

A new interpolation method to measure delta evolution and sediment flux: Application to the late Holocene coastal plain of the Argens River in the western Mediterranean

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1	A new interpolation method to measure delta evolution and sediment flux:
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23	ABSTRACT
24	Rapid environmental changes along the Mediterranean coasts influenced the

25 sedimentary dynamics, shoreline position and human settlements in deltaic areas over

26 the last millennia. An innovative and multiproxy approach using geostatistical modelling 27 was developed to estimate geomorphic evolution and sediment fluxes in deltaic areas, 28 while palaeoecological conditions and environmental changes were assessed from 29 geochemistry and cluster analyses of molluscan fauna. This method was applied to the 30 coastal plain of the Argens River (southern France) in the western Mediterranean. The 31 study of facies associations led to the identification of depositional environments 32 representative of prodelta, delta front, river channel, floodplain, marsh and abandoned 33 channel. The late Holocene sediment flux in the Argens River bayhead delta ranged from $15,800 \pm 2300$ to 52,000 + 8500 m³.yr⁻¹. The highest rates of sediment deposition and 34 35 delta growth occurred between 2500 and 2000 cal yr BP during a period of high flood 36 frequency in the western Mediterranean. A geomorphically-induced decrease in 37 subaqueous sediment flux over the past 2500 years and an increase in fluvial activity 38 from 1600 to 500 cal yr BP were primarily controlled by autogenic processes due to 39 change in accommodation space, delta slope and tributary junctions. A sharp increase in 40 subaerial sediment flux from 500 to 0 cal yr BP was probably driven by hydroclimatic 41 change characterized by more frequent flooding during the Little Ice Age. High sediment 42 fluxes and climate-driven fluvial activity were associated with a southward shift of the 43 North Atlantic westerlies or a northward migration of the intertropical convergence 44 zone.

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46 KEYWORDS

47 Mediterranean delta; Coastal environment; Sediment flux; Elevation modelling;
48 Inorganic geochemistry; Molluscan fauna; Late Holocene; Western Mediterranean
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51 1. INTRODUCTION

52 In the western Mediterranean, the vulnerability of Holocene coastal landscape increased 53 in response to rapid climate change characterized by pervasive millennial-scale 54 variability as well as decadal- to centennial-scale abrupt transitions (Fletcher et al., 55 2013; Fletcher and Zielhofer, 2013; Melis et al., 2018). These rapid climate changes 56 occurred during cooling events characterized by high storm activity and flood 57 frequency, which led to an increased fluvial and coastal sediment supply (Dezileau et al., 58 2011, 2016; Sabatier et al., 2012; Fletcher and Zielhofer, 2013; Degeai et al., 2015, 59 2017). These episodes of intense storm activity and high flood frequency could have 60 been caused by changes in atmospheric circulation patterns such as the North Atlantic 61 oscillation (NAO) or the intertropical convergence zone (ITCZ) (Benito et al., 2015a, 62 2015b; Goudeau et al., 2015; Dezileau et al., 2016; Sanchez-Lopez et al., 2016; Degeai et 63 al., 2017). 64 For instance, a southward shift of the North Atlantic westerlies and storm tracks related 65 to the negative phase of the NAO was associated with cooler and wetter climatic 66 conditions in southwestern Europe during the Little Ice Age (Nieto-Moreno et al., 2011, 67 2013a, 2013b; Trouet et al., 2012; Wirth et al., 2013; Goudeau et al., 2014, 2015; Sanchez-Lopez et al., 2016). An increase in riverine flooding and fluvial activity occurred 68 69 at this time (Macklin et al., 2006; Benito et al., 2015a; Degeai et al., 2017), which led to 70 higher inputs of terrigenous material in the marine sediments of the western 71 Mediterranean basin (Frigola et al., 2007; Nieto-Moreno et al., 2011, 2013a, 2013b; 72 Goudeau et al., 2014, 2015; Jalali et al., 2016). 73 High fluvial sediment supply to coast and river mouths led to widespread coastal plains

and an increased growth of bayhead deltas in incised-valley systems (Grove, 2001;

Amorosi et al., 2009; Anthony et al., 2014). However, the quantitative impact of late

Holocene climate change on the sedimentation rates and fluvial activity of western
Mediterranean deltas is poorly known, although these coastal areas are important to
estimate sediment transport and flux from continent to ocean, or to assess the balance
between onshore erosion and offshore deposition.

80 The evolution of deltas developed in western Mediterranean incised-valley systems was 81 generally characterized by facies associations showing a succession of marine, lagoonal 82 or fluvial conditions over the past millennia (Amorosi et al., 2009, 2013b, 2013c; 83 Bertoncello et al., 2014a; Anthony, 2015). The identification of depositional 84 environment and the study of sedimentary processes and palaeogeographic changes in 85 the western Mediterranean delta plains during the Holocene were generally based on 86 multiproxy analyses of sedimentary sequences from cores or trenches (e.g. Giraudi et al., 87 2009; Bertoncello et al., 2014a; Amorosi et al., 2016; Melis et al., 2018; Devillers et al., 88 2019; Ruiz-Pérez and Carmona, 2019). These methods are effective for coastal 89 palaeoenvironmental reconstruction, but are often poorly suited to the geometric 90 reconstruction of sedimentary units and the measurement of sediment volumes. 91 Geostatistics or numerical modelling were used to estimate palaeoelevation and 92 sediment flux in coastal plains and incised-valley systems over the last millennia (e.g. 93 Koster et al., 2017; Clement and Fuller, 2018). These interpolation methods are very 94 useful in the case of limited age data and can provide important information on soil 95 erosion, sediment flux, mass balance and geomorphic evolution in coastal areas. 96 In this paper, a new quantitative approach was developed to (1) calculate sediment flux 97 associated with the progradation of bayhead deltas in incised-valley systems and (2) 98 decipher the role of climate change and autogenic processes on delta growth. This new 99 method was applied to the coastal plain of the Argens River in southeastern France, 100 which is an area of particular interest for the study of atmospheric circulation patterns

between northern and southern Europe. The depositional environment and sedimentary
dynamics over the past 2500 years were studied using an integrated methodology based
on cores and archaeological excavations, sedimentological analyses (geochemistry,
magnetic susceptibility), malacology and radiocarbon chronology, then geomorphic
change and sediment flux were estimated from geostatistics and elevation modelling.
The variability of sediment flux was discussed in relation to late Holocene hydroclimate
change and geomorphic setting.

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- 109

110 2. ENVIRONMENTAL SETTING

111 The Argens River drains a catchment of ca. 2800 km² located in southeastern France 112 (Fig. 1A). This 114 km long river flows into the western Mediterranean between the Gulf 113 of Lions and the Ligurian Sea along the southwestern Alps. This mountainous coastal 114 area is bordered by a very narrow shelf (< 5 km) and a steep continental slope that 115 stands above the floor of the Provençal Basin at depths greater than 2800 m. The source 116 of the Argens River is at 268 m above sea level (Durozoy et al., 1970), giving an average slope of ca. 2.35% for the modern streambed. Over the last 50 years (1970-2019), the 117 118 hydrological regime at Roquebrune-sur-Argens (Fig. 1B) was characterized by 119 maximum and minimum river discharge rates in winter (32.2 m³.s⁻¹) and summer (5.8 120 m³.s⁻¹), respectively, due to a Mediterranean rainfall regime (DREAL PACA data, 121 www.hydro.eaufrance.fr). 122 The lower valley of the Argens River is bounded on the south by the late Proterozoic to 123 early Carboniferous granitic and gneissic rocks of the Massif des Maures (Fig. 1B), and

124 on the north by the Permian volcano-sedimentary basin of the lower Argens valley

125 (Toutin-Morin et al., 1994). At the end of the Last Glacial Maximum (LGM, ca. 20 cal kyr

BP), sea level dropped more than 100 m along the French Mediterranean coast
(Lambeck and Bard, 2000), leading to an adjustment of the longitudinal profile of the
river by regressive erosion (Bertoncello et al., 2014a). Thus, the lower valley of the
Argens River was deeply incised in Lower Pliocene marine sediments composed of
blueish or grey marl, sand and gravel (Toutin-Morin et al., 1994). This incised-valley
system was then buried by transgressive deposits during the post-LGM sea level rise
(Dubar, 2004).

In the lower valley, the post-LGM deposits form a >100 m thick sedimentary sequence 133 134 composed of (1) fluvial sand and gravel at the bottom, (2) marine sand and clay 135 intercalated with fluvial sand and gravel, and (3) fluvial silt, sand and gravel at the top 136 (Durozoy et al., 1970; Toutin-Morin et al., 1994; Dubar, 2004). The latter deposits were 137 formed by delta progradation and fluvial aggradation, which led to the formation of a 138 coastal plain over the past 4000 years (Dubar, 2004; Bertoncello et al., 2014a). This 139 coastal plain does not exceed 5 m in elevation over a distance of up to 5 km from the sea 140 (Fig. 1B), giving a mean slope of 1‰.

141 Previous works studied the palaeogeographic and environmental changes in the Argens 142 River coastal plain from the Neolithic to the Roman period and their impacts on human 143 settlement (Dubar, 2004; Bertoncello et al. 2014a). The analyses of the palaeogeography 144 at the mouth of the Argens River over the past three centuries from historical maps or 145 archives highlighted a highly mobile coastal environment under the influence of the 146 progradation of the Argens River along with its two main northern tributaries (i.e. the 147 Garonne and Reyran rivers) and the southern branch of this delta system upstream from 148 the Villepey Pond (Fig. 1B) (Cohen, 1997; Bertoncello et al., 2008, 2014a). In fact, these 149 multiple channels formed a deltaic plain, hereafter called the Argens River delta system 150 (ARDS).

151 The ARDS was classified as a bayhead delta that prograded into a protected bedrock 152 embayment with low incident wave energy as a result of significant refraction, 153 diffraction and nearshore dissipation (Anthony et al., 2014; Anthony, 2015). Bayhead 154 deltas prograde in river-mouth setting along the inner part of wave-dominated estuaries 155 (Dalrymple et al., 1992; Amorosi et al., 2005; Anthony et al., 2014), and become 156 increasingly shaped by waves (Anthony, 2015). In the western Mediterranean, wave-157 dominated estuaries were filled with sediments during transgression then transformed 158 into wave-dominated deltas during sea-level highstand (Amorosi and Milli, 2001; Milli et 159 al., 2013). Confined or unconfined bayhead deltas are defined according to the geological 160 and environmental setting: the former prograde into incised-valleys and their geometry 161 is controlled by the morphology of the valley, while the latter prograde into open 162 interdistributary bay and are a part of larger deltaic complexes (Simms and Rodriguez, 163 2015; Simms et al., 2018). The ARDS bayhead delta is confined in the Argens River valley 164 west of Fréjus, and unconfined in the interdistributary bay between Saint-Raphaël and 165 Saint-Aygulf (Fig. 1B). 166 During the Late Holocene, or the Meghalayan (i.e. the last 4200 years b2k), 167 palaeoenvironmental and palaeoecological studies showed that the progradation of the 168 ARDS occurred in a marine depositional environment that gradually transformed into a 169 delta plain during the late phases of sedimentary filling of the lower valley (Allinne et al., 170 2006; Devillers and Bonnet, 2006; Bertoncello et al., 2014a). Pollen data indicate that 171 the first human impact on the landscape of the lower valley of the Argens River could

172 have occurred around 7000 cal yr BP with indices of forest clearing and evidence of

173 cereal cultivation during the late Neolithic period (Dubar et al., 2004; Bertoncello et al.,

174 2014a). After ca. 3000 cal yr BP, human activities led to radical changes in the vegetation

175 cover, high forest clearing and metal pollution in the sediments of the ARDS (Dubar et

176 al., 2004; Bertoncello et al., 2014a; Véron et al., 2018).

177 The past human occupation in the study area was more intense during the Roman

- period, especially from the creation of the Roman colony of *Forum Iulii* (Fréjus) (Fig. 1B)
- in the late first century BCE (Gascou, 1982; Gascou and Janon, 1985). The Roman
- 180 harbour of *Forum Iulii* is now more than 1 km inland from the sea owing to the
- 181 progradation of the ARDS during the last two thousand years (Excoffon et al., 2006,
- 182 2010; Gébara and Morhange, 2010; Bertoncello et al., 2011; Bony et al., 2011; Giaime et
- 183 al., 2019).
- 184
- 185
- 186 3. MATERIAL AND METHODS
- 187 3.1. Cores and trenches

188 A geological database was created from sedimentary cores and archaeological trenches

- 189 (Fig. 1B). It is based on published cores with geochronological and palaeoenvironmental
- data as well as on two new cores (BN1 and BN5) located in the downstream part of the
- 191 ARDS. The 12-m long BN1 core (43°25'22"N, 6°44'02"E, 1.65 m in elevation) was
- sampled with a stationary piston corer. The 18.5-m long BN5 core (43°25'02"N,
- 193 6°43'44"E, 1.67 m in elevation) was sampled with a stationary piston corer from 0 to 3
- m depth, a solid tube sampler equipped with a hydraulic hammer from 3 to 7 m depth,
- and a rotary drill auger from 7 to 18.5 m depth. The sediments at 3-4 m depth in the BN1
- 196 core and at 3.4-4 and 5.63-6.75 m depth in the BN5 core were unsampled due to

197 creeping and difficulty in coring layers with very coarse particles.

198

199 3.2. Molluscan analysis

200 A volume of sediment of ca. 1000 cm³ per sample was collected at an average interval of 201 50 cm along the BN1 and BN5 cores for a malacological study. These samples were 202 sieved using a 1 mm mesh (Sabatier et al., 2008, 2012; Dezileau et al., 2016). A total of 203 2314 mollusc shells were identified and counted under stereomicroscope, representing 204 85 species living in marine (44 species), brackish (7 species), freshwater (17 species) or 205 terrestrial (17 species) environments. The mollusc species were assigned to ecological 206 groups from previous work in the Argens River valley (Devillers and Bonnet, 2006) and 207 from the World Register of Marine Species (www.marinespecies.org), the AnimalBase of 208 the University of Göttingen (www.animalbase.uni-goettingen.de) and the INPN database 209 of the Museum National d'Histoire Naturelle (inpn.mnhn.fr).

210

211 3.3. Geochemistry

212 Geochemical analyses of sediments may provide information about the degree of salinity 213 in subaqueous depositional environment from coastal or lacustrine areas: high (low) 214 values of S, Cl and Sr can evidence brackish (freshwater) conditions (Lopez-Buendia et 215 al., 1999; Chagué-Goff et al., 2002; Schofield et al., 2010; Moreno et al., 2012; Haenssler 216 et al., 2013; Degeai et al. 2015, 2017). Moreover, the bulk of sulphur in coastal lakes can 217 reflect the organic content of a sediment as a function of primary productivity in the 218 water column (Striewski et al., 2009; Haenssler et al., 2013). High ratios of S/Cl may 219 indicate organic-rich layers as S is high in organic-rich sediments (Croudace et al., 2006). 220 Besides, an increase in Si, K and Rb is generally associated with riverine input in the 221 western Mediterranean (Frigola et al., 2007; Martin-Puertas et al., 2010, 2011; Nieto-222 Moreno et al., 2011, 2013a; Rodrigo-Gamiz et al., 2011, 2014; Moreno et al., 2012; 223 Martinez-Ruiz et al., 2015; Degeai et al., 2017). Elemental concentrations of coastal 224 sediments from the western Mediterranean are usually normalized to Al to minimize

matrix effects (Dezileau et al., 2011, 2016; Sabatier et al., 2012). Therefore, we have
used the Cl/Al and Sr/Al ratios as paleosalinity indicators, the Si/Al, K/Al and Rb/Al
ratios as proxy records of riverine input, and the S/Cl ratio as an indicator of organic
matter content.

229 A total of 108 and 138 sediment samples from the BN1 and BN5 cores, respectively, 230 were measured by energy-dispersive X-ray fluorescence (ED-XRF) spectrometry at an 231 average sampling interval of ca. 10 cm with a Delta Innov-X spectrometer equipped with 232 a 4 W Au-tube. Each sample was dried then mechanically crushed and homogenized to a fine powder, which was placed in a crystal polystyrene tube of 22 mm length and inner 233 234 diameter then covered with an ultrafine polyethylene film. The measurement of S, Cl, Rb 235 and Sr was undertaken with the soil analytical mode, while Al, Si and K were measured 236 with the mining analytical mode. The parameters of voltage, amperage and counting 237 times were as follows: 10 kV, 200 µA and 10 s for Al, Si and K; 15 kV, 200 µA and 45 s for 238 S and Cl; 40 kV, 100 µA and 30 s for Rb and Sr. The elemental concentrations were 239 obtained using the Compton Normalization calibration method (EPA, 2007). The 2- σ measurement uncertainty is lower than ± 5% for Si, K, Rb and Sr, and around ± 10% for 240 241 S and Cl.

242

243 3.4. Magnetic susceptibility

In deltaic environment, the magnetic susceptibility can reflect the terrigenous flux

245 derived from fluvial processes (Delile et al., 2015). Higher values of magnetic

susceptibility are generally driven by the presence of Fe-bearing minerals (e.g. biotite)

in the sample (Dearing, 1999).

248 The mass-specific magnetic susceptibility (χ) of sediments from the BN1 and BN5 cores

was measured at an average sampling interval of ca. 10 cm with a Bartington MS2

susceptibilimeter and a MS2B probe. Each powder sample in the tubes used for
geochemical analyses (i.e. 246 samples for BN1 and BN5) was weighed then measured
twice with a time period of 12s, an operating frequency of 4.65 kHz, an amplitude of
applied magnetic field of 250 μT, and a resolution of 1x10⁻⁸ kg.m⁻³ (range 0.1). The
relative standard deviation of both measurements is systematically lower than 3%.

256 3.5. Geochronology

257 The geochronological data are based on a compilation of available radiocarbon ages in 258 the ARDS, including a set of 10 new Accelerator Mass Spectrometry (AMS) ¹⁴C dates on 259 wood and charcoal from the BN1 and BN5 cores (Table 1). The new AMS ¹⁴C ages were 260 performed by the Centre de Datation par le RadioCarbone at Lyon and by the Poznan 261 Radiocarbon Laboratory. The Calib 7.10 software (Stuiver et al., 2019) was used to 262 calibrate the new ¹⁴C dates and to update the calibration of radiocarbon ages published 263 in previous studies. The radiocarbon ages of terrestrial material (seed, wood, charcoal, 264 peat) were calibrated using the IntCal13 calibration curve (Reimer et al., 2013), while 265 the radiocarbon ages of marine shell were calibrated using the MARINE13 calibration 266 curve (Reimer et al., 2013) and a ΔR of -15 years (Siani et al., 2000). The 2- σ uncertainty 267 interval and median probability of calibrated ages are reported in Table 1.

268

269 3.6. Geostatistics

Three 2D age-elevation transects (A1-3, see location in Fig. 1B) and five maps of
sediment thickness in the ARDS were computed for the last 2500 years using the ArcGIS
10 Geostatistical Analyst extension. The geochronological and sediment thickness data
were interpolated using radial basis functions. The computation of rasters of sediment
thickness with a spatial cell of 15 m in size for each 500-year interval over the past 2500

275 years was based on the isochrones of the 2D age-elevation models and on a null 276 thickness assigned to the boundary of Holocene deposits. The models were limited to 277 the last 2500 years because there are not enough ¹⁴C dates without age reversals to 278 make a robust data interpolation before 2500 cal yr BP (Table 1). 279 The palaeoelevation Z_n was then calculated at a 500-year interval from n = 1 (500 cal yr 280 BP) to 5 (2500 cal yr BP) using the ArcGIS 10 raster calculator and equation [1]. For 281 spatial coordinates (x, y), Z_n was obtained by substracting the cumulative sediment 282 thickness T from the modern elevation Z_0 , which is deduced from LiDAR data available 283 for the ARDS (Litto3D, IGN/SHOM, year 2015), and by adding the sea level *S_n* at the time 284 *n* from Vacchi et al. (2016). 285 286 $Z_n(x, y) = Z_0(x, y) - \sum_{i=1}^n T_i(x, y) + S_n$ [1] 287 288 Error estimates for sediment thickness and palaeoelevation were calculated with an 289 uncertainty of ± 100 years for the isochrones of the 2D age-elevation models. This 200-290 year interval was chosen because it closely matches the mean $2-\sigma$ uncertainty interval 291 (225 years) of calibrated radiocarbon ages (Table 1). 292

293

4. RESULTS

295 4.1. Lithology

Both BN1 and BN5 cores were divided into 4 stratigraphic units from bottom (unit 1) totop (unit 4) (Fig. 2, 3).

298 Unit 1 in BN1 core (12-9 m depth) is characterized by alternating 10-20 cm thick layers

of grey to greyish brown clay, silt and fine sand (Fig. 2). Moreover, particulate organic

matter is abundantly present in the matrix of sediments deposited from 11.8 to 10.2 m
depth. In BN5 core, unit 1 (18.5-11.4 m depth) is primarily composed of grey clay and
silty clay alternating with grey clayey silty fine sand at ca. 16 and 14.1-13.4 m depth,
with brownish grey sand at ca. 18.4, 17.5, 16.8 and 16.4 m depth, and with grey sand and
gravel at 16.5 and 12.7-12 m depth (Fig. 3). The clay content decreases above 14.85 m
depth.

306 Unit 2 in BN1 core (9-4 m depth) is mainly composed of grey, greyish brown, brownish 307 grey or brown sand interbedded with silt and sandy silt (Fig. 2). The upper part shows a coarsening-upward trend from silt to medium-coarse sand. Thin organic-rich layers can 308 309 be observed at 8.7-8.6, 8.25-8, 6.3-6.1 and 5.3-5.2 m depth, while charcoals are present 310 around 5.7 m depth. In BN5 core, unit 2 (11.4-6.75 m depth) exhibits a succession of 311 grey sediments characterized by a general coarsening-upward trend showing fine sandy 312 silt and silty fine sand overlaid with fine sand from 8.5 to 8.1 m depth then with clayey 313 medium-coarse sand and rounded gravel from 8.1 to 7.3 m depth. Particulate organic 314 matter appears at ca. 11.3-11.2, 9.6-9.4 and 8.7-8.5 m depth (Fig. 3). The top of unit 2 in 315 BN5 core is composed of organic-rich grey clay between 7.3 and 6.75 m depth with black 316 peat from ca. 7 to 7.2 m depth. This layer is absent in BN1. 317 Unit 3 in BN1 core (3-1.4 m depth) is represented by brown sand and clast-supported

318 rounded gravel showing a general upward-fining trend with fine-medium sand

overlying coarse sand and gravel (Fig. 2). In BN5 core, the coarse sand and clast-

320 supported rounded gravel of unit 3 (5.63-4 and 3.4-2.9 m depth) have a brownish grey

321 or greyish brown colour in the lower part and a brown colour above 4.85 m depth (Fig.

322 3). The bottom of this unit is uncertain in BN5 core given the coring gap from 6.75 to

323 5.63 m depth.

324 Unit 4 in BN1 core (1.4-0 m depth) is composed of brown clay, silt and fine sand with 325 many rootlets and of a 20 cm thick dark brown clayey layer at the bottom (Fig. 2). In 326 BN5, unit 4 (2.9-0 m depth) is mainly composed of brownish grey or greyish brown fine 327 sand interbedded with thin layers of clay and silty clay between 2.9 and 2.2 m depth, of 328 greyish brown clay between 2.2 and 1.5 m depth, and of brown silt, clayey silt or fine 329 sand between 1.5 and 0 m depth (Fig. 3). Small fragments of modern ceramics and a 15 330 cm thick layer of dark brown silt and matrix-supported angular or rounded gravel 331 overlaid with a 4 cm thick layer of black organic silt appear in the upper 40 cm of BN5 332 core. Unit 4 in BN5 core contains particulate organic matter, rootlets and charcoals. 333 Overall, bedding contacts are gradual apart from sand layers which are generally in 334 sharp contact with clay or silt beds. Marine, brackish, freshwater gastropods and 335 bivalves are present in units 1 and 2, while terrestrial gastropods are over-represented 336 in unit 4. Unit 3 is barren of fossils.

337

338 4.2. Molluscan shells

The palaeoecological conditions prevailing in the BN1 and BN5 cores were deducedfrom a multivariate cluster analysis of molluscan assemblages (Fig. 4, 5).

341 Unit 1 is characterized by a salinity-stressed marine environment with low to moderate 342 brackish and freshwater influences. The dominant species of this ecosystem correspond 343 to Turbonilla lactea for BN1 (41% on average) and to Moerella donacina for BN5 (19% 344 on average). Subordinate marine species (1-10% on average) are represented by Abra 345 prismatica, Corbula gibba, Eulima bivittata, Macomangulus tenuis, Smaragdia viridis, 346 *Thracia pubescens, Trivia arctica* and *Trophonopsis muricata* for BN1, and by *A*. 347 prismatica, Acanthocardia sp., Antalis dentalis, C. gibba, Gouldia minima, Lentidium 348 mediterraneum, M. tenuis, Mathilda quadricarinata, Megastomia conoidea, Mimachlamys

349 *varia, Nucula nucleus, Saccella commutata* and *Venus verrucosa* for BN5. A low to

350 moderate mean abundance (1-12% in BN1 or BN5) of molluscs such as *Abra alba*,

351 *Cerastoderma glaucum, Hydrobia acuta, Loripes orbiculatus, Parvicardium exiguum,*

352 Ancylus fluviatilis, Bithynia tentaculata, Islamia spirata and Valvata sp. suggests that the

- 353 marine environment experienced brackish and freshwater conditions with fluvial
- discharges.

355 Unit 2 is characterized by brackish and freshwater environmental conditions with

356 marine influences (Fig. 4, 5). The molluscan assemblages are predominantly composed

of *A. Alba*, *C. glaucum*, *H. acuta* and *Valvata* sp. (> 10% on average in BN1 or BN5). Other

358 subordinate brackish or freshwater species in BN1 or BN5 include *A. fluviatilis*,

359 Belgrandia gibba, B. tentaculata, Euglesa casertana, I. spirata and Theodoxus fluviatilis

360 (1-10% on average). The rare occurrence (< 1%) of terrestrial molluscs usually found in

361 sandy dunes such as *Cernuella virgata*, *Cochlicella conoidea*, *Trochoidea trochoides* and

362 *Xerosecta explanata* suggests reworking of coastal sediment, while a low to moderate

363 mean abundance (1-12% in BN1 or BN5) of *Abra nitida*, *A. prismatica*, *C. gibba*, *L.*

364 mediterraneum, Lucinella divaricata, M. donacina, N. nucleus, Retusa umbilicata, and T.

365 *lactea* highlight marine influences.

366 Unit 3 is barren of molluscan shells, whereas unit 4 is characterized by a terrestrial

367 environment with low to moderate humid conditions (Fig. 4, 5). The most abundant land

368 snails are represented in BN1 and BN5 by species living in dry habitats of coastal areas

369 like sandy dunes or beaches: C. virgata (45 and 30% on average, respectively) and C.

370 *conoidea* (7 and 43% on average, respectively). The most frequent subordinate

371 terrestrial species (1-10% on average) correspond to gastropods living in dry vegetation

372 or on sandy dunes in coastal areas (*Cochlicella acuta*, *Trochoidea pyramidata*, *T*.

373 *trochoides, X. explanata*). Other uncommon land molluscs (< 6% on average) live in

374 either humid habitats (*Cecilioides acicula, Oxychilus draparnaudi, Oxyloma elegans,*

375 *Vallonia pulchella*) or open habitats with very variable humidity (*Vertigo pygmaea*). The

376 main species of freshwater molluscs are *B. tentaculata, Galba truncatula, I. spirata,*

377 *Musculium lacustre* and *Valvata* sp. (1-10% on average). A rare occurrence (< 3%) of *H*.

acuta is found between 1.5 and 1 m depth in BN1, while the absence of marine shells isobserved in both BN1 and BN5 cores.

380

381 4.3. Geochemical data

382 Unit 1 generally exhibits high values of Cl/Al and Sr/Al, which reflect high levels of 383 salinity (Fig. 2, 3). However, higher than average riverine inputs (K/Al and Rb/Al but not 384 Si/Al) accompanied with lower values of Cl/Al and Sr/Al at the bottom of BN1 and BN5 385 cores could reveal fluvial influence in a marine depositional environment. The layers of 386 brownish grey sand or grey sand and gravel in unit 1 of BN5 are associated with peaks 387 of Si along with low levels of salinity (Fig. 3). These coarse sediments could reflect local 388 and brief episodes of high-energy fluvial input. The high content of S at the transition 389 between units 1 and 2 from 9.3 to 8.1 m depth in BN1 is associated with abundant tabular crystals of euhedral gypsum in the sand fraction (Fig. 2). 390 391 Unit 2 is characterized by a general decrease of salinity indicators paralleled by an 392 increase in riverine input indicators (Fig. 2, 3). The values of elemental ratios in the

393 upper part of unit 2 show low levels of salinity and high riverine input in both BN1 and

BN5 cores. In this latter, the organic-rich clayey layer at the top of unit 2 exhibits high

ratios of K/Al, Rb/Al and S/Cl, while the highest concentration of S is found in the black

396 peat horizon between ca. 7.2 to 7 m depth.

397 Unit 3 presents geochemical and sedimentological features characterized by high Si/Al
398 and K/Al ratios, low salinity indices, and coarse sand and gravel most likely deposited in
399 high-energy environment (Fig. 2, 3).

400 Unit 4 in BN1 core shows higher than average riverine discharge and very low levels of 401 salinity (Fig. 2). Large peaks of K/Al and Rb/Al are associated with a 20 cm thick layer of 402 dark brown clayey sediment at the bottom of this unit. Besides, a major peak of S/Cl is 403 found at the transition between units 3 and 4 in BN1 core, and is perhaps indicative of 404 higher organic matter content. In BN5 core, the lower half of unit 4 between 2.9 and 1.55 405 m depth generally shows higher than average Rb/Al and K/Al ratios, low Si/Al ratio, and 406 peaks of Sr/Al, whereas lower than average indicators of riverine input and salinity are 407 most often found in the upper half of this unit above 1.55 m depth (Fig. 3). 408 Overall, high (low) levels of salinity in both BN1 and BN5 cores generally correspond to 409 low (high) levels of riverine input. Hence, the variability of salinity indicators in the

410 ARDS could have been mainly driven by fluvial activity. Besides, the geochemical

411 indicators of riverine discharge seem to have been influenced by the grain size of

412 sediments. High values of Si/Al are found only in sand and gravel, and more specifically

413 in the coarse sand and gravel fraction (Fig. 2, 3). High values of Rb/Al are observed only

414 in clay and silt, while high values of K/Al can be recorded in coarse or fine sediments. K

and Rb are commonly associated with detrital clay (Croudace et al., 2006). However, the

416 K/Al ratio is also influenced by the contribution of detrital K-feldspar, which may be

417 significant in the coarse fraction (Martinez Ruiz et al., 2015).

418

419 4.4. Magnetic susceptibility

420 The values of mass-specific magnetic susceptibility (χ) are generally lower than 10×10^{-8} 421 m⁻³.kg⁻¹ in BN1 and BN5 (Fig. 2, 3). Higher than average χ values are found in the basal

part of both cores (unit 1) from 12 to ca. 10 m depth in BN1 and from 18.5 to 16.9 m 422 423 depth in BN5. χ values ranging between ca. 10 and 20x10⁻⁸ m⁻³.kg⁻¹ appear from 5.8 to 4.5 m depth in unit 2 of BN1 and from 3.3 to 2 m depth at the transition between units 3 424 and 4 in BN5. The highest χ values around 30×10^{-8} m⁻³.kg⁻¹ are found at the top of BN1 425 426 and BN5 cores. 427 Higher χ values in deltaic environment should reflect a terrigenous supply from catchment by fluvial processes (Delile et al., 2015). However, the riverine sediments of 428 429 unit 3 are characterized by low χ values (Fig. 2, 3). In fact, the mineralogical composition 430 of coarse sand and gravel can preclude the use of magnetic susceptibility as a reliable 431 proxy record of fluvial activity. The low χ values of sediments from unit 3 suggest that 432 these coarse alluvial deposits are mainly composed of diamagnetic material such as 433 quartz (Dearing, 1999). High values of Si and Si/Al as in unit 3 can be indicative of high 434 relative content of quartz (Martinez-Ruiz et al., 2015). This is confirmed by 435 stereomicroscopic observations, which show that the alluvial deposits of unit 3 exhibit 436 an important content of quartz for particles coarser than 1 mm. 437 By contrast, the high χ values of the fine sediments in the upper 1.5 m of BN5 are not 438 related to high values of elemental ratios indicative of riverine input. In fact, this 439 increase in magnetic susceptibility could have been caused by post-depositional 440 oxidation and weathering of sediments at the top of the BN5 core, which is characterized 441 by a terrestrial environment as evidenced by molluscan fauna (Fig. 3, 5). 442 443 444 **5. PALAEOENVIRONMENTAL INTERPRETATION**

445 Delta systems in the northwestern Mediterranean are often characterized by a vertical
446 stacking of mid-late Holocene prodelta, delta front and delta plain deposits (Dubar and

Anthony, 1995; Amorosi et al., 2008, 2013c). These deltaic sequences are interpreted as 447 448 highstand systems tracts deposited during the deceleration of sea level rise after 7500 449 cal yr BP (Stanley and Warne, 1994; Somoza et al., 1997; Vacchi et al., 2016). 450 The depositional environments of the BN1 and BN5 sedimentary successions can be 451 defined on the basis of a comparison of molluscan assemblages, lithology, sediment 452 properties and geochemical data (Fig. 2-5) with facies associations described for the 453 northwestern Mediterranean deltas of (1) the Argens River (Allinne et al., 2006; 454 Devillers and Bonnet, 2006; Bertoncello et al., 2014a), Hérault River (Devillers et al., 455 2019), Rhône River (Amorosi et al., 2013c) and Var River (Dubar and Anthony, 1995) in 456 southern France, (2) the Arno River (Amorosi et al., 2009, 2013a, 2013b), Biferno River 457 (Amorosi et al., 2016), Po River (Amorosi et al., 2008), Posada River (Melis et al., 2018) 458 and Tiber River (Giraudi et al., 2009) in Italy, and (3) the Ebro River (Somoza et al., 459 1997) and Turia River (Ruiz-Pérez and Carmona, 2019) in eastern Spain (Fig. 1A). 460 Holocene prodeltas in the northwestern Mediterranean are characterized by (1) 461 salinity-stressed and strongly fluvial-influenced marine environment with abundant 462 marine gastropods and mixed brackish-normal saline conditions, and (2) clay-sand 463 alternations showing an upward decrease in clay content accompanied with an increase in the frequency and thickness of sandy layers (Somoza et al., 1997; Amorosi et al., 2008, 464 465 2009, 2013b, 2013c). In both BN1 and BN5 cores, unit 1 is mostly composed of grey to 466 brownish grey clay and silty clay with intercalations of sand layers deposited in a 467 marine environment with high levels of salinity (Fig. 2, 3). From the bottom to the top of 468 unit 1, the sandy layers are thicker or more frequent and clay content decreases. In BN5 469 core, the presence of massive mud below 17 m depth and the increasing sand content 470 above this depth fit well with the bottom sediment distribution of the modern inner 471 shelf environment along the ARDS, which shows a minimum upper limit of onshore

blue-grey mud and a minimum lower limit of nearshore terrigenous sand at ca. 15 m
below sea level (Toutin-Morin et al., 1994). The abundance of marine molluscs ranges
from ca. 45 to 75% in BN1 core and from ca. 50 to 95% in BN5 core. Moreover, riverine
inputs highlight a marine environment with fluvial influence. Hence, unit 1 is defined as
a prodelta.

477 In the northwestern Mediterranean, the sedimentary features associated with the 478 transition from prodelta to delta front typically exhibits an increasing proportion of 479 sand, with a delta front composed of an upward-coarsening sand body deposited in high-energy environment (Somoza et al., 1997; Amorosi et al., 2008, 2013c). The lower 480 481 and middle parts of unit 2 in BN1 and BN5 cores are characterized by many layers of 482 sandy silt or fine sand, while the upper part shows upward-coarsening sand bodies in 483 which the sand fraction changes from fine sand to medium-coarse sand from bottom to 484 top (Fig. 2, 3). Unit 2 is also characterized by a decrease in marine shells (from ca. 50 to 485 0%) paralleled by a decrease in Cl/Al and an increase of riverine input (Fig. 2, 3). This 486 unit can be considered as a ca. 5 m thick delta front, which is consistent with the delta 487 front thickness of 5.5 and 7 m in PL1 and PAL1 cores, respectively (Fig. 6). In BN5 core, 488 the delta front is overlaid with organic-rich sediments and black peat, which were likely 489 deposited in a marsh environment (Fig. 3).

The sediments of unit 3 are composed of poorly sorted coarse sand and clast-supported
gravel devoid of fossils (Fig. 2, 3). Fining-upward trends are found in unit 3 at BN1 and
from the top of unit 3 to the lower half of unit 4 at BN5. These sedimentological and
biological features are typical of high-energy environments such as fluvial channels
(Dubar and Anthony, 1995; Somoza et al., 1997; Amorosi et al., 2009, 2013a, 2013b,
2013c, 2016; Giraudi et al., 2009; Melis et al., 2018; Devillers et al., 2019; Ruiz-Pérez and
Carmona, 2019). Moreover, a deposition of the coarse sediments of unit 3 in a river

497 channel is consistent with the geochemical data, which show very low levels of salinity498 and high values of Si/Al and K/Al indicative of riverine input.

499 Unit 4 in BN1 core and the upper half of unit 4 in BN5 core are mainly composed of 500 bioturbated brown clay, silt and fine sand with land molluscs and very low levels of 501 salinity (Fig. 2, 3). These sedimentological and palaeoecological features are compatible 502 with a depositional environment in a well-drained floodplain occasionally affected by 503 river floods (Somoza et al., 1997; Allinne et al., 2006; Devillers and Bonnet, 2006; 504 Amorosi et al., 2013a, 2016; Bertoncello et al., 2014a; Melis et al., 2018; Devillers et al., 505 2019; Ruiz-Pérez and Carmona, 2019). The delta plain of the ARDS includes both 506 floodplain and river channels. More specifically, the malacological data point to a delta 507 plain environment subject to subaerial exposure, given the high abundance of terrestrial 508 species living in coastal habitats such as Cernuella virgata, Cochlicella conoidea, 509 Trochoidea trochoides and Xerosecta explanata at the top of BN1 and BN5 cores (Fig. 4, 510 5). Land snails and freshwater molluscs associated with humid or aquatic environment 511 are mostly found in the lower part of unit 4 and are probably indicative of a depositional 512 setting more frequently subjected to riverine flooding or hydromorphic conditions. 513 In the lower half of unit 4 in BN5 core, thin layers of brownish grey to greyish brown 514 sand and silt (ca. 2.9-2.2 m depth) then organic-rich greyish brown massive clay with 515 rootlets at the top (ca. 2.2-1.5 m depth) were deposited on the coarse fluvial sediments 516 of unit 3 (Fig. 3). These sediment properties and the stratigraphic relationship with 517 fluvial-channel sediments are typical of alluvial sub-environments close to river channel 518 (Amorosi et al., 2013a, 2016). Besides, the organic-rich sediments of the lower half of 519 unit 4 in BN5 core show a general fining-upward trend such as in abandoned channel 520 fills, which are characterized by an initial proximal fill during the abandonment stage 521 then by a distal fill collected in the abandoned palaeochannel during the disconnected

522 stage (Toonen et al., 2012). These two stages of channel-fill development could explain 523 the sediment deposition at ca. 2.9-2.2 and 2.2-1.5 m depth, respectively, in BN5 core. 524 Higher than average values of Sr and Sr/Al are recorded in the lower half of unit 4 in 525 BN5 core and could be due to episodes of shallowing and salinization of water bodies in 526 an abandoned fluvial channel, for example during hydroclimatic changes. Indeed, 527 enhanced Sr may indicate the presence of high-Sr aragonite which requires a shallow-528 water source (Croudace et al., 2006). Besides, higher Sr/Al ratios were recorded by 529 Holocene lacustrine deposits in the western Mediterranean during periods of evaporite 530 precipitation due to lower lake levels characterized by Sr-rich sediments (Martin-531 Puertas et al., 2011). 532 533 534 6. NUMERICAL MODELLING 535 6.1. 2D age-elevation models 536 An accurate chronology of deltaic sequences can be difficult to determine because of ¹⁴C 537 age reversals due to sediment storage and remobilization of older deposits during 538 downslope alluvial transport that prevails in fluvial and delta plains (Stanley and Hait, 539 2000; Stanley, 2001). 540 The median probability of 10 new radiocarbon dates obtained for the BN1 and BN5 541 cores ranges from 3565 to 440 cal yr BP (Table 1). However, the 4 radiocarbon ages of 542 charcoals and wood sampled in delta front deposits are too old and show a reversal of 543 the median probability (Fig. 2, 3). This seems to indicate that reworking of older 544 terrestrial organic matter from the catchment was more important in the delta front 545 depositional environment. By excluding these 4 radiocarbon ages, the sediments from

the BN1 and BN5 cores would have been deposited approximately during the last 2500years.

548 Amongst the 41 radiocarbon dates compiled in the ARDS, we discarded 12 dates (ca. 549 30% of total) showing age reversals with a median probability older than expected 550 (Table 1). These reversal patterns are probably due to the remobilization and 551 redeposition of older material (Stanley and Hait, 2000; Stanley, 2001). The cores from 552 the ancient harbour of Forum Iulii (FIX, FXI, SC6) (Fig. 1B) were not used owing to 553 potential sedimentary hiatuses caused by possible dredging in this port basin during the 554 Roman period (Bony et al., 2011). Besides, several ¹⁴C dates from the PL1 core exhibit 555 large 2- σ uncertainty intervals higher than 300 years because of a very large plateau of 556 the calibration curve from ca. 800 to 400 cal yr BCE (i.e. a conventional ¹⁴C age around 557 2450 BP), the so-called Hallstatt plateau (Fontugne, 2004; Van der Plicht, 2004). Finally, 558 a total of 29 radiocarbon ages were used to compute three 2D age-elevation models in 559 the ARDS: one longitudinal age profile (A1) and two transverse age profiles (A2 and A3) 560 across the valley (Fig. 6). The isochrones are displayed with a 100-year resolution. 561 Overall, profile A1 shows a downstream thickening of deposits from PL1 to BN5 over the 562 last 2500 years. The thickest part is located between PAL1 and BN5 where a mean 563 sediment thickness of ca. 18 m was deposited during the last 2500 years. The highest 564 accumulation occurred between 2500 and 2000 cal yr BP in this part of profile A1. 565 Profile A2 and the northwestern part of profile A1 exhibit a lower sediment 566 accumulation with a thickness of ca. 5 m upstream of the ARDS during the last 2500 567 years. Profile A3 shows that the sediments deposited downstream of the coastal plain 568 over the past 2500 years are thicker in the central part of the valley around BN5 and 569 thinner on the southern edge of the ARDS, with a thickness of ca. 7 m in the VIL2 core 570 (Fig. 6). On the northern side, the trench dug during the Théâtre d'agglomération T2

571 (TAT2) archaeological excavation revealed that Permian rocks were buried under
572 coastal sediments during the second half of the first century BCE (Excoffon et al., 2006;
573 Devillers et al., 2007). These Permian rocks were then covered by 2-3 m of sediments
574 over the last 2000 years.
575 Although the lack of chronological data in the easternmost part of the ARDS coastal plain
576 does not allow the age models to be robustly estimated in this area, LiDAR data were

577 shifted along the BN5 age-elevation model in order to propose a hypothetical

reconstruction of the isochrones of the unconfined bayhead delta that prograded onto

the inner shelf (Fig. 6). The depositional architecture and stacking pattern east of BN5

are thus very similar to those of steep-face deltas studied by Dubar and Anthony (1995)

along the southern French Alps, such as the Var River delta located at ca. 70 km

582 northeast of the Argens River (Fig. 1A).

583

584 6.2. Sediment thickness

585 The numerical models of sediment thickness of the Argens River bayhead delta cover an 586 area of 12.2 km² (Fig. 7). These spatial models were focused on the incised-valley system 587 west of Fréjus, in which the eastward progradation of the bayhead delta of the Argens 588 River was confined and controlled by the morphology of the valley. Moreover, the lack of 589 data in the easternmost part of the ARDS coastal plain prevents the models to be 590 extended east of Fréjus. Therefore, the eastern limit of the modelled area corresponds 591 approximately to the eastern boundary of the Argens River valley between Fréjus and 592 the eastern side of the Massif des Maures.

593 The thickness of sediments deposited from 2500 to 2000 cal yr BP ranges from < 1 to 8

m in the study area, with maximum values around the PAL1 site (Fig. 7A). The

595 depocenter was located in the central part of the lower valley of the Argens River. From

596 2000 to 1500 cal yr BP, the depocenter moved downstream in an eastward direction

597 (Fig. 7B). The main depositional area was located between BN5 and BN1, with a

598 maximal thickness of deposits estimated between 4 and 5 m. Another area of moderate

accumulation ranging from 1 to 3 m in thickness can be observed along a northwest-

- 600 southeast direction of the valley from PL1 to PAL1.
- After 1500 cal yr BP, the maximum accumulation of sediments was lower than 4 m for

602 each 500-year period and the depocenter stayed close to the modern mouth of the

Argens River (Fig. 7C-E). The depocenter moved southward from 1500 to 1000 cal yr BP

and shows a maximum thickness of 3-4 m at BN5 (Fig. 7C). The sediment thickness

pattern between 1000 and 500 cal yr BP was quite similar to the previous period, with

606 maximum values of 2-3 m (Fig. 7D). From 500 to 0 cal yr BP, the highest sediment

accumulation occurred in two areas located, on the one hand, between Verteil and ESC1,

and, on the other hand, close to the mouth of the Argens River around BN5 (Fig. 7E).

609

610 6.3. Elevation modelling and coastal palaeogeography

Deltaic areas with elevation above or below sea level correspond to subaerial or
subaqueous delta, respectively (Anthony, 2015). The past sea levels are taken from
Vacchi et al. (2016) (Table 2).

At 2500 cal yr BP, the elevation model reveals that the lower valley of the Argens River was largely occupied by a marine bay with a subaqueous delta extended until the Verteil site and a prodelta/delta front transition located between PL1 and PAL1 (Fig. 8A). The reduction in the surface area of the subaqueous delta between 2500 and 2000 cal yr BP was probably due to the progradation of the Argens, Reyran and Garonne rivers, which would have led to an eastward displacement of the prodelta/delta front transition between PAL1 and BN5 and to a decrease of the maximum depth of the marine bay from ca. 15 to 10 m (Fig. 8A-B). Then the transition between the delta front and the prodelta
moved east of BN5 after 2000 cal yr BP (Fig. 8B-C).

623 During the first millennium CE, elevation models suggest that land emergence seems to 624 have been more important in the northern part of the ARDS, while subaqueous delta 625 environment would have persisted on the southern side of the ARDS between PAL1, 626 BN5 and VIL2 at the same time (Fig. 8B-D). The progradation in the southern part of the 627 ARDS occurred mainly during the last millennium and would have been characterized by 628 the formation of a fan-delta that grew between BN5 and VIL2 (Fig. 8D-E). Historical 629 maps show that the downstream flow of the Argens River was divided into two branches 630 at least from the end of the 16th century CE to the end of the 18th century CE (Cohen, 631 1997; Bertoncello et al., 2008, 2014a). In addition to the southern branch of the Argens 632 River, the Villepey River might also have contributed to the progradation in the southern 633 part of the delta system, as shown by maps from the 18th century CE (Bertoncello et al., 634 2008, 2014a).

635 The transition from subaqueous to subaerial delta at PAL1 and BN1 occurred between 636 1000 and 500 cal yr BP. At 500 cal yr BP, the elevation model suggests that a large part 637 of the delta plain was above sea level, while a shallow body of water (< 1 m depth) 638 oriented in a NNE-SSW direction persisted along the coastline (Fig. 8E). Moreover, a 639 seaward progradation of ca. 1 km on average should have occurred between BN5 and 640 VIL2 during the last 500 years, which yields a mean progradation rate of 2 m.yr⁻¹. 641 The present topography of the coastal plain exhibits an area of lower elevation between 642 ESC1 and PL1 around the Garonne River, which is surrounded by two levees along the 643 Reyran River north of PAL1 and along the Argens River between Verteil and PAL1 (Fig. 644 1B). The elevation models suggest that these topographic features were largely 645 inherited over the past 2000 years (Fig. 8B-E). More specifically, the small shallow

- 646 depression above sea level south of ESC1 at 500 cal yr BP (Fig. 8E) is consistent with the
- 647 palaeoecological and chronological data from the ESC1 core that shows the presence of a

648 freshwater marsh from ca. 700 to 330 cal yr BP (Allinne et al., 2006).

649

650

651 7. DISCUSSION

- 652 7.1. Sediment flux and fluvial activity
- 653 The elevation models were used to estimate late Holocene sediment fluxes in the ARDS
- (Fig. 9a). The total sediment flux in the confined bayhead delta was greatest from 2500
- to 2000 cal yr BP, with a deposition rate estimated at 52,000 ± 8500 m³.yr⁻¹. A decrease

656 in total sediment flux occurred after 2000 cal yr BP, with rates ranging from 15,800 ±

657 2300 to 26,700 ± 4100 m³.yr⁻¹. The sediment flux in the unconfined bayhead delta along

658 the modern delta front east of BN5 (Fig. 6) shows a general decrease from 47,900 ±

659 9600 to 27,000 ± 5400 m³.yr⁻¹ over the last 2500 years (Fig. 9a).

660 Subaerial and subaqueous sediment fluxes in the confined bayhead delta were estimated661 by calculating the volume of sediments above and below sea level, respectively. The

662 subaqueous sediment flux followed a similar trend to the total sediment flux. More

663 specifically, the rates of subaqueous sediment flux declined sharply from 41,200 ± 8200

664 to $18,700 \pm 3700 \text{ m}^3.\text{yr}^{-1}$ between 2500 and 1500 cal yr BP, then from 6200 \pm 1200 to

665 700 ± 100 m³.yr⁻¹ during the last millennium. Overall, the subaerial sediment flux

666 exhibits an inverse trend with rates ranging from 8000 ± 1600 to $10,700 \pm 2100$ m³.yr⁻¹

between 2500 and 500 cal yr BP, then increasing to $17,800 \pm 3600 \text{ m}^3.\text{yr}^{-1}$ between 500

and 0 cal yr BP (Fig. 9a). Moreover, it is worth noting that the rate of subaerial sediment

flux from 2500 to 2000 cal yr BP (10,700 \pm 2100 m³.yr⁻¹) was slightly higher than that

670 from 2000 to 1500 cal yr BP ($8000 \pm 1600 \text{ m}^3.\text{yr}^{-1}$).

671 The subaqueous sediment flux was ca. 2 to 4 times higher than the subaerial sediment 672 flux between 2500 and 1500 cal yr BP, while the latter exceeded the former during the 673 last millennium. The subaerial sediment flux was 25 times higher than the subaqueous 674 sediment flux over the past 500 years. In fact, the rate of subaerial sediment flux 675 increased sharply between 500 and 0 cal yr BP and was two times higher than during 676 the period from 1000 to 500 cal yr BP. This sharp increase was associated with a slight 677 increase in the total sediment flux of the confined bayhead delta between 500 and 0 cal 678 yr BP, although this latter should be interpreted with caution given uncertainties of flux 679 rates. Besides, the sediment flux of the unconfined bayhead delta does not seem to have 680 significantly changed during the last millennium (Fig. 9a).

681 The high rates of sediment flux during the late Iron Age Cold Period (IACP) and the early 682 Roman Warm Period (RWP) between 2500 and 2000 cal yr BP were accompanied by 683 high levels of fluvial activity in the easternmost part of the lower Argens valley, as 684 suggested by the geochemical indicators of riverine input from the BN1 and BN5 cores 685 (Fig. 9b-c). The decrease in sediment fluxes between 2000 and 1500 cal yr BP was 686 contemporaneous with a period of low fluvial activity in the second half of the RWP. 687 Then the fluvial activity increased and was generally higher than average during the 688 Dark Ages Cold Period (DACP), the Medieval Warm Period (MWP) and the Little Ice Age 689 (LIA), except for BN5 over the last 300 years.

The periods of higher than average fluvial activity in the ARDS during the late Holocene were most often associated with high riverine flooding in the southern French Alps and, more generally, with more frequent river floods and extreme fluvial events in southeastern France (Fig. 9b-f). Overall, the Mediterranean deltas grew slowly during the Middle Ages, then high fluvial sediment supply to coast led to their rapid growth after 500 cal yr BP (Anthony et al., 2014), i.e. during the LIA and 20th century global warming. In the western Mediterranean, the LIA was generally characterized by more
frequent flooding and higher fluvial activity (Barriendos Vallve and Martin-Vide, 1998;
Wilhelm et al., 2012; Wirth et al., 2013; Benito et al., 2015a; Degeai et al., 2017). Besides,
Gutiérrez-Elorza and Pena-Monné (1998) found that climate-driven sediment
accumulation in valleys from northeastern Spain was more widespread during the IACP
than during the LIA, which is consistent with the pattern of total sediment flux in the
ARDS (Fig. 9a).

703

704 7.2. Autogenic control

Autogenic processes play an important role in controlling the stratigraphic architecture
of bayhead delta (Simms et al., 2018). For fluvio-deltaic systems developed in incised
valleys, geomorphologically-driven autogenic processes such as channel avulsion,
meander cutoff, increasing levee height and tributary junction can have been controlled
by changes in accommodation space and valley gradient, confluence of tributary valleys,
or channel extension (Vis et al., 2010; Amorosi et al., 2013a; Simms and Rodriguez,
2015; Simms et al., 2018).

712 In the lower valley of the Argens River, the evolution of subaqueous sediment fluxes 713 paralleled a progressive decrease in the mean slope of the longitudinal valley gradient 714 and in the volume of the accommodation space, which were higher than 2.8‰ and 20 715 Mm³, respectively, between 2500 and 2000 cal yr BP (Fig. 10a-c). The general evolution 716 of the rates of subaqueous sediment deposition in the ARDS could therefore have been 717 primarily driven by geomorphic factors over the last 2500 years. In general, lower rates 718 of sediment flux into estuarine systems and late stages of sedimentary fill in a largely 719 infilled estuarine basin, i.e. in a shallow subaqueous environment, are controlled by 720 limited accommodation space rather than sediment flux (Clement and Fuller, 2018).

More specifically, the lowest rate of subaqueous sediment flux in the incised-valley
system of the Argens River would have been driven by a very limited accommodation
space over the last 500 years (Fig. 10b).

724 The sediment flux and fluvial activity associated with the marine environment of 725 prodelta in unit 1 (Fig. 2, 3), which spans the period from ca. 2700 to 1800 cal yr BP 726 from the age-elevation models (Fig. 6), were high in the late IACP then decreased during 727 the first half of the RWP (Fig. 10c-e). This is inconsistent with the environmental 728 evolution of western Mediterranean prodeltas, which generally show an upward 729 increase in riverine influence (Amorosi et al., 2008, 2013c). Therefore, the period of high fluvial activity and sediment flux between 2500 and 2000 cal yr BP could be indicative of 730 731 an increased fluvial sedimentation caused by allogenic forcing rather than by autogenic 732 control during delta progradation.

By contrast, the increasingly strong riverine input in the easternmost part of the Argens
valley from ca. 1650 to 1300 cal yr BP corresponds to the formation of a delta front,
which suggests that sediment flux was mainly controlled by autogenic processes during
this period (Fig. 10e). Besides, a regional increase in riverine flood frequency occurred
in southeastern France during the DACP (Fig. 9d-f). Thus, the fluvial sediment supply
associated with the advancing delta front could have been strengthened by an increase
in riverine input in response to allogenic forcing (e.g. climate change).

The high fluvial activity of the DACP was maintained in the MWP from ca. 1100 to 600

cal yr BP (Fig. 10e). During the period between 1500 and 500 cal yr BP, the elevation

models suggest that delta progradation led to tributary junctions of the Argens, Reyran

and Villepey Rivers in the easternmost part of the lower Argens valley (Fig. 8C-E). In

744 general, the progradation rates of bayhead deltas increase when crossing

745 geomorphologically and environmentally controlled thresholds such as tributary

junctions in the absence of allogenic forcing mechanisms (Simms and Rodriguez, 2015). 746 747 Moreover, the fluvial sediments filling the river channels identified in BN1 and BN5 748 cores were deposited between ca. 1350 and 450 cal yr BP and between ca. 1000 and 550 749 cal yr BP, respectively (Fig. 2, 3, 6). Thus, the period of high fluvial activity during the 750 MWP could have been driven by autogenic processes such as tributary junction or 751 channel avulsion. Furthermore, each flood record in the southern Alps exhibits at least 752 one episode of more frequent riverine flooding during the MWP (Fig. 9d-e). Therefore, it 753 cannot be ruled out that the high levels of fluvial activity in the ARDS were partially 754 controlled by allogenic forcing during the MWP. 755 From 500 to 0 cal yr BP, the formation of the floodplain in the easternmost part of the 756 lower Argens valley was associated with the highest rate of subaerial sediment flux and 757 with higher than average levels of fluvial activity (Fig. 10d-e), although the mean slope 758 and accommodation space were very low (Fig. 10a-b). Besides, this period was 759 characterized by high flood frequency and fluvial activity in southeastern France (Fig. 760 9d-f). All this suggests that fluvial sedimentation in the ARDS was closely related to 761 allogenic factors during the LIA.

762

763 7.3. Impact of climate change

The impacts of rapid climate change on vegetation cover and hydrological regime are
allogenic factors that may have controlled the fluvial and coastal sedimentation in the
western Mediterranean during the Holocene (Fletcher and Zielhofer, 2013). At this time,
the periods of higher flood frequency in this region coincided with transitions to cool
and wet climatic conditions during persistent negative phases of the North Atlantic
oscillation (NAO) (Wirth et al., 2013; Benito et al., 2015a).

770 The NAO is one of the most prominent and recurrent modes of atmospheric circulation 771 variability in Europe (Hurrell et al., 2003). A southward (northward) shift of the North 772 Atlantic westerly winds and storm tracks during the negative (positive) phase of the 773 NAO leads to a wetter (drier) climate in southern Europe and to a drier (wetter) climate 774 in northern Europe (Hurrell, 1995; Serreze et al., 1997; Osborn et al., 1999; Trigo et al., 775 2002). A well-known speleothem record of late Holocene climate change in northwest 776 Scotland (Fig. 1A) was supposedly driven by the bipolar seesaw of storminess and 777 humidity related to the NAO (Proctor et al., 2002; Trouet et al., 2009). In the lower 778 Argens valley, the episodes of high fluvial activity in the late IACP, the first half of the 779 RWP, the DACP and the LIA occurred during dry periods in northern Europe and should 780 thus have been associated with a southward shift of the westerly winds typically found 781 in a negative phase of the NAO (Fig. 10e-g, 11A), except for the MWP during which the 782 high fluvial activity was mainly driven by autogenic control due to delta progradation 783 (see section 7.2). However, increased flood frequencies in Europe may also have been 784 influenced by other small-scale flow patterns in the atmosphere (Jacobeit et al., 2003). 785 Hence, we undertook an empirical orthogonal function (EOF) analysis of winter 786 (December to February) sea level pressure (SLP) using the 500 yr data set from 787 Luterbacher et al. (2002) in order to highlight atmospheric circulation patterns in the 788 North Atlantic and Europe (Fig. 11). The respective contributions of the three first EOFs 789 to the total variance are estimated at 51, 20 and 14%. The leading EOF mode (EOF1) 790 exhibits a north-south dipole with zonal flow over Europe (Fig. 11A). This NAO-like 791 pattern is characterized by high (low) SLP anomalies over the Icelandic Low and low 792 (high) SLP anomalies over the Azores High during its negative (positive) phase 793 (Luterbacher et al., 2002; Josey et al., 2011), although its southern centre is offset to the 794 east (Fig. 1A, 11A). The second EOF mode (EOF2) shows anomalously low (high) SLP

over the North Sea that may be associated with the negative (positive) phase of the East
Atlantic/West Russian (EA/WR) pattern (Fig. 11B) (Josey et al., 2011). The third EOF
mode (EOF3) is characterized by high (low) SLP anomalies centred west of the British
Isles around 55°N latitude and 25°W longitude (Fig. 11C), which correspond to the
negative (positive) phase of the East Atlantic (EA) pattern (Luterbacher et al., 2002;
Josey et al., 2011).

801 The EA/WR and EA patterns were mentioned as potential drivers of atmospheric 802 circulation and climate change in western Europe during the late Holocene (Dermody et 803 al., 2012; Moffa-Sanchez et al., 2014; Sanchez-Lopez et al., 2016; Sicre et al., 2016; 804 Degeai et al., 2017). From observational data, the negative phases of the EA/WR and EA 805 patterns were associated with higher precipitation and enhanced ocean heat loss in the 806 northwestern Mediterranean (Qian et al., 2000; Josey et al., 2011), while strong (weak) 807 surface winds in France were associated with either a negative (positive) phase of the 808 EA/WR pattern or a positive (negative) phase of the EA pattern (Najac et al., 2009). 809 Hence, given the late Holocene storm activity in France and the SST record in the 810 northwestern Mediterranean (Fig. 10h-k), the late IACP should have been characterized 811 by a negative phase of the EA pattern, then the lower RWP (upper RWP-DACP) and 812 MWP (LIA) would have been associated with a positive (negative) phase of the EA/WR 813 pattern (Fig. 10l).

In the northwestern Mediterranean, the coupling of cold northerly winds generated by
winter North Atlantic blocking events due to anomalously high SLP in the eastern North
Atlantic during the negative phase of the EA pattern (Shabbar et al., 2001; Jiang et al.,
2003; Trigo et al., 2004; Josey et al., 2011; Moffa-Sanchez et al., 2014; Sanchez-Lopez et
al., 2016; Sicre et al., 2016) with warm and dry summer due to the northward position of
the ITCZ (Goudeau et al., 2015) could have led to enhanced convection and higher

rainfall along the southern Alps in the late IACP (Fig. 10l-m, 11C), similarly to modern
heavy precipitating systems over southeastern France (Nuissier et al., 2008; Duffourg
and Ducrocq, 2011). This synoptic situation could explain the high fluvial activity and
sediment flux in the ARDS during the late IACP (Fig. 10e).

824 The positive phase of the EA/WR pattern in the lower RWP from ca. 2300 to 1800 cal yr 825 BP was associated with a large-scale circulation pattern characterized by a negative or 826 positive phase of the NAO (Fig. 10g, l). This period shows a decreasing trend in fluvial 827 activity in the ARDS apart from the 2nd and 1st centuries BCE (Fig. 10e). The ITCZ was 828 still in a northern position during the lower RWP, but the absence or low frequency of 829 atmospheric blocking events and cold northerly winds would have limited atmospheric 830 convection and precipitation in the southern Alps at this time. During the late RWP and 831 the DACP from 1800 to 1150 cal yr BP, the negative phase of the EA/WR pattern 832 corresponded to a general increase in fluvial activity in the ARDS, with higher than 833 average values when the negative phases of the EA/WR and NAO patterns were coupled 834 in the late DACP between 1600 and 1150 cal yr BP (Fig. 10e, g, l, 11B). 835 In the Western Mediterranean, there is a broad consensus on the climatic conditions of 836 the MWP (LIA), which were generally warm (cold) and dry (wet) from continental 837 records in Spain (Martin-Puertas et al., 2008, 2010; Martin-Chivelet et al., 2011; Moreno 838 et al., 2012; Sanchez-Lopez et al., 2016), Italy (Frisia et al., 2005; Mensing et al., 2016; 839 Sadori et al., 2016), southern France (Degeai et al., 2017) and northern Morocco 840 (Wassenburg et al., 2013). Overall, the MWP (LIA) would have been characterized by the 841 combination of a positive (negative) phase of the NAO with a positive (negative) phase 842 of the EA/WR pattern (Fig. 10g, l). 843 The climatic context of the LIA would have been similar to that of the DACP (Fig. 11B).

844 During the cold and humid period of the LIA, high detrital inputs of terrigenous material

845 were identified in marine sediments from the western Mediterranean basin (Fig. 1A).

846 Interestingly, the sharp increase in subaerial sediment flux in the ARDS after 500 cal yr

847 BP is consistent with this high offshore discharge of riverine sediment (Fig. 10d).

848 Overall, the LIA was characterized by high fluvial sediment supply to Mediterranean

deltas in an increasingly human-influenced environment (Anthony et al., 2014).

850 However, it is currently difficult to assess the real contribution of human land use to late

851 Holocene sediment flux in the ARDS because of the lack of quantitative data to measure

the magnitude of the geomorphic impact of anthropogenic forcing.

Finally, the climate-driven fluvial activity in the ARDS over the past 2500 years would have been higher than average when the negative phase of the NAO-like pattern was associated with either a negative phase of the EA/WR pattern or the combination of a negative EA pattern with a northward migration of the ITCZ.

857

858

859 8. CONCLUSION

860 The elevation modelling and palaeogeographic reconstruction of incised valleys in 861 coastal areas is challenging because this kind of environment is often characterized by 862 rapid changes in relative sea level, shoreline progradation, delta sedimentation and 863 climatic conditions. A new multi-method approach using cores and archaeological 864 trenches, geochemistry and magnetic susceptibility of sediments, malacology, 865 radiocarbon chronology, geostatistics and numerical modelling was undertaken in order 866 to robustly characterize changes in environmental conditions, geomorphic setting and 867 sediment flux associated with delta growth in incised-valley systems. This quantitative 868 method was applied to the bayhead delta of the Argens River in the western
869 Mediterranean, in order to assess the impacts of autogenic control and climate change 870 on the variations in sediment flux and fluvial activity of this deltaic system. 871 The coastal dynamics of the Argens River delta system were studied over the last 2500 872 years. The sedimentary successions of this deltaic depositional system revealed four 873 major facies associations from bottom to top: (1) a prodelta with clay-sand alternations 874 deposited in a marine environment with brackish or freshwater influences and high 875 levels of salinity, (2) a delta front characterized by higher sand content as well as by 876 coarsening-upward trend and gradual increase in riverine input, (3) river channels filled 877 with sand and gravel, and (4) a floodplain with a high abundance of land snails and low 878 salinity levels. 879 Late Holocene sediment fluxes were estimated from numerical modelling then 880 compared with hydroclimate proxy records and geomorphic factors. An overall decrease 881 in subaqueous sediment supply over the last 2500 years could have been 882 geomorphologically controlled by autogenic processes such as changes in 883 accommodation space and delta slope. High rates of sediment flux between 2500 and 884 2000 cal yr BP coincided with higher precipitation and more frequent riverine flooding 885 in the western Mediterranean. The maximum rate of subaerial sediment flux from 500 to 886 0 cal yr BP could have been caused by higher fluvial activity and more frequent flood 887 events during the cold and wet climatic conditions of the Little Ice Age. 888 In general, the high sediment fluxes at 2500-2000 and 500-0 cal yr BP were associated 889 with higher than average levels of fluvial activity in the Argens River delta system. The 890 periods of high fluvial activity driven by climate change were associated with a 891 southward shift of the North Atlantic westerlies and storm tracks in southern Europe. 892 This atmospheric circulation pattern was combined with either (1) a negative phase of 893 the East Atlantic pattern coupled to a northward migration of the intertropical

894 convergence zone in the late Iron Age Cold Period, or (2) a negative phase of the East 895 Atlantic/West Russian pattern during the Dark Ages Cold Period and Little Ice Age. 896 This quantitative approach will need to be applied to other Mediterranean deltas in 897 order to compare the variation of coastal sediment fluxes in response to hydroclimate 898 change and geomorphic setting over the past millennia. Besides, the study of other 899 deltaic sequences will be needed to firmly establish a link between delta sedimentation 900 and atmospheric circulation patterns in the western Mediterranean. Furthermore, the 901 analysis of palaeobotanical, archaeological and historical data will be required to 902 accurately estimate the impact of human land use on sediment flux and geomorphic 903 change in these coastal areas.

904

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915 Data availability

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1326 Legend of tables and figures

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1328

Table 1. Radiocarbon dates from the Argens River delta system. Elevation in reference to
LiDAR data (Litto3D, IGN/SHOM). LPT1: La Plaine T1, LPT3: La Plaine T3, TAT2: Théâtre
d'agglomération T2. *: age inversion.

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Table 2. Nominal relative sea level (RSL) with minimum and maximum errors from
Vacchi et al. (2016). The RSL prediction was computed using the open source code
SELEN (Spada and Stocchi, 2007) that employs the ICE-5G (VM2) model from Peltier
(2004).

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1340Fig. 1. Location of the Argens River in the northwestern Mediterranean (A) and

1341 topographic and geological setting of the Argens River delta system (B). Coordinate

1342 system: WGS 1984 World Mercator (A) and RGF 1993 Lambert-93 (B). ITCZ:

1343 intertropical convergence zone. 1, 2: Durance River sites (Arnaud-Fassetta et al., 2010;

1344 Benito et al., 2015a). Marine records of terrigenous material in sediments from the

1345 western Mediterranean basin: KSGC-31 site (Jalali et al., 2016); MD99-2343 site (Frigola

1346 et al., 2007); 305G, 306G, 384B and 436B sites (NietoMoreno et al., 2011, 2013a, 2013b);

1347 DP30 site (Goudeau et al., 2014, 2015). References of published cores and archaeological

1348 surveys: Dubar (2004) for Verteil core, Allinne et al. (2006) for ESC1 core, Devillers and

1349 Bonnet (2006) for VIL2 core, Excoffon et al. (2006) and Devillers et al. (2007) for

1350 Théâtre d'agglomération T2 archaeological survey, Excoffon et al. (2010) for Villa

1351	Romana archaeological survey, Bony et al. (2011) for FIX, FXI and SC6 cores, Bertoncello
1352	et al. (2014a) for PL1 and PAL1 cores, Bertoncello et al. (2016) for La Plaine T1-T3
1353	archaeological survey.
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1356	Fig. 2. Stratigraphy, magnetic susceptibility, geochemistry and malacofauna of BN1 core
1357	in the Argens River delta system. Vertical dashed lines correspond to the mean values of
1358	variables.
1359	
1360	
1361	Fig. 3. Stratigraphy, magnetic susceptibility, geochemistry and malacofauna of BN5 core
1362	in the Argens River delta system. Vertical dashed lines correspond to the mean values of
1363	variables.
1364	
1365	
1366	Fig. 4. Malacological diagram of BN1 core. Main environmental conditions inferred from
1367	a cluster analysis by constrained incremental sum of squares (CONISS) with a
1368	dissimilarity coefficient corresponding to Edwards and Cavalli-Sforza's chord distance
1369	(Grimm, 1987).
1370	
1371	
1372	Fig. 5. Malacological diagram of BN5 core. Main environmental conditions inferred from
1373	a cluster analysis by constrained incremental sum of squares (CONISS) with a
1374	dissimilarity coefficient corresponding to Edwards and Cavalli-Sforza's chord distance
1375	(Grimm, 1987).

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1378 Fig. 6. 2D age-elevation models in the coastal plain of the Argens River delta system. The 1379 data used in the age-elevation models correspond to the median probability of 1380 calibrated ages (white dot) from the compilation of radiocarbon ages available in the 1381 Argens River delta system (Table 1) and to modern elevation based on LiDAR data 1382 (Litto3D, IGN/SHOM) for the year 2015 (blue triangle). Numbers in brackets represent 1383 the 2- σ uncertainty interval of calibrated ages (Table 1). The radiocarbon ages in red 1384 were discarded to avoid age reversals (see explanation in text). References for 1385 radiocarbon ages, lithology and depositional environment (if available): Dubar (2004) 1386 for Verteil core, Allinne et al. (2006) for ESC1 core, Devillers and Bonnet (2006) for VIL2 1387 core, Devillers et al. (2007) for Théâtre d'agglomération T2 archaeological survey, 1388 Excoffon et al. (2010) for Villa Romana archaeological survey, Bertoncello et al. (2014a) 1389 for PL1 and PAL1 cores, Bertoncello et al. (2016) for La Plaine T1 and T3 archaeological 1390 surveys, and this study for BN1 and BN5 cores. The data point for Villa Romana was 1391 estimated from the reconstruction of the shoreline around 2000 cal yr BP based on 1392 archaeological data, radiocarbon ages and relative sea level reconstruction (Excoffon et 1393 al., 2010; Vacchi et al., 2016). Location of age-elevation models A1-3 in Fig. 1B. Models 1394 interpolated with a thin plate spline kernel function and an 8-sector neighbourhood 1395 search algorithm. The isochrones of the unconfined bayhead delta were obtained by 1396 shifting the LiDAR data along the BN5 age-elevation model. (For interpretation of the 1397 references to colour in this figure legend, the reader is referred to the web version of 1398 this article.)

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Fig. 7. Interpolated maps of sediment thickness in the Argens River delta system over
the past 2500 years. LPT1 and LPT3: La Plaine T1 and T3 archaeological surveys, TAT2:
Théâtre d'agglomération T2 archaeological survey, VR: Villa Romana archaeological
survey. Models interpolated with a multiquadratic kernel function and an 8-sector
neighbourhood search algorithm.

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1408 Fig. 8. Palaeoelevation in the Argens River delta system over the past 2500 years. LPT1 and LPT3: La Plaine T1 and T3 archaeological surveys, TAT2: Théâtre d'agglomération 1409 1410 T2 archaeological survey, VR: Villa Romana archaeological survey. The past sea levels 1411 used in the elevation models are based on the relative sea level (RSL) in the western 1412 Ligurian Sea determined by the ICE-5G (VM2) geophysical model (Peltier, 2004) for the 1413 last 2500 years (Table 2). In this area, a good fit between the model and available 1414 geological sea level proxies was observed along the whole Holocene (Vacchi et al., 2016). 1415 The reconstructed RSL at 2000 cal yr BP is consistent with the Roman sea level deduced 1416 from marine encrustation found on archaeological harbour structures, a very robust 1417 methodology to infer the RSL position in the Mediterranean Sea (Morhange and 1418 Marriner, 2015; Vacchi et al., 2016). The RSL in the Forum Iulii harbour (Fig. 1B) was 1419 estimated at -40 ± 10 cm (Morhange et al., 2013) or -33 ± 6 cm (Devillers et al., 2007) 1420 between the first century BCE and the first century CE. 1421 1422

Fig. 9. Evolution of sediment flux and fluvial activity in the Argens River delta system
(ARDS) and comparison with riverine flooding frequency in the southern French Alps
over the past 2500 years. (a) Sediment fluxes calculated for each 500-year interval by

1426 estimating the volume of sediment using the ArcGIS 10 Surface Volume tool with either 1427 (1) the elevation models in Fig. 8 and reference planes corresponding to the mean 1428 relative sea level from Vacchi et al. (2016) for each 500-year interval for the confined 1429 bayhead delta, or (2) the LiDAR data shifted along the BN5 age-elevation model (Fig. 6, 1430 age profile A1) and a reference plane at 20 m below present sea level for the unconfined 1431 bayhead delta. (b) Proxy records of riverine input and fluvial activity corresponding to 1432 the average of the standard scores of Si/Al, K/Al and Rb/Al ratios from BN1 core with an 1433 age scale established by using AnalySeries 2.0 (Paillard et al., 1996). (c) Proxy records of 1434 riverine input and fluvial activity corresponding to the average of the standard scores of 1435 Si/Al, K/Al and Rb/Al ratios from BN5 core with an age scale established by using 1436 AnalySeries 2.0 (Paillard et al., 1996). (d) Periods of relatively high flood frequency of 1437 the lower Durance River and its tributaries (site 1 in Fig. 1A) in the southern French 1438 Alps (Arnaud-Fassetta et al., 2010; Benito et al., 2015a). (e) Periods of relatively high 1439 flood frequency of the middle Durance River and its tributaries (site 2 in Fig. 1A) in the 1440 southern French Alps (Arnaud-Fassetta et al., 2010; Benito et al., 2015a). (f) Relative 1441 cumulative probability density function of radiocarbon dates from flood and extreme 1442 fluvial event units of rivers in southeastern France (Fig. 1A) (Benito et al., 2015a). IACP: 1443 Iron Age Cold Period, RWP: Roman Warm Period, DACP: Dark Ages Cold Period, MWP: 1444 Medieval Warm Period, LIA: Little Ice Age, GW: Global Warming.

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Fig. 10. Comparison of sediment flux and fluvial activity in the Argens River delta system
(ARDS) with geomorphic factors and climate proxy records over the last 2500 years. (a)
Mean slope of the ARDS longitudinal gradient between Verteil and BN5 from age profiles
in Fig. 6. (b) Volume of accommodation space of the ARDS estimated using the ArcGIS 10

1451 Surface Volume tool with the elevation models in Fig. 8 and reference planes 1452 corresponding to the relative sea level from Vacchi et al. (2016). (c) Subaqueous 1453 sediment flux in the ARDS with an error range of \pm 20%. (d) Subaerial sediment flux in 1454 the ARDS with an error range of \pm 20%. (e) Fluvial activity in the easternmost part of the 1455 lower Argens valley corresponding to the average of the standard scores of fluvial 1456 activity at BN1 and BN5 (Fig. 9b-c) resampled at a 5-years interval by linear 1457 interpolation using AnalySeries 2.0 (Paillard et al., 1996), along with period of autogenic 1458 control (see section 7.2). (f) Mean band width of three radiometrically dated 1459 speleothems from the Uamh an Tartair cave in northwestern Scotland (Fig. 1A) (Proctor et al., 2002). (g) Potential leading mode of atmospheric circulation in Europe (see 1460 1461 section 7.3). (h) Alkenone-based sea surface temperature (SST) in the northwestern 1462 Mediterranean with cold relapse episodes (CR4-6) from the KSCG-31 core in the Gulf of 1463 Lions (Fig. 1A) (Jalali et al., 2016). (i) Storm activity in southeastern France from the 1464 sedimentary record of the Bagnas lagoon (Fig. 1A) (Degeai et al., 2015). (j) Storm 1465 activity in southeastern France from the sedimentary record of the Pierre Blanche 1466 lagoon (Fig. 1A) (Sabatier et al., 2012). (k) Storm activity from sedimentary records in 1467 the Seine River estuary in northwestern France (Fig. 1A) (Sorrel et al., 2009). (l) 1468 Potential second or third mode of atmospheric circulation in Europe (see section 7.3). 1469 (m) 5-yr averaged oxygen isotope record of authigenic carbonate from Lake Bosumtwi 1470 in Ghana (Fig. 1A) (Shanahan et al., 2009). NAO: North Atlantic oscillation, EA: East 1471 Atlantic pattern, EA/WR: East Atlantic/West Russian pattern, IACP: Iron Age Cold 1472 Period, RWP: Roman Warm Period, DACP: Dark Ages Cold Period, MWP: Medieval Warm 1473 Period, LIA: Little Ice Age, GW: Global Warming. 1474

- 1476 Fig. 11. Late Holocene atmospheric circulation patterns and climatic conditions in the
- 1477 North Atlantic and Europe during the periods of high fluvial activity driven by climate
- 1478 change in the Argens lower valley. Empirical orthogonal functions (EOF1-3) calculated
- 1479 from sea level pressure reconstructed by Luterbacher et al. (2002) for the period 1500-
- 1480 1999 CE. 1: Argens River delta system, 2: Gulf of Lions, 3: Bagnas and Pierre Blanche
- 1481 lagoons, 4: Seine River estuary, SST: sea surface temperature, NAO: North Atlantic
- 1482 oscillation, EA: East Atlantic pattern, EA/WR: East Atlantic/West Russian pattern, ITCZ:
- 1483 intertropical convergence zone, IACP: Iron Age Cold Period, RWP: Roman Warm Period,
- 1484 DACP: Dark Ages Cold Period, LIA: Little Ice Age.
- 1485

Table 1

Core or trench	Elevation (m)	Laboratory code	Material	¹⁴ C age (BP)	Reference	2-σ interval of calibrated age (cal yr BP)	Median probability of calibrated age (cal yr BP)
BN1	0.30	Poz-81546	Charcoal	375 ± 30	This study	504-318	442
BN1	-3.60	Poz-81547	Charcoal	2520 ± 30	This study	2744-2492	2595*
BN1	-5.60	Poz-81608	Charcoal	2215 ± 30	This study	2322-2151	2231*
BN1	-7.00	Poz-81548	Charcoal	3330 ± 35	This study	3678-3467	3565*
BN1	-8.60	Poz-81549	Charcoal	2085 ± 30	This study	2143-1990	2058
BN1	-10.10	Poz-81550	Charcoal	2480 ± 30	This study	2723-2385	2585
BN5	-7.86	Lyon-13936	Wood	2055 ± 30	This study	2118-1934	2023*
BN5	-9.98	Lyon-13937	Wood	1975 ± 30	This study	1993-1871	1925
BN5	-12.33	Lyon-13938	Wood	2175 ± 30	This study	2310-2072	2231
BN5	-13.67	Lyon-13939	Wood	2195 ± 30	This study	2313-2133	2237
ESC1	0.18	Poz-12168	Seed	680 ± 30	Allinne et al. (2006)	680-561	652
ESC1	-1.57	Poz-12167	Seed	2535 ± 30	Allinne et al. (2006)	2747-2495	2625
ESC1	-2.52	Poz-12166	Seed	3735 ± 35	Allinne et al. (2006)	4227-3980	4089*
ESC1	-4.82	Poz-9593	Charcoal	3535 ± 35	Allinne et al. (2006)	3904-3702	3816
LPT1	3.44	Poz-78590	Charcoal	155 ± 30	Bertoncello et al. (2016)	284-0	174
LPT1	3.14	Poz-78591	Charcoal	175 ± 30	Bertoncello et al. (2016)	293-0	180
LPT1	2.44	Poz-78589	Charcoal	1315 ± 30	Bertoncello et al. (2016)	1295-1183	1259
LPT3	3.05	Poz-78593	Charcoal	4125 ± 30	Bertoncello et al. (2016)	4817-4531	4667*
LPT3	2.45	Poz-80846	Charcoal	960 ± 50	Bertoncello et al. (2016)	960-764	859
PAL1	-0.74	Lyon-5430	Seed	1075 ± 35	Bertoncello et al. (2014a)	1057-930	982
PAL1	-1.84	Lyon-5431	Seed	1285 ± 35	Bertoncello et al. (2014a)	1294-1097	1232*

PAL1	-2.34	Lyon-5432	Seed	1285 ±	Bertoncello	1294-1097	1232
				35	et al.		
	0.04	L	C l	2145	(2014a)	2204 2000	2127
PALI	-8.84	Lyon-5061	Seed	2145 ± 30	et al	2304-2009	2137
				30	(2014a)		
PAL1	-14.34	Lvon-5062	Seed	2210 +	Bertoncello	2320-2149	2232
	1 110 1		beeu	30	et al.	2020 2117	2202
					(2014a)		
PAL1	-17.34	Lyon-5433	Seed	2800 ±	Bertoncello	2998-2791	2904
				40	et al.		
					(2014a)		
PL1	1.49	Lyon-4223	Charcoal	1540 ±	Bertoncello	1525-1356	1448
				35	et al.		
DI 1	1.21	1 4001	C I	2515.	(2014a)	2746 2467	2502*
PLI	-1.21	Lyon-4221	Seed	2515 ±	Bertoncello	2/46-246/	2593*
				40	(2014_{2})		
DI 1	-1 76	Lvon-4228	Seed	2510+	Bertoncello	2740-2490	2590*
1 1 1	1.70	Lyon 1220	beeu	30	et al	27 10 2190	2390
				00	(2014a)		
PL1	-3.41	Lyon-4226	Charcoal	3120 ±	Bertoncello	3442-3234	3340*
		·		35	et al.		
					(2014a)		
PL1	-3.56	Lyon-5434	Seed	2395 ±	Bertoncello	2687-2345	2425
				35	et al.		
					(2014a)		
PL1	-4.31	Lyon-4225	Charcoal	2450 ±	Bertoncello	2704-2361	2528
				30	et al.		
DI 1	6.76	Luon 4224	Sood	2400 +	(2014a) Portoncollo	2720 2201	2502
PLI	-0.70	Ly011-4224	Seeu	2490 ±	ot al	2/30-2301	2505
				10	(2014a)		
PL1	-7.51	Lyon-5435	Charcoal	4200 ±	Bertoncello	4851-4584	4730*
		5		45	et al.		
					(2014a)		
PL1	-8.81	Lyon-4222	Charcoal	2820 ±	Bertoncello	3057-2846	2924*
				35	et al.		
	10.00	1. 1910		0.64.0	(2014a)	0040.0540	0.5.40
PL1	-10.06	Lyon-4219	Charcoal	2610 ±	Bertoncello	2843-2540	2749
				40	(2014_{2})		
ΤΔΤ2	-0.33	Poz-14372	Marine	2345 +	Devillers et	2087-1888	1983
11112	0.55	102 11372	shell	30	al. (2007)	2007 1000	1905
Verteil	-2.41	Ly 5889	Peat	3050 ±	Dubar et al.	3443-3008	3247
		5		75	(2004)		
VIL2	-2.25	Poz-10916	Charcoal	915 ±	Devillers	920-762	848
				30	and Bonnet		
					(2006)		
VIL2	-2.90	Poz-10917	Charcoal	1340 ±	Devillers	1305-1185	1280
				30	and Bonnet		
		D- 10010	Ch 1	222F -	(2006) Dest'll	2426 2220	2245
VILZ	-5.65	P0Z-10918	Unarcoal	2325 ±	Devillers	2426-2208	2345
				50	(2006)		
VII.2	-7 75	P07-10922	Marine	5190 +	Devillers	5634-5478	5571
¥1114	1.1.5	102 10722	shell	30	and Bonnet	5051 5770	0071
					(2006)		

1490 Table 2

Age (cal yr	Minimum	Nominal	Maximum
BP)	RSL (m)	RSL (m)	RSL (m)
0	0.00	0.00	0.00
500	-0.14	-0.07	0.06
1000	-0.28	-0.13	0.12
1500	-0.43	-0.20	0.19
2000	-0.58	-0.28	0.26
2500	-0.78	-0.38	0.32





1497 Figure 2 (two-column) BN1 core Magne



Figure 3 (two-column) BN5 core

1501 Figure 4 (two-column)






1505 Figure 6 (two-column)















1515 Figure 11 (single column) A. EOF1 (51%) NAO-like pattern in its negative phase





