer was less favourable to deflation
771 because of vegetation growth and ground-ice melting. Based on laboratory experiments,
772 McKenna-Neuman (1989) also showed that the release of sand material susceptible to be
773 transported by the wind can very active when the temperature drops below -20°C because of
774 ice sublimation. On the contrary, between -5 and -15°C, the erosion threshold increases
775 considerably due to the presence of a significant amount of liquid water in the ground.
776 Dijkmans and Koster (1990), for Alaska, underline that when the snow cover is thick, the
777 season of aeolian transport shifts towards the end of winter. According to Seppälä (2004), in

778 arctic and subarctic regions where snow is abundant, deflation is effective only during
779 summer, i.e. after snowmelt.

780

781 In the following, we compare our proxy data and palaeoclimatic simulations in an attempt to
782 specify the periods of the year when dust activity occurred during the investigated climate
783 intervals. The result depends on how reliable the simulated variables, especially the wind
784 directions, are. In the region studied here, the atmospheric circulation is largely determined by
785 the surface conditions (temperatures, sea-ice extent) in the North Atlantic; in glacial times,
786 another key factor was the size of the Laurentide ice sheet (e.g., Löffverström et al., 2014,
787 Beghin et al., 2015). In our simulation for 21 ka, these elements are imposed from
788 reconstructions for the LGM (GLAMAP, ICE-5G). Even though these reconstructions have
789 their uncertainties, we tend to be confident in the general aspect of the modelled annual cycle
790 of wind directions. For the 39 ka simulation, the sea-level drop is realistic and the ice-sheet
791 configuration is adapted to this sea-level drop, the main uncertainty being related to the
792 North-Atlantic surface conditions. Results of numerical experiments with the LOVECLIM
793 earth system model of intermediate complexity (Van Meerbeek et al., 2009) show a
794 difference of air temperature of 3°C on average over the North Atlantic between an average
795 MIS 3 stadial state and the LGM. For the MIS 3 run examined here, the difference is 2.5°C,
796 close enough to the Van Meerbeek et al. value, knowing that there was no unique (or
797 average) MIS 3 stadial climate, but many stadial episodes differing in ice-sheet size and
798 orbital parameters.

799

800 Since we are reasonably confident in the simulated wind directions, constraints on the dusty
801 season can be derived using the comparison between the modelled wind directions and those
802 reconstructed from field proxies, together with the results of sensitivity experiments with
803 lower roughness length more typical of glacial vegetation. The only other constraining
804 element that we can use is the simulated soil moisture, keeping in mind that it is
805 underestimated in the cold part of the year.

806

807 In the LGM simulation, soil moisture conditions allow dust activity to start in the beginning
808 of March, while reconstructed W-NW wind directions suggest that the dusty season should
809 not begin before mid-March, when simulated winds change from W-SW to W (Fig. 17A).
810 Sand transport took place during the spring/summer times because of scarce vegetation,
811 adequate wind speeds and well-drained sandy soils. According to recent investigations

812 (Andrieux et al., 2016), permafrost, which might have impeded drainage (Kasse, 1997), was
813 lacking during most of the period of sand accumulation. Deflation may have been more
814 restricted in summer in the loess area because of the denser vegetation cover. This scenario
815 contrasts with that proposed by Renssen et al. (2007) for more northern latitudes, which
816 suggested winter to be the most suitable period for deflation during MIS 2. In such a
817 permafrost context and by comparison with the modern periglacial environments investigated
818 by Dijkmans (1990) and Dijkmans and Törnqvist (1991), deflation was thought to have been
819 impeded during the summer because the ground was mostly wet (thaw of the active layer),
820 vegetation was growing, and the wind speeds were lower than in the winter season. It is worth
821 noting that the observations in modern milieus were made on aeolian systems supplied by
822 proglacial rivers. These rivers fed by glaciers are mostly active after the spring melt. In
823 summer, the water table rises and, as a consequence, the sediment availability to deflation is
824 low (Mountney and Russell, 2009). Such an aeolian system was probably characteristic of the
825 margins of the Scandinavian ice sheet (Koster, 1988), but cannot be transposed to regions far
826 from an ice sheet such as the Aquitaine basin. In this oceanic region, the ground was wet and
827 largely protected by a snow cover (largely underestimated in our simulation) during the
828 passage of winter storms, but dried in summer, thus becoming more susceptible for deflation.

829

830 In the MIS 3 simulation, the soil becomes dry enough in the beginning of March, while the
831 reconstructed westerly winds suggest that the dusty season should not begin before the end of
832 March. Both data and model agree that the dust activity should cease in the end of May, when
833 the simulated winds turn from W (which is the reconstructed direction) to WNW, and the
834 maximum wind speeds, recalculated with a roughness length value more typical of steppe
835 vegetation (green curve in Fig. 17D), become too weak. The modelled wind directions
836 coincide again with the reconstructed (westerly) ones in autumn, but the simulated increase in
837 soil moisture suggests that little, if any, dust activity should occur.

838

839 Wind strength, soil moisture and snow cover determine the dusty season(s) in scarcely
840 vegetated environments, Where vegetation has strong seasonal variations, its impact is also
841 important. The developing vegetation contributes to wind weakening through increasing
842 roughness length, and protects the surface against erosion, The later effect cannot be
843 estimated here due to the unrealistically high vegetation cover in the study region, but Sima et
844 al. (2009) have found that it stopped dust activity in the warm season under MIS 3 conditions
845 in Europe at latitudes around 50°N.

846

847 We do not have a simulation for the YD, but we note that the annual cycle of wind direction
848 does not considerably change between present day, LGM and MIS 3 (in our simulations, also
849 in those by van Huissteden and Pollard 2003), and even from the present day. Therefore, we
850 may speculate that winter, typified by dominantly W-SW wind directions, was the main
851 period of sand transport during the YD. In Western Europe, winter is the season that
852 corresponds to the passage of the main perturbations and associated storms, and that records
853 the highest wind speeds (Fig. 17C, D). These wind directions are in agreement with those
854 reconstructed for the YD by Isarin et al. (1997). The YD orientation of parabolic dunes in
855 Aquitaine is identical to that observed in the Netherlands (Isarin et al., 1997) and should be
856 associated with increased speed of SW winds throughout Northwestern and Western Europe
857 (Isarin et al., 1997). Differences should have been important in the surface conditions, which
858 for the YD allowed dust activity in winter: scarce vegetation, reduced snow cover compared
859 to LGM and MIS 3, and reduced soil moisture favoured by improved ground drainage due to
860 the lack of permafrost and to the weaker influence of meltwater from the ice sheets (for
861 Northern Europe).

862

863

864 **7. Conclusions**

865

866 The multi-approach study of the aeolian deposits of southwest France allows better
867 understanding of sediment sources and depositional dynamics. The maps of interpolated grain
868 size and the geochemical composition converge to show a genetic relationship between
869 coversands and finer-grained deposits. In the southern part of the basin, dust derived mainly
870 from aeolian abrasion of the coversands and is dominantly composed of minerals resistant to
871 alteration. In contrast, both the coversands and the floodplain of the Garonne river, richer in
872 Ca-bearing minerals, contributed to loess in the northern part of the basin.

873

874 The low-amplitude dunes lacking slipfaces, dated to MIS 4-3, are interpreted as zibars and
875 show variable morphologies (transverse, barchanoid, isolated barchanoid) occurring as strips
876 subparallel to the coastline. This pattern seems to be mainly controlled by the slope gradient.
877 The parabolic dunes dated to the YD are simple, complex or compound. These are clustered
878 along the rivers but are rare on the plateaus, which correspond to deflation areas. In the
879 valleys oriented W-E, i.e. parallel to the prevailing winds, corridors of elongated parabolic

880 dunes developed, whereas more simple shapes occur on both sides of the valleys
881 perpendicular to the winds. These formed as a consequence of relief-induced deceleration of
882 the airflows. Overall, the dune distribution indicates that the sand supply came mainly from
883 the coast.

884

885 The spatial analysis of dune shapes and grain-size gradients, as well as the study of dust
886 sources, made it possible reconstructing the Last Glacial wind regimes in the study region.
887 The efficient winds changed significantly through time. They came from the W to the NW
888 during the glacial, and the W-SW during the YD. The comparison between proxy data and
889 both the current and simulated Last Glacial wind regimes strongly suggests that aeolian
890 transport of sand and dust occurred mostly during the spring/summer season for the glacial,
891 and the winter for the YD. We assume that these variations were not related to changes in the
892 pattern of atmospheric circulation, but rather to changes in the deflation season. This was
893 mainly controlled by the length of the season with a snow cover and frozen or moist ground,
894 and by the seasonal evolution of the wind speeds able to cause deflation. Vegetation could
895 impose additional constraints in areas or periods when its seasonal variations were significant.

896

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898

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906

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1320

1321 **Figures**

1322 Fig. 1: European aeolian deposits. A – General map from Bertran et al. (2016) and location of
1323 the study area; B – Close-up view of southwest France. The distribution of ventifacts is from
1324 Bertran et al. (2011).

1325

1326 Fig. 2: Modern meteorological parameters in the studied area. A - Roses of seasonal winds in
1327 Aquitaine (period 1999-2007) according to MétéoFrance (Ile d'Yeu weather station); B -
1328 Mean wind speeds ($\text{m}\cdot\text{s}^{-1}$) at a height of 80 m, from Agri4cast (2016); the dashed line
1329 indicates the coversand limit.

1330

1331 Fig. 3: A, B - Detailed topography of barchanoid ridges (photogrammetry); location shown in
1332 Figure 4A. pr – Pleistocene ridge field; hs: Holocene sand cover at the front of a parabolic
1333 dune. North to the top. A', B' – Aerial views of the same areas (photos IGN / Google Earth).

1334

1335 Fig. 4: View of transverse ridges dated to MIS 2, Pontonx-sur-l'Adour.

1336

1337 Fig. 5: Ridges of the Plateau Girondin. A – Distribution of low-amplitude ridges (zibars) and
1338 orientation of the vector perpendicular to dune crests. B – Topographic transect between the
1339 Atlantic coast and the Garonne river, and associated dunes morphologies. C – Photo-mosaic
1340 of fields showing low-amplitude ridges (photos IGN / Google Earth) with superimposed 5 m
1341 contour lines, and close-up view of ridge fields.

1342

1343 Fig. 6: Distribution of parabolic dunes (in red) according to the 1:50,000 geologic map
1344 (Bureau des Recherches Géologiques et Minières) and orientation of the bisector of the angle
1345 between the two arms for simple and complex dunes. The rectangles indicate the location of
1346 figures 6A-D.

1347

1348 Fig. 7: Parabolic dunes, 5 m DEM (RGE Alti® 2.0, IGN 2014). The areas represented are
1349 shown in Figure 5. A – Dunes (indicated by white arrows) on both left and right banks of a
1350 river; B – Compound elongated parabolic dune; C - Parabolic dunes on the right (i.e. wind-
1351 facing) bank of a river; D - Dunes on the left bank of a river.

1352

1353 Fig. 8: Grain-size maps of the aeolian deposits interpolated by ordinary kriging. The white
1354 dots correspond to the sites used for kriging. The dotted line indicates the coversand limit.

1355

1356 Fig. 9: Detailed maps of the coarse (upper panel) and fine (bottom panel) silt content in the
1357 northern (A) and the southern (B) part of the Aquitaine basin.

1358

1359 Fig. 10: Evolution of the grain-size mode of some representative samples from the southern
1360 part of the basin. The modes were established after deconvolution of the grain-size curves
1361 using the software Fityk 0.9.8 (Wojdyr, 2010). The location of the samples is shown in
1362 Figure 8.

1363

1364 Fig. 11: Grain size distribution of some representative samples of sandy silts and loess from
1365 the Landes district. The location of the samples is shown in Figure 8.

1366

1367 Fig. 12: Map of the thickness of the loess located on the right bank of the Garonne river (A)
1368 and on the left bank of the Adour river (B). The coversand limit is indicated by a dashed line.

1369

1370 Fig. 13: Maps of the content in some elements of the aeolian and alluvial deposits. Fines = silt
1371 + clay.

1372

1373 Fig. 14: Ti, Zr, K and Al – Si scattergrams of aeolian and alluvial samples of the Aquitaine
1374 basin. UCC: upper continental crust composition according to Taylor and McLennan (1995).
1375 a – adjustment line for loess and the alluvial deposits of the Leyre and Ciron rivers; b –
1376 adjustment line for other alluvial deposits.

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1378 Fig. 15: Principal component analysis adapted to compositional data (“compositional biplot”)
1379 of major and minor elements (fraction > 63 μm) of aeolian and potential source deposits.

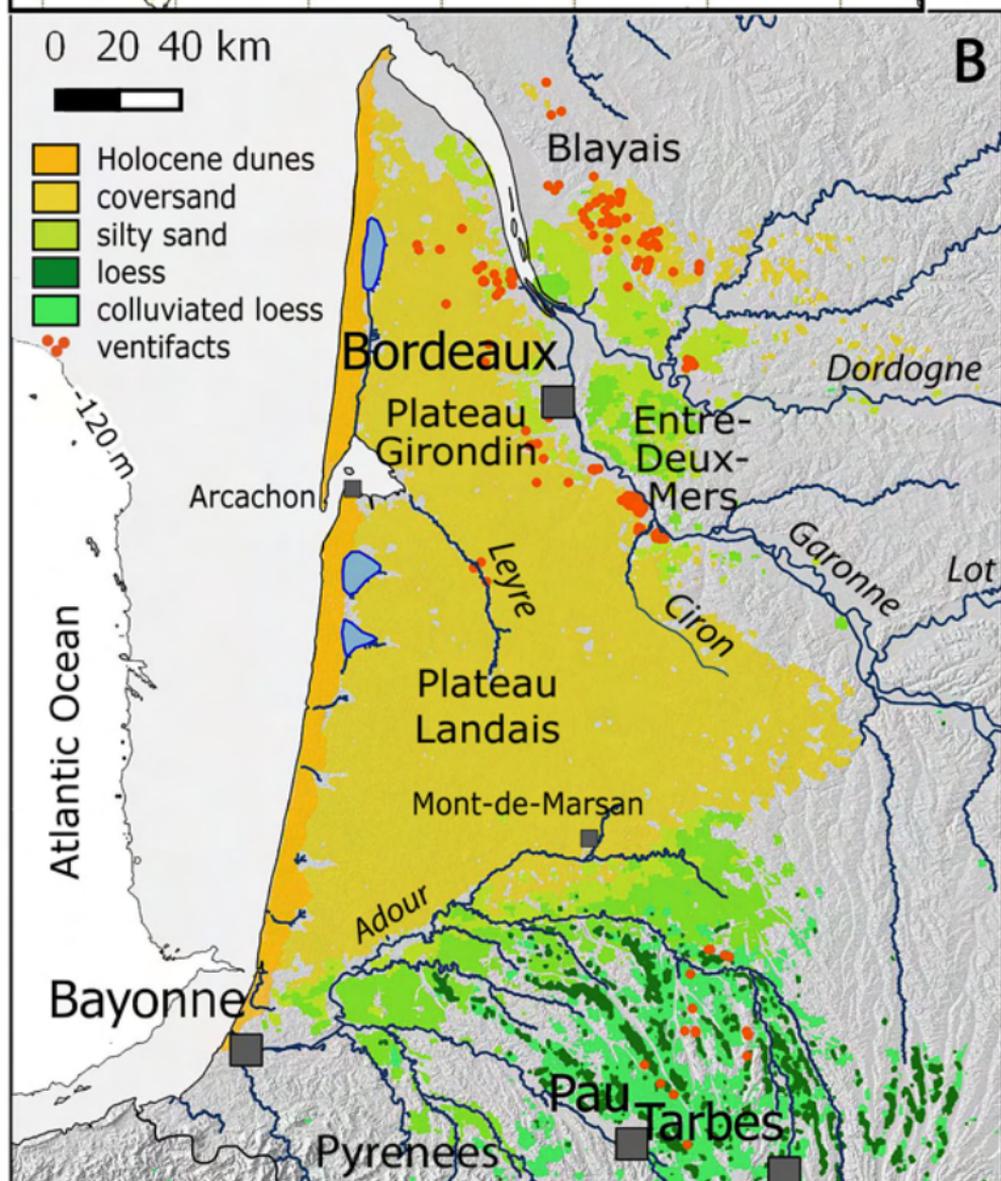
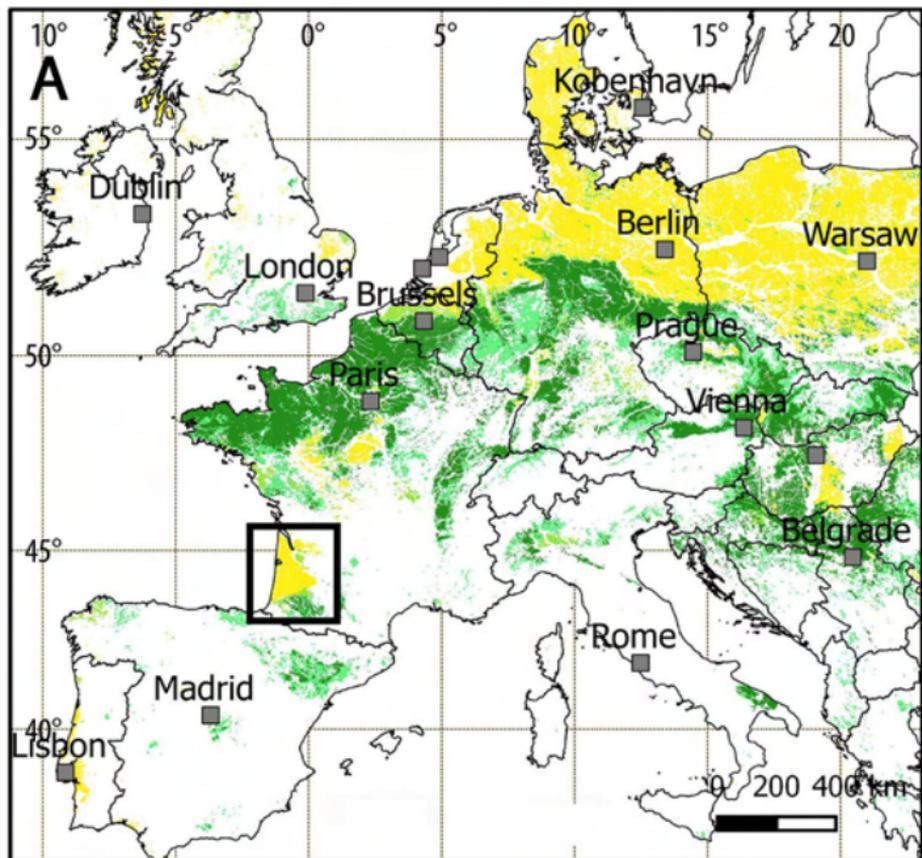
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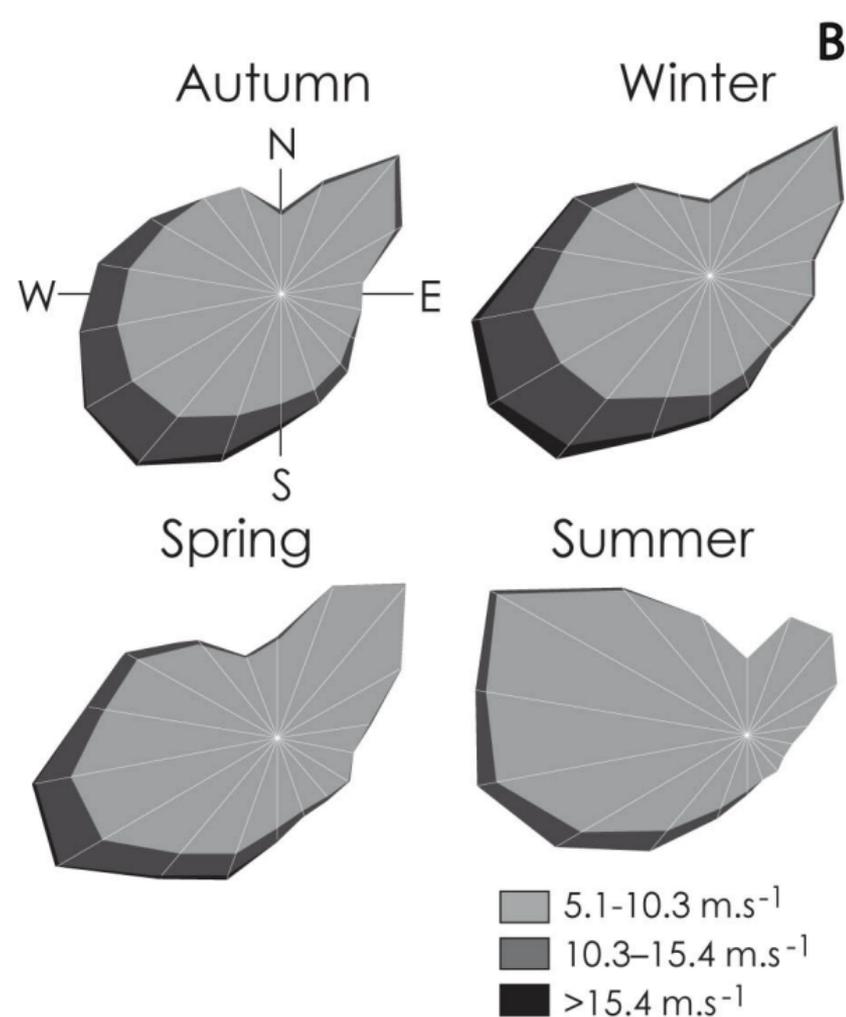
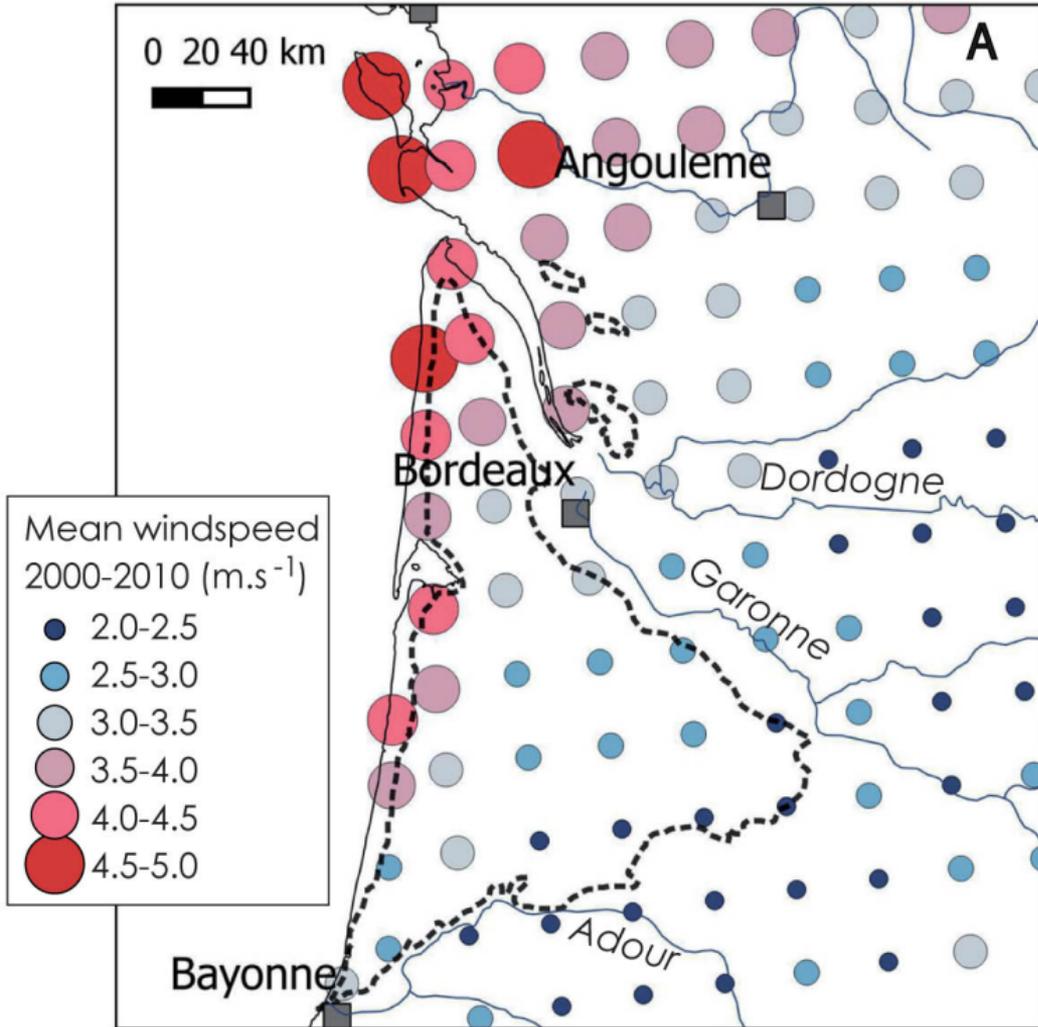
1381 Fig. 16: Compositional biplot of major and minor elements (fraction < 63 μm) of aeolian and
1382 potential sources deposits.

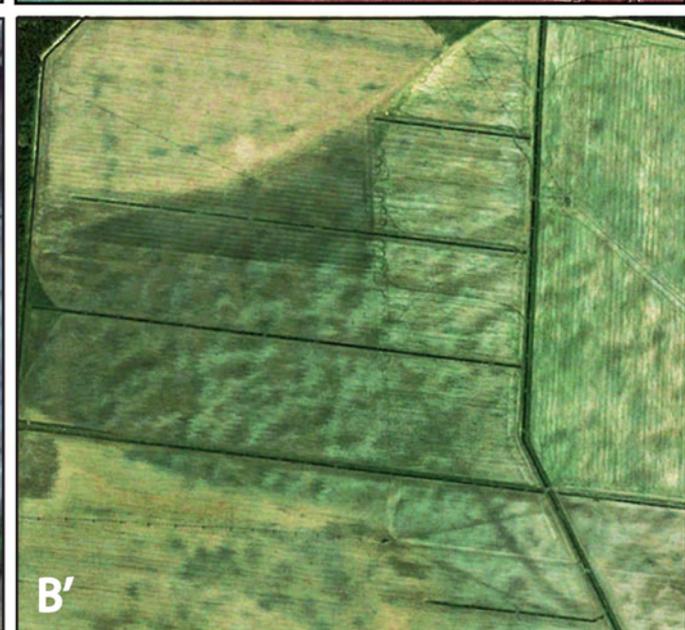
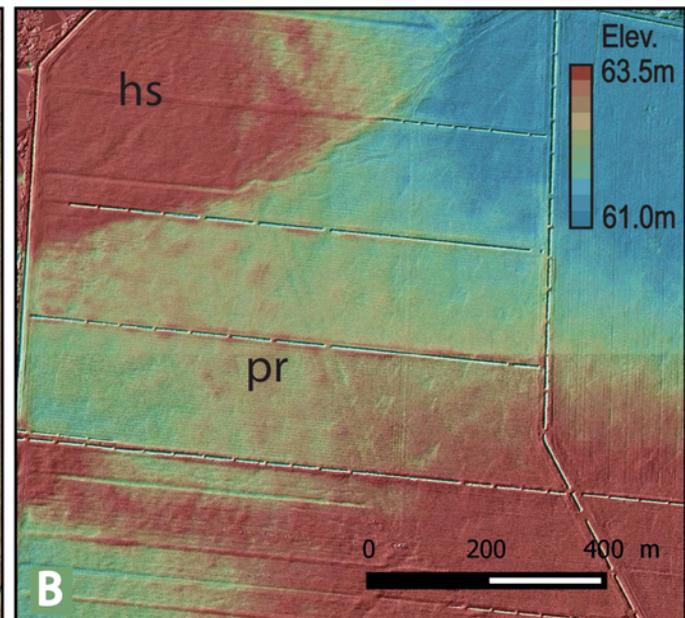
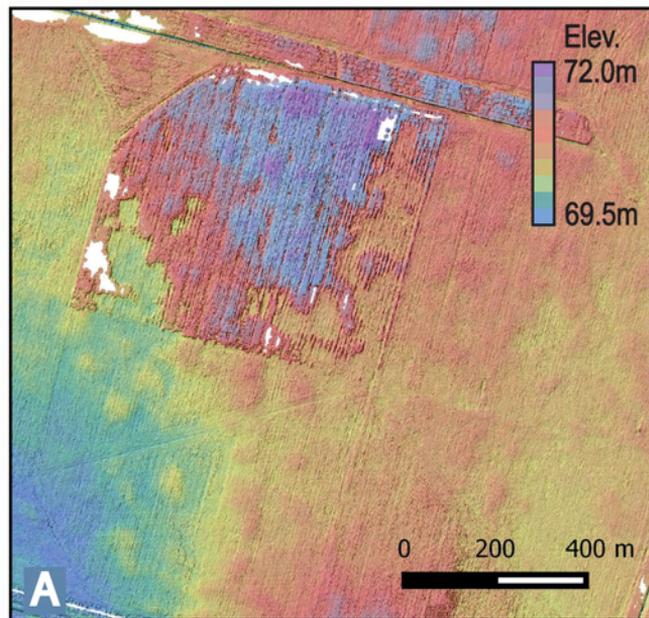
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1384 Fig.17: Annual cycle of surface winds and ground conditions for the LGM and MIS 3
1385 simulations (averaged over the domain between longitude 2°W-1°E and latitude 43°-45.5°N,

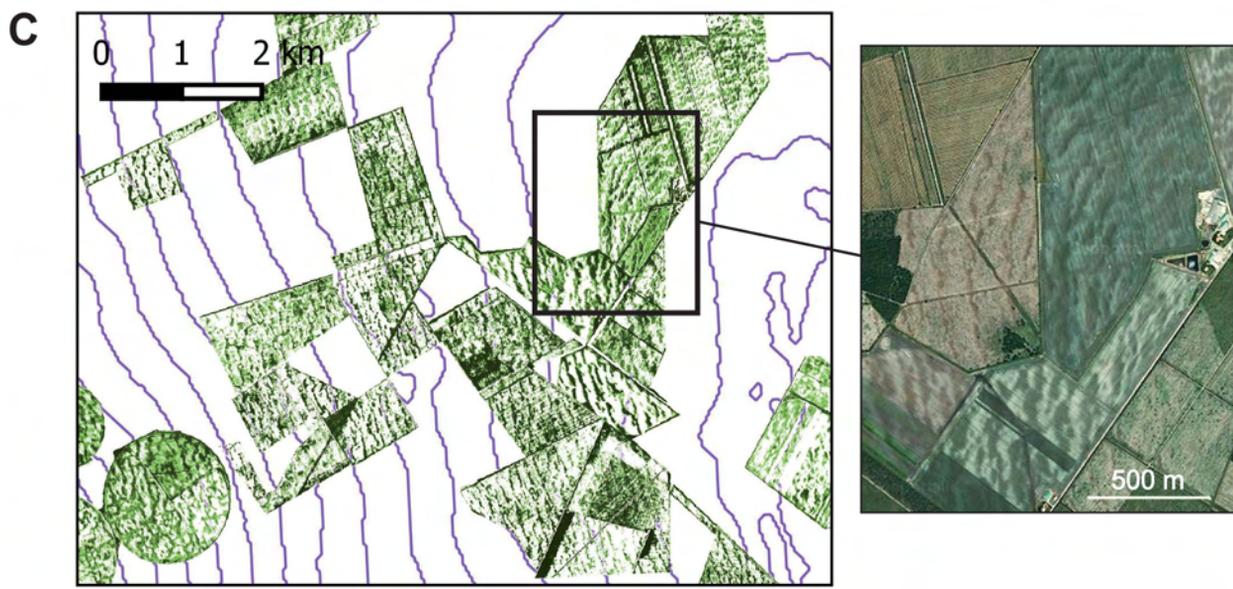
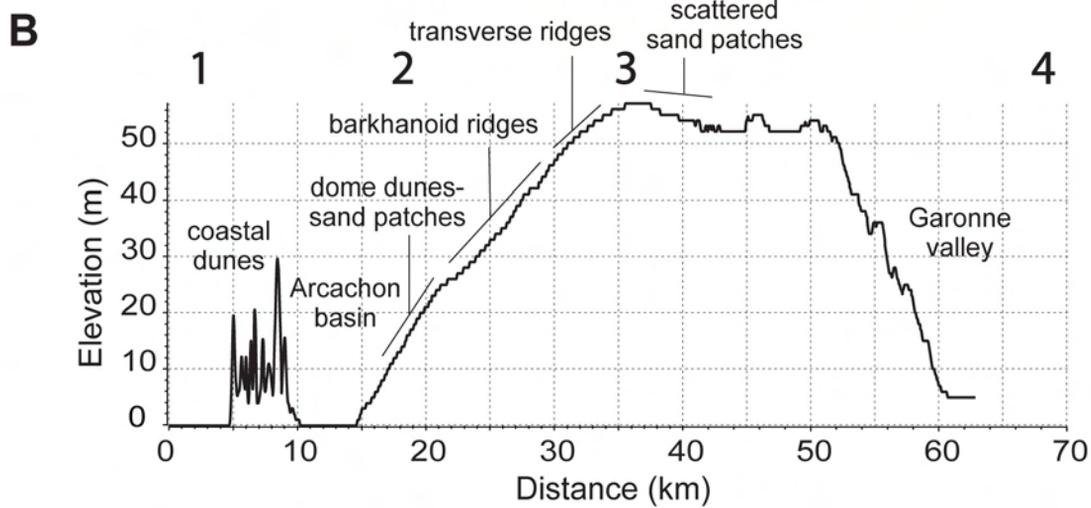
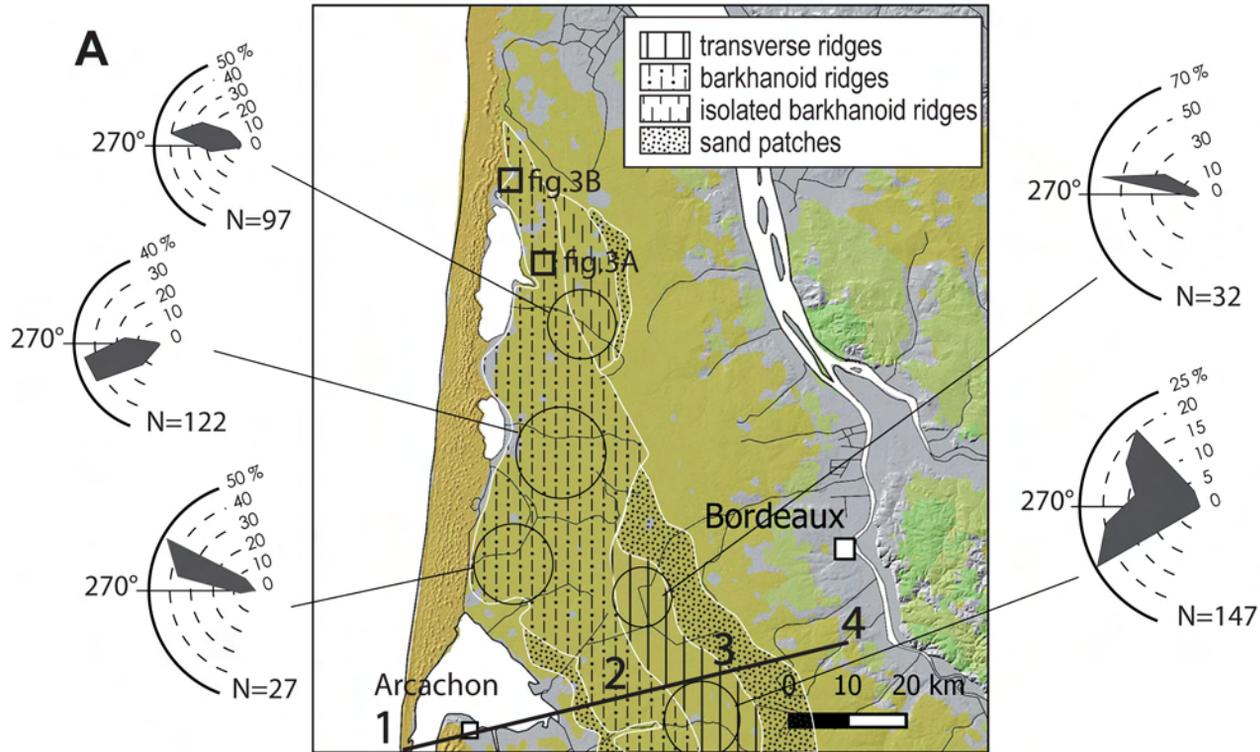
1386 and over 20 years for each run). From top to bottom: wind direction, daily maximum wind
1387 speed for different values of roughness length (z_0), and ground conditions (dry soil depth and
1388 percent of snow-free surface). Red dashed lines mark the different thresholds discussed in the
1389 text. Red rectangles indicate the most probable period of deflation according to the available
1390 constraints.

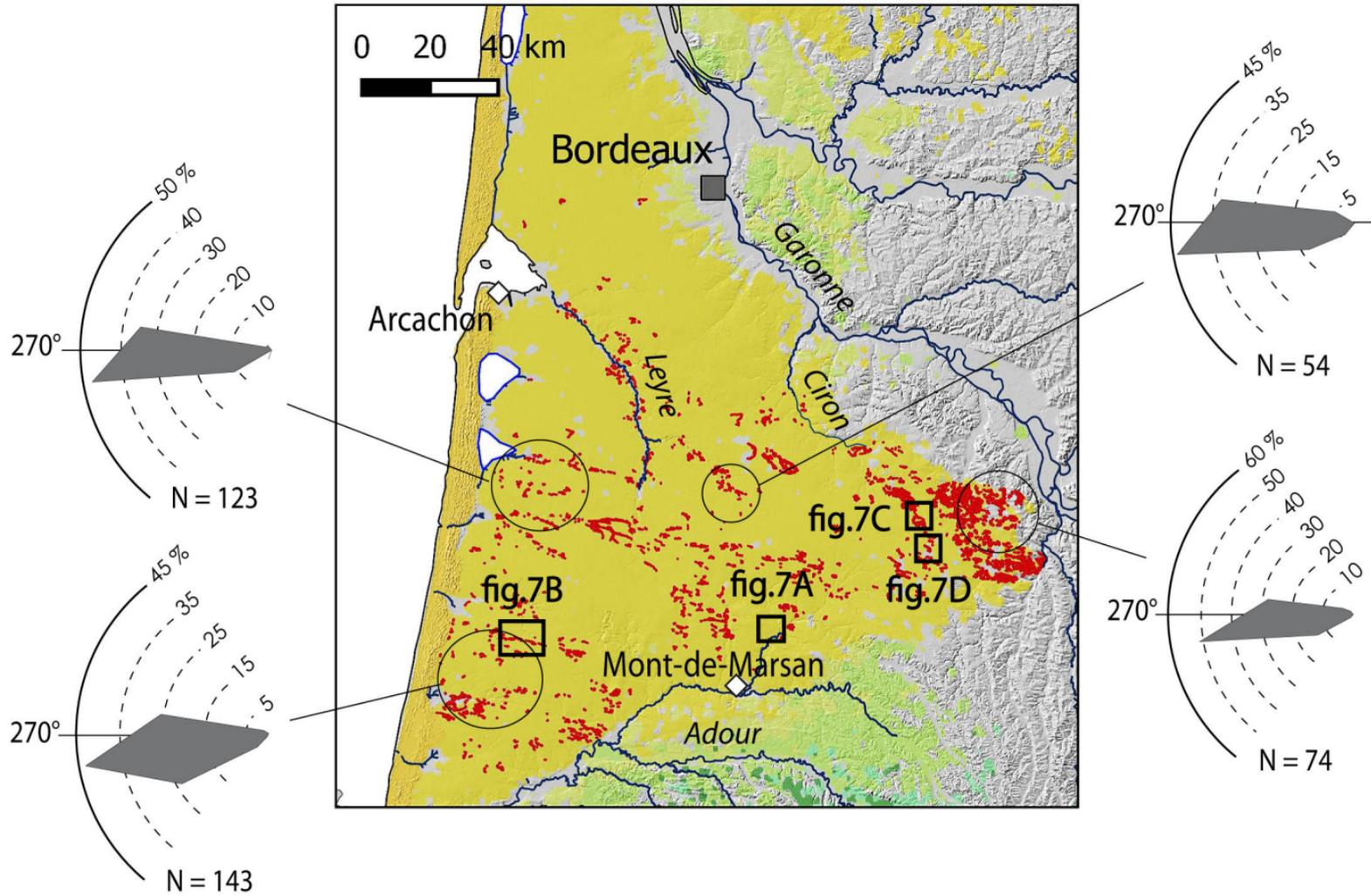


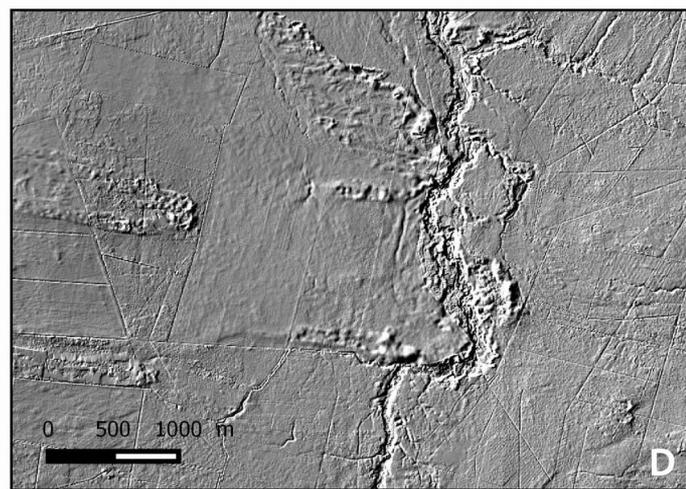
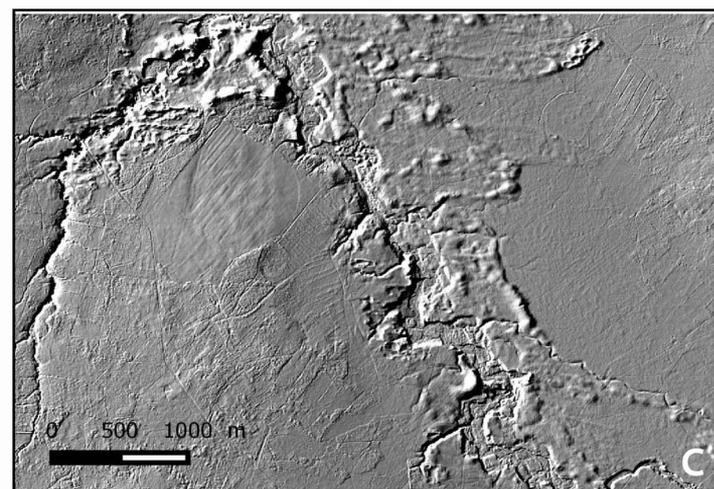
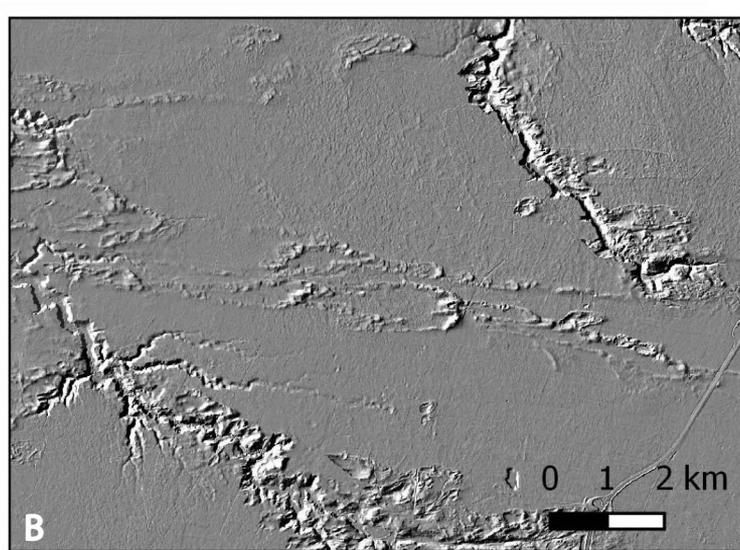
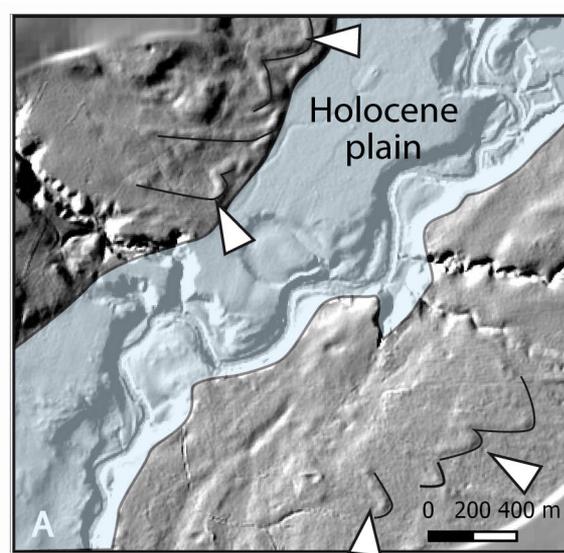




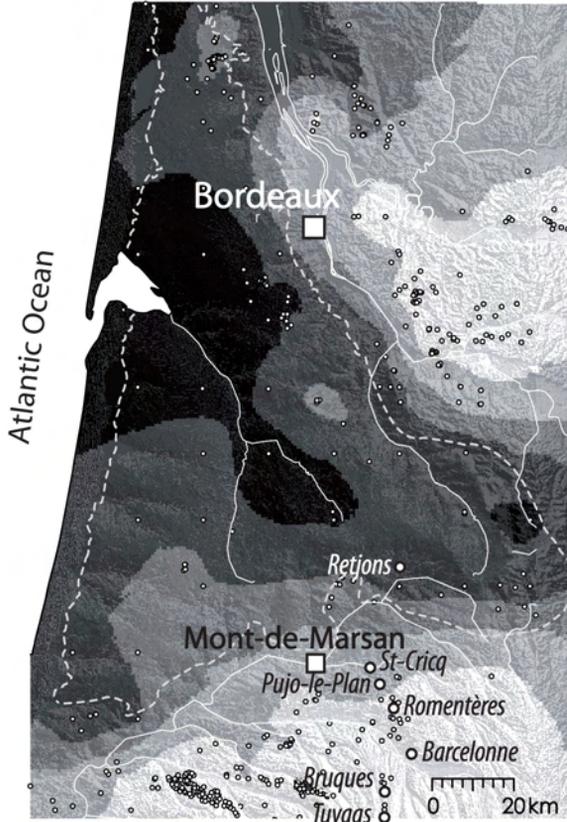
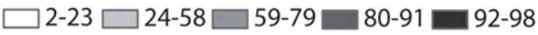




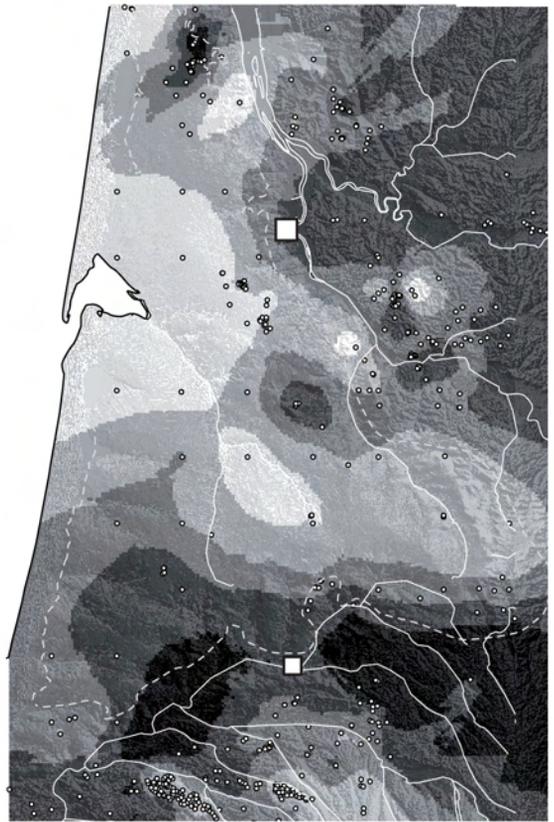




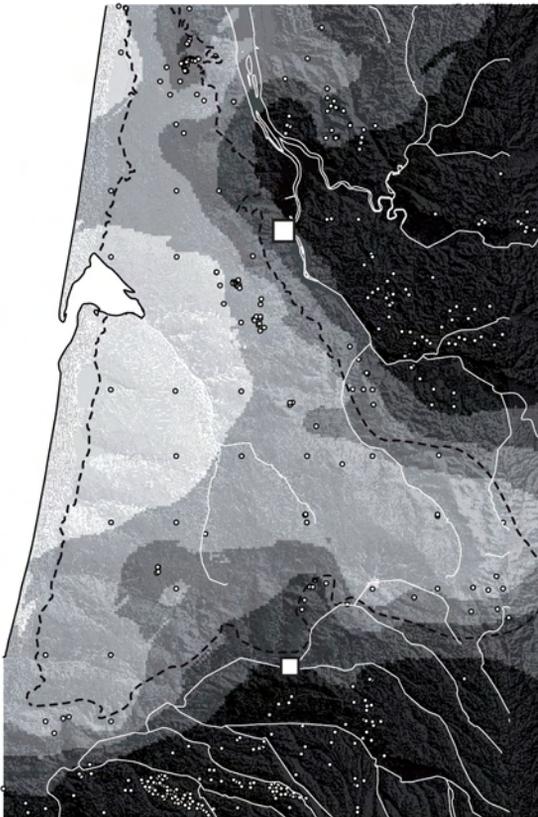
Medium to coarse sand (2000 - 200 μm) %



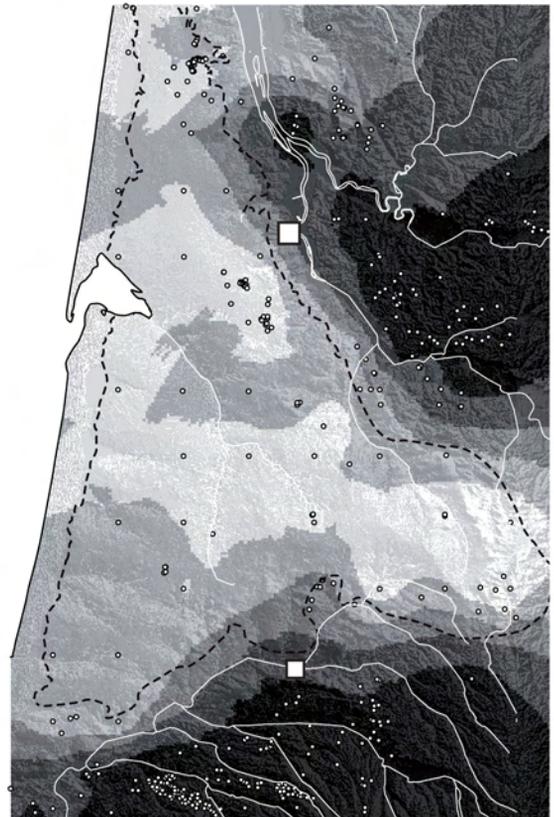
Fine sand (200 - 50 μm) %

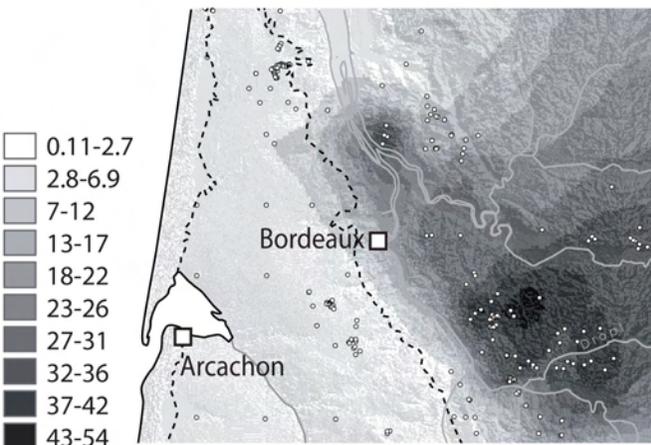
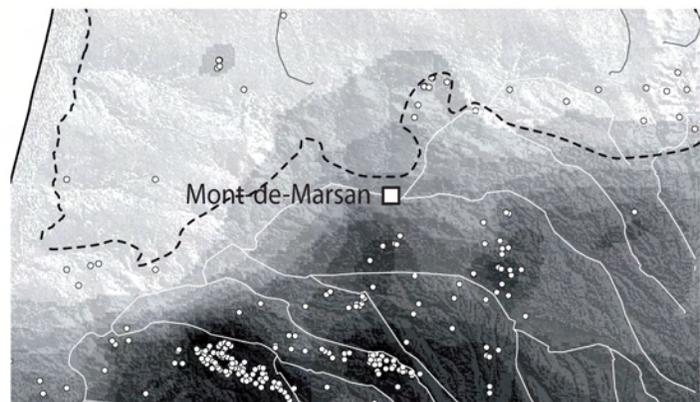
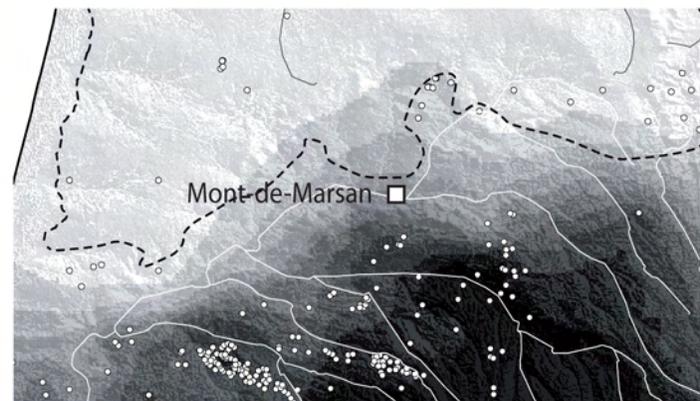
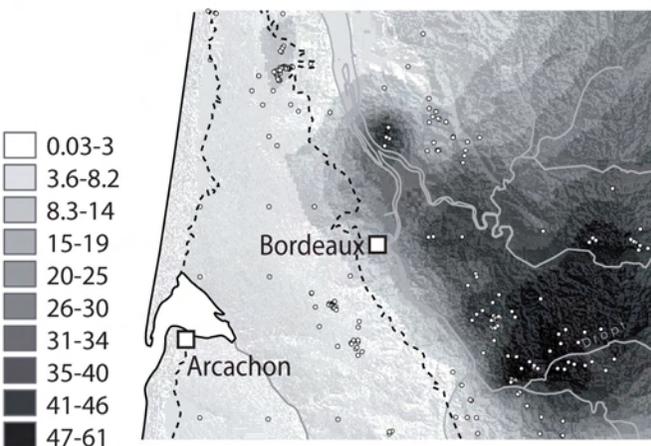


Coarse silt (50 - 20 μm) %



Fine silt (20 - 2 μm) %



A**B**Coarse silt (50 - 20 μm) %Fine silt (20 - 2 μm) %

0 16 km

