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Timing and mechanisms of sediment accumulation and pedogenesis: Insights from the Po Plain (northern Italy)

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Abstract:	Paleosol stratigraphy in the Po Plain, reconstructed through sedimentological analysis and correlation of ca. 170 core data chronologically constrained by 376 radiocarbon dates, revealed strict relationships between pedogenetic and fluvial processes in alluvial and paralic environments. Vertically stacked, weakly developed paleosols within Upper Pleistocene and Holocene mud-prone strata testify to intermittent bedogenesis, periodically interrupted by overbank sedimentation. Individual paleosols are laterally traceable for tens of km and exhibit A-Bk-Bw, A-Bk or A-Bw profiles. Stratigraphically ordered 14C calibrated ages from A horizons testify to slow aggradation during 4-6 thousand years-long exposure periods. Burial ages, with an error of few centuries, are provided by plant debris at the top of A horizons. Millennial- scale climate oscillations and glacio-eustasy are the main drivers of the pedo- sedimentary evolution of the area during the last 50 kyr. Pleistocene (P) paleosols developed in well-drained floodplain environments, dissected by fluvial incision, during relatively warm periods. Overbank sedimentation and paleosol burial occurred during colder phases. High-sediment-supply during the Last Glacial Maximum hindered bedogenesis and led to the accumulation of 3-10 m-thick overbank strata. Widespread soil development (paleosol PH) occurred at the end of Last Glacial Maximum, following glaciers retreat and afforestation of drainage basins. At distal locations, paleosol PH was progressively buried under estuarine sediments during the Holocene phases of post-glacial sea-level rise. Beyond the area of marine influence, burial ages of paleosol PH change from a place to another without specific spatial trends and reflect upstream luvial sedimentation dominated by avulsions and deposition of spatially restricted alluvial units. Holocene (H) paleosols show a poor correlation potential and laterally variable degree of maturity that reflect avulsive sedimentation patterns and crevassing. Deforestation and land
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Dear Editor,

please find enclosed the research paper "**Timing and mechanisms of sediment accumulation and pedogenesis. Insights from the Po Plain (Northern Italy)**" by Bruno L., Campo B., Hajdas I., Hong W., Amorosi A. that we would like to submit for publication in your journal *Palaeo 3*.

The manuscript has been prepared in accordance with the authors guidelines.

Best regards,

Dr. Luigi Bruno

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1	Highlights
2	
3	Po Plain pedostratigraphy reveals relations between pedogenetic and alluvial events
4	
5	Upper Pleistocene-Holocene stacked paleosols testify to intermittent pedogenesis
6	
7	Radiocarbon dates permitted to estimate the duration of the main pedogenetic phases
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9	Climate change and eustasy controlled Late Pleistocene pedo-sedimentary evolution
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11	Discontinuous Holocene paleosols reflect avulsive sedimentation and crevassing

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26 Abstract

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Paleosol stratigraphy in the Po Plain, reconstructed through sedimentological analysis and 28 correlation of ca. 170 core data chronologically constrained by 376 radiocarbon dates, revealed 29 30 strict relationships between pedogenetic and fluvial processes in alluvial and paralic environments. 31 Vertically stacked, weakly developed paleosols within Upper Pleistocene and Holocene mud-prone 32 strata testify to intermittent pedogenesis, periodically interrupted by overbank sedimentation. 33 Individual paleosols are laterally traceable for tens of km and exhibit A-Bk-Bw, A-Bk or A-Bw profiles. Stratigraphically ordered ¹⁴C calibrated ages from A horizons testify to slow aggradation during 4-6 34 thousand years-long exposure periods. Burial ages, with an error of few centuries, are provided by 35 36 plant debris at the top of A horizons.

Millennial-scale climate oscillations and glacio-eustasy are the main drivers of the pedo-37 sedimentary evolution of the area during the last 50 kyr. Pleistocene (P) paleosols developed in well-38 drained floodplain environments, dissected by fluvial incision, during relatively warm periods. 39 40 Overbank sedimentation and paleosol burial occurred during colder phases. High-sediment-supply 41 during the Last Glacial Maximum hindered pedogenesis and led to the accumulation of 3-10 m-thick 42 overbank strata. Widespread soil development (paleosol PH) occurred at the end of Last Glacial Maximum, following glaciers retreat and afforestation of drainage basins. At distal locations, 43 44 paleosol PH was progressively buried under estuarine sediments during the Holocene phases of 45 post-glacial sea-level rise. Beyond the area of marine influence, burial ages of paleosol PH change from a place to another without specific spatial trends and reflect upstream fluvial sedimentation 46 47 dominated by avulsions and deposition of spatially restricted alluvial units. Holocene (H) paleosols 48 show a poor correlation potential and laterally variable degree of maturity that reflect avulsive

49	sedimentation	patterns	and	crevassing.	Deforestation	and	land	reclamation	affected	pedo-
50	sedimentary pr	ocesses d	uring	the last 6 ky	r.					

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53 Keywords

1. INTRO

Paleosol stratigraphy, pedo-sedimentary evolution, radiocarbon dates, climate change, eustasy,
Late Quaternary

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In alluvial settings, the development and burial of soil horizons are intimately related to fluvial 60 activity (Kraus and Bown, 1993) and governed by a variety of factors. High sediment supply and 61 fluvial aggradation inhibit pedogenesis (Demko et al., 2004), whereas low sediment supply and 62 63 fluvial incision promote soil development in interfluvial areas (Blum and Törngvist, 2000). Climate influences the amount of sediment delivered to the basins, modulating the magnitude and type of 64 65 precipitations, the vegetation cover of drainage areas and the river's transport capacity (Vandenberghe, 2003). Alluvial sedimentation can be locally hindered or enhanced by tectonic 66 67 movements (Holbrook and Schumm, 1999; Viseras et al., 2003; Benvenuti et al., 2008; Pati et al., 2019), differential subsidence (Cohen et al., 2003) or by avulsions and other processes related to 68 the intrinsic dynamics of the fluvial apparatus (Kraus, 1999; Cleveland et al., 2007; Hajek and 69 70 Wolinsky, 2012). In coastal settings, fluvial dynamics are strongly influenced by sea-level oscillations 71 (Atchley et al., 2004; Blum et al., 2013). Incised valleys (and associated interfluve paleosols) related

to glacio-eustasy are a widely recognized trait in the sedimentary record (Shanley and McCabe,
1994; Aitken and Flint, 1996; McCarthy and Plint, 1998; Plint et al., 2001; Gibling et al., 2011).

The causal relationships between paleosol formation and alluvial sedimentation have been 74 reconstructed in ancient alluvial successions, where the geometric relationships between paleosols 75 and associated fluvial sediment bodies are superbly exposed (Gibling and Bird, 1994; McCarthy and 76 77 Plint, 2003; Dubiel and Hasiotis, 2011; Soares et al., 2020). However, the role of distinct controlling 78 factors on paleosol development and burial can hardly be assessed (Wright and Marriott, 1993; 79 Atchley et al., 2004), because of the temporal resolution of pre-Quaternary dating methods, insufficient to describe pedo-sedimentary processes which typically act at millennial time scales 80 (Rasbury et al., 1997; Hofmann et al., 2017). Furthermore, diagenesis may have substantially 81 82 modified soil properties after burial, hindering comparison with modern soils and the reconstruction 83 of the soil-forming environment (Trendell et al., 2012).

Improved techniques of subsurface investigation have recently prompted the study of buried upper Quaternary alluvial successions (Choi, 2005; Tsatskin et al., 2015; Amorosi et al., 2017a,b). In this relatively narrow time window, drainage areas, ecosystems and soil-forming environment did not change significantly. Moreover, the high temporal resolution of Quaternary dating methods permits to link unequivocally pedogenetic processes to a well-known climate and eustatic history. Radiocarbon is probably the most reliable dating method for the last 50 kyr (Reimer et al., 2020).

carbon is continuously incorporated in soils, and subjected to complex, and not totally understood, biological and physical transformations (Larionova et al., 2017; Meingast et al., 2020). These processes can act at different velocity and magnitude, depending on climate, morphology of the exposed surface, parent material, and vegetation (Alexandrovskiy and Chichagova, 1998; Kleber,

95 2010).

This study is based on refined pedostratigraphic reconstructions and on a vast ¹⁴C dataset which 96 permitted to chronologically constrain the major architectural and facies changes. The primary aim 97 of this research is to understand the causes and mechanisms of soil development and burial in an 98 alluvial-deltaic setting and its relation to fluvial and coastal dynamics. To this purpose, we focused 99 on Upper Pleistocene-Holocene paleosols and on their stratigraphic relationship with older, coeval 100 101 and younger pedo-sedimentary units. Secondary objectives are to discuss: (i) the relative role of 102 global and local controlling factors on soil development and burial; (ii) the potential use (and major 103 limitations) of radiocarbon to assess the exposure age of paleosols. 104 105 2. GEOLOGICAL SETTING 106 107 2.1. Structural setting and basin-scale stratigraphy 108 109 The Po sedimentary basin is enclosed in a natural amphitheatre bordered by the Alps, to west 110 and to the north, and by the Apennines, to the south. The eastward prolongation of the Po Plain is 111 112 the Adriatic epicontinental sea (Fig. 1). The most external thrusts of the Southern Alps and of Apennines, showing opposite vergence, locally face each other beneath the Po Plain (Fantoni et al., 113 114 2004). South-Alpine buried thrusts are arranged in a single wide arc running from Milan to the Garda Lake (Vannoli et al., 2015). The Apennine front is composed of three tectonically active arcuate 115 thrust systems: from west to east, the Monferrato, Emilia and Ferrara-Romagna Arcs (Ghielmi et al., 116 117 2013). Thrusts of the Emilia and Ferrara Arcs are sealed beneath a pile of sediments with extremely variable thicknesses (from a few hundred metres to ca. 8 km; Amadori et al., 2019). These lateral 118 119 variations in thickness are controlled by lateral changes in subsidence rates (Campo et al., 2020),

ranging between zero and 2 mm/ 2 mm/ 2 sruno et al., 2020a). The northern margin of the Apennines,
which corresponds to an out-of-sequence system of N-NE verging thrusts (Maestrelli et al., 2018),
has been involved in a generalized uplift since the Early Pleistocene (Argnani et al., 2003).

The Po Basin fill shows and overall regressive trend from deep-marine to continental and coastal 123 units (Ghielmi et al., 2013). Widespread alluvial sedimentation started in the Po Plain around 870 124 125 kyr B.P. (Muttoni et al., 2003, 2011; Gunderson et al., 2014). The Po River, flowing in W-E direction, 126 divides the alluvial plain in two sectors (Fig. 1A). In the norther sector, Alpine tributaries flow 127 entrenched in Pleistocene sandy units (Ravazzi et al., 2012; Fontana et al., 2014; Bruno et al., 2018). To the south, paleosol-bearing Pleistocene muds are sealed by a thick (up to 30 m) Holocene alluvial 128 and paralic succession (Amorosi et al., 2017a, b). Early Holocene estuarine deposits typically onlap 129 130 onto Upper Pleistocene alluvial strata, testifying to the progressive drowning of the alluvial plain 131 after 11.5 cal kyr B.P. (Bruno et al., 2017). Following stabilization of sea-level, the Middle-Late Holocene records the progressive filling of the Po estuary and the successive delta progradation 132 (Amorosi et al., 2017b, 2019). 133

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135 2.2. Paleosols of the Po Basin

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Highly weathered rubified soils have been reported from the Po Basin margin (Cremaschi, 1987; Garzanti et al., 2011) and from isolated hills in the alluvial plain, generated by South-Alpine (Zerboni et al., 2015) and Apennine blind thrusts (Zuffetti et al., 2018). In depocentral areas, mature paleosols are replaced by sequences of aggradationally stacked, weakly developed paleosols laterally traceable for tens of km (Amorosi et al., 2017a). A series of closely spaced paleosols, marking the MIS 3/MIS 2 transition, has been described as the equivalent of the Interfluve Sequence Boundary (McCarty and Plint, 2003) in a high accommodation setting, due to its peculiar geometric 144 relationships with genetically related fluvial-channel bodies (Amorosi et al., 2017a). The most widespread paleosol reported in the Po Plain marks the Pleistocene-Holocene boundary and 145 responds to several local names: 'Caranto' in the Venetian Plain (Mozzi et al., 2003), 'Ringhiera 146 paleosol' in the Romagna area (Ravazzi et al., 2006), 'YD paleosol' in the Po Delta Plain (Amorosi et 147 148 al., 2017b; Morelli et al., 2017), 'PH paleosol' in the Bologna urban area (Amorosi et al., 2014; Bruno 149 et al., 2020b; this name will be used in this paper). Although a wide literature reported the existence 150 of this pedogenized horizon, the mechanisms and causes of its formation and burial are not yet fully understood. More discontinuous paleosols have been described from the Holocene succession 151 (Bruno et al., 2015, 2013). 152

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155 **3. METHODS**

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This work is based on the sedimentological and stratigraphic analysis of core material. Sixty-nine 157 cores, 30-50 m deep (red circles in Fig. 2), were analysed and sampled for radiocarbon dating. For 158 159 each core lithology, grainsize, colour and accessory materials (vegetal and shell macrofossils, 160 carbonate concretions, redoximorphic features) were reported. The reaction to HCl was tested as a 161 qualitative indicator of the presence of calcium carbonate. Resistance was measured on fine-grained 162 sediments with a Pocket Penetrometer (PP). One hundred and seventy-two wood, peat, vegetalremains and bulk-soil samples were dated at the KIGAM AMS laboratory (Daejeon, Republic of 163 Korea) and at the Ion Beam Physics laboratory of the ETH (Zurich, Switzerland, Synal et al., 2007). 164 165 Samples were collected fresh from the innermost part of the core and desiccated in a 40° oven. The 166 humin fraction was separated through acid-alkali-acid (AAA) pre-treatment and AMS dated. Humic

167 acids, extracted from 9 bulk soil samples were also dated. Conventional ¹⁴C ages were calibrated using OxCal 4.4, with the IntCal 20 curve (Reimer et al., 2020). 168 More than 100 ore descriptions from previous works and from the database of the Geological, 169 Seismic and Soil survey of Regione Emilia-Romagna (yellow circles in Fig. 2) were reinterpreted 170 following calibration with new cores. These descriptions provide information on lithology, colour, 171 172 consistency, reaction to 10% HCl, pocket penetrometer measurement and radiocarbon dates. 173 Additional 204 radiocarbon dates derive from previously published works (Geological Map of Italy 174 to scale 1:50.000; Ravazzi et. al., 2006; Cacciari et al., 2017; Marabini and Vai, 2020). 175 176 4. RESULTS 177 178 4.1. Sedimentary facies 179 180 The sedimentological analysis of the studied cores led to the identification of several facies 181 182 associations which belong to alluvial, deltaic, estuarine and shallow-marine depositional 183 environments. To the purposes of this paper, we will describe in detail four main facies associations, which are part of the alluvial, upper delta-plain and inner estuary depositional systems. For other 184 185 facies associations, the reader is referred to Amorosi et al. (2017b, 2019) and Bruno et al. (2017). 186 4.1.1. Fluvial channel-related facies association (FCR) 187 188 189 This facies association is composed of cross-stratified or laminated coarse-to-fine sand with local 190 mud intercalations (Fig. 3). The occasional presence of floated woods or plant debris is recorded.

Marine or lagoonal macrofossils are absents, whereas shell fragments of freshwater gasteropods
are seldom encountered.

Based on lithology, sedimentary structures and on the lack of marine and lagoonal macrofossil, these coarse-grained deposits are interpreted as related to fluvial activity (in-channel deposition, bar migration, crevassing, proximal overbank; Miall, 1985; Toonen et al., 2012; Aslan, 2013).

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197 4.1.2. Freshwater swamp and marsh (SM)

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This facies association is dominated by fine-grained material, with abundant wood, vegetal remains and peat (Fig. 3). Peat horizons, 5-30 cm thick, are composed of poorly decomposed woody material and are separated by grey (2.5 YR 7.1, 7.5 YR 8/1), peat-free clays. Freshwater fossils were locally encountered. Consistency is very weak and PP values are < 1 kg/cm². Carbonate concretions and redoximorphic features are absent. This facies association shows poor lateral extent in the Upper Pleistocene succession. On the contrary, in Holocene distal deposits it is volumetrically dominant and only locally interrupted by channel-related sands.

The grainsize of this facies group denotes a low-energy depositional environment. The abundance of poorly decomposed organic material and the lack of oxidation suggest the setting of reducing conditions in submerged environments. The latter were likely isolated ponds during the Late Pleistocene, and swamp-marshes part of wider delta-plain and bayhead delta systems during the Holocene (Amorosi et al., 2017b; Bruno et al., 2017; Cacciari et al., 2020). Peat horizons separated by peat-free clays testify to intermittent sedimentation, with peat accumulation during sedimentation breaks (Hijma and Cohen, 2011; Ishii et al., 2016; Bruno et al., 2019).

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4.1.3. Poorly drained floodplain (PDFP)

This facies association is composed of grey (7.5 YR 6/1) silty clays with faint lamination given by millimetre-scale silt intercalations. Soft and isolated carbonate concretions (CC in Fig. 4) and sparse plant debris (UOM in Fig. 4) were seldom encountered. Redoximorphic features are lacking. Freshwater gasteropods were seldom encountered. Consistency is weak to firm. Pocket penetrometer values are in the range of 1-2 kg/cm².

Based on grainsize and fossil content, this facies association is interpreted as deposited in a distal fluvial environment, occasionally subjected to river flooding. Lack of redoximorphic features suggests a groundwater table fluctuating close to the topographic surface (poorly drained floodplain). The amount, distribution and type of secondary calcite are consistent with this interpretation.

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4.1.4. Well-drained floodplain (WDFP)

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This facies association is composed of firm-to-hard clayey silts and silty clays. Sedimentary 229 230 structures are absent. By contrast, polygonal structures, desiccation cracks and rootlets are 231 commonly observed. A well-expressed horizonation is given by colour and CaCO₃ distribution (Fig. 232 3). Dark horizons, showing weak or no reaction to HCl alternate with lighter horizons showing strong 233 reaction to HCl. Carbonate concretions, in the form of coalescent nodules, coatings, filaments and sub-horizontal bands are observed in lighter horizons. Redoximorphic features, in the form of 234 orange mottles or bands (RF in Fig. 6), are commonly observed at the mesoscale. Plant debris and 235 236 freshwater gasteropods are seldom encountered. PP values are $> 2 \text{ kg/cm}^2$.

As for facies PDFP, grainsize and fossil content point to a distal fluvial environment, occasionally
 inundated by river waters (Collinson, 1996). The alternation of dark and light horizons is interpreted

as the result of multiple phases of soil development interrupted by overbank deposition and paleosol burial. Dark colour reflects the accumulation and decomposition of organic matter in the topsoil (Bruno et al., 2020b). Carbonate distribution, with carbonate-free horizons overlying CaCO₃ enriched horizons, is interpreted as the result of carbonate leaching from the topsoil (A horizon) and illuviation to the underlying Bk horizon (Boul et al., 2011). Carbonate mobilization requires the percolation of meteoric waters in a vadose zone. We, thus, argue that the groundwater was located a few metres below the topographic surface during subaerial exposure (well-drained floodplain).

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248 **4.2.** Paleosol stratigraphy

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Paleosols were identified at three stratigraphic intervals in the Upper Pleistocene-Holocene
 succession of the Po Plain.

The lower interval is composed of a series of closely-spaced paleosols (paleosols P in Figs. 4, 5 252 and 6), whose cumulative thickness is in general < 5 m. Calibrated ages span from the ¹⁴C detection 253 254 limit (~50 kyr B.P.) to ~23 kyr B.P. (Figs. 5 and 6). These paleosols show A-Bk-Bw or A-Bk profiles. A 255 horizons, 30-50 cm thick, are decalcified (Bruno et al., 2020b; Amorosi et al., this volume) and present diffuse organic matter (DOM in Fig. 4) and slight bands of Fe oxides and/or hydroxides (Fig. 256 257 3). Undecomposed organic matter was not observed and carbonate concretions have locally been 258 reported from the base and the top of these horizons. Bk horizons are typified by high CaCO₃ content (30-40% of the total volume, Bruno et al., 2020b). Carbonate concretions are abundant as 259 260 coalescent nodules or whitish sub-horizontal bands (evolutionary stage II of Gile et al., 1981; 261 Machette, 1985). Transition to overlying and underlying horizons is in general gradational. Bw 262 horizons show moderate CaCO₃ content (Bruno et al., 2020b; Amorosi et al., this volume) and faint

263 mottles of Fe oxides and-or hydroxides. Evidences of clays mobilization (Bt horizons) were not 264 observed. At distal locations, these paleosols may show less clear horizonation (A-Bw profile) or may 265 be replaced by coeval peat-bearing swamp deposits (Fig. 5).

In the second interval, a single paleosol (paleosol PH, Figs. 4, 5 and 6) is encountered at the top 266 267 of a 3-10 m thick sedimentary unit composed of overbank muds and subordinated channel-related 268 sands (Last Glacial Maximum Sedimentary Unit, LGM SU in Figs. 4, 5 and 6). Textural characters of 269 paleosol PH vary from proximal to distal locations. At inland locations (core EM15, Fig. 3), paleosol 270 PH shares many characteristics with older P paleosols: (i) a carbonate-free, brown (10YR 4/4, 3/3) A horizon, up to 100 cm thick, with DOM and RF; (ii) a lighter Bk horizon with abundant nodules or 271 272 bands of secondary calcite (evolutionary stage II); (iii) a Bw horizon showing evidence of incipient 273 alteration. At more distal locations (cores EM1 and EM3, Fig. 3), paleosol PH exhibit a grey (5Y 4/1, 274 2.5Y 6/1) decalcified A horizon, 30-50 cm thick, with DOM. Plant debris, absent in the lower part, become increasingly abundant upsection and, at places, grade into the overlying peats (see core B3, 275 276 Fig. 4). The underlying Bk horizon, when present, is relatively thin (a few dm), with secondary calcite in the form of nodules, coatings or filaments (evolutionary stages I and II). Redoximorphic features 277 278 were not observed at the mesoscale along the whole soil profile.

In the third interval, dated to the Holocene, paleosols (H1 and H2 in Figs. 4, 5 and 6) have been observed only at inland locations. At distal locations, they are replaced by coeval estuarine and deltaic deposits. H paleosols have poor correlation potential and their textural features vary dramatically from a place to another, without specific spatial trends. In few localities, they show characteristics similar to those of their Pleistocene counterparts. At places, they are barely identifiable due to a poorly developed horizonation. Archeological artifacts are increasingly abundant in younger paleosols. Paleosols P and PH bound nearly tabular sediment packages and are locally replaced by coeval incisions filled with FCR or SM deposits (see cores CNT, RN3 and EM2, Fig. 5). On the contrary, Holocene paleosols overlie or are overlain by CR facies (see cores MW and 15, Figs. 5 and 6). Alluvial units delimited by these soil horizons are poorly extended and typically wedge out laterally or downdip (Figs. 5 and 6).

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- 293 4.3. Radiocarbon dating of paleosols
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Upper Quaternary successions provide the great opportunity to link pedo-sedimentary events to the detailed climatic and eustatic history described in a vast literature (eg. Clark and Huybers, 2009; Lambeck et al., 2011; Rasmussen et al., 2014). Assessing the timing of soil development and burial is crucial to disentangle the relative influence of global and local factors. This chapter summarizes observations deriving from the analysis of a chronological database composed of 376 radiocarbon dates (160 from paleosols).

301 Along single cores, multiple dating of A horizons yielded stratigraphically ordered ages, spanning 302 3-5 kyr (see cores B3, RW28 and 13, Figs. 4, 5C and 6). Only in core SPR (Fig. 5C) two samples from 303 the base and the top of the same A horizon provided similar ages. In distal cores, plant debris from 304 topmost A horizons yielded ages that are considerably younger (up to 5 kyr) than those from bulk-305 soil samples and just a few centuries older than those from overlying SM peats (see cores EM1, EM3 306 and B3, Figs. 3, 4 and 5). Dates from A horizons indicate aggradation at extremely low rates (0,05-307 0,23 mm yr-1) during soil development. Accumulation rates above or between paleosols are at least 308 one order of magnitude grater (> 1 mm/yr). Distinct ages were also obtained from different fractions extracted from the same sample. Particularly, humic acids yielded, in 9 cases out of 10,
ages older than those obtained from the humin fraction (Table 1).

Dates from bulk-soil samples change also laterally, along the same paleosol, spanning in a range of 4-6 kyr (Fig. 3). No specific trend was observed from proximal to distal locations. Ages also change over short distances, as observed in the Bologna area. In the Two Tower Medieval district, where distance between cores is just 10-20 m, six samples from paleosol P3 yielded ages in the range of 3-4 kyr (Fig. 6).

316 The frequency distribution of radiocarbon dates from paleosols (grey rectangles in Fig. 7) shows a major peak between 14 and 10 kyr B.P., related to the PH paleosol. Samples from the P paleosols 317 318 yielded ages clustered around three minor peaks (P1, P2 and P3, Fig. 7), at 46-41, 38-32 and 29-25 319 cal kyr B.P., respectively. The overall decay with time in the number of ¹⁴C dates across the 320 Pleistocene simply reflects the fact that older sediments were penetrated by a progressively lower number of cores. A decline in the number of ¹⁴C dates from soil samples is also recorded towards 321 younger ages, together with a sharp increase in dates from SM and PDFP deposits. Ages from 322 Holocene soil samples are clustered around two peaks. The older peak reflects, in part, the youngest 323 324 ages from paleosol PH and in part dates from a distinct paleosol (H1 in Figs 4, 5 and 6) observed at 325 several inland locations. In the Bubano Quarry (BQ in Fig. 5C), 5 m-thick overbank strata separate H1 from PH. In the Bologna area (core 16, Figs. 4 and 6), paleosol H1 bears Neolithic and Eneolithic 326 327 artefacts (Bruno et al., 2013). The second peak (H2, Fig. 7) corresponds to another archeological level, rich in Late Bronze Age, Iron-Age and Roman rests (IR paleosol of Bruno et al., 2013), which 328 was buried in large part of the Po Plain during the Middle Ages (see core CRV, Fig. 5A). Locally, this 329 paleosol is still exposed (Fig. 5). Interestingly, the maximum number of ¹⁴C dates from facies FCR is 330 331 recorded between H2 and H1.

334 **5. DISCUSSION**

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5.1 Use and limitations of radiocarbon for the estimation of the exposure age of paleosols

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338 The use and reliability of radiocarbon dates from paleosols have been widely debated because, unlike plant macrofossils, which preserve the particular ¹⁴C content of the moment of grow, soils 339 are open systems where organic carbon is continuously incorporated, humified and-or oxidized to 340 CO₂ before burial (Orlova and Panychev, 1993; Martin and Johnson, 1995; Gaudinski et al., 2000). 341 Therefore, ¹⁴C ages from paleosols reflect the mean residence time of the organic carbon plus the 342 343 time of burial (Matthews, 1980; Geyh et al., 1983). The amount of organic carbon added to a soil 344 and the velocity of its transformation processes change from place to place as a function of climate, vegetation and parent material (Reimer et al., 2020). Moreover, turnover and breakdown of organic 345 matter may lead to the redistribution of ¹⁴C ratios in different soil fractions. As a result, apparently 346 conflicting ages can be obtained from the same paleosol (Figs. 5 and 6) and from different soil 347 348 fractions (Table 1). Because of this inherent variability, many authors concluded that radiocarbon 349 dates from soils are unreliable and the concept of "age" cannot be strictly applied to paleosols (Gilet-Blein et al., 1980; Becker-Heidmann et al., 2002). However, if not subjected to post-burial 350 351 contamination (e.g. rootlets), paleosols should yield ages scattered within its formation period. The 352 estimation of the length of this period, which can attain several millennia, is an intriguing goal for Quaternary paleosol stratigraphers. 353

The lack of evidence of clay mobilization, the evolutionary stage of secondary calcite and the limited rubefaction suggest that Po Plain paleosols evolved over relatively short periods. Radiocarbon ages from deposits below and above paleosols constrain their exposure period within 357 10-12 thousand years (Figs. 5 and 6). Dates from paleosols are within the narrower range of 4-6 millennia and provide a robust constrain for stratigraphic correlations over long distances (Amorosi 358 et al., 2017a). More complex is the precise assessment of the ages that mark the onset and end of 359 pedogenesis for each paleosol, even if radiocarbon dates are available from the base and top of A 360 361 horizons. Previous work has demonstrated that, although some organic carbon can remain in soils 362 for millennia (Feng, 2009), the oldest organic fraction accumulated at the onset of pedogenesis may 363 have been completely replaced by younger organic carbon (Scharpenseel, 1971; Alexandrovskiy and Chichagova, 1998) introduced by roots in the deepest soil horizons. Thereby, radiocarbon dates 364 from organic-sediment samples collected from the base of the A horizon may not be representative 365 of the onset of soil formation (Gilet-Blein et al., 1980; Alexandrovskiy and Chichagova, 1998). Several 366 factors influence carbon turnover times in soils (Kleber, 2010; Chen et al., 2013; Pandey et al., 2014). 367 368 Particularly, carbon renewal is slower in colder climates and in deep soil horizons (Alexandrovskiy and Chichagova, 1998; Gaudinski et al., 2000; Fontaine et al., 2007). Therefore, for soils that 369 developed in cold climatic conditions, such as paleosols P3, which formed at the onset of Last Glacial 370 Maximum, dates from the base of the A horizon should better approximate the onset of 371 372 pedogenesis. This assumption is corroborated by dates from core B3 (Figs. 3, 4 and 5), where the 373 oldest age from paleosols P3 partially overlaps the age from underlying paleosol P2. For paleosols developed in warmer periods (PH and H paleosols), ages from the base of A horizons may post-date 374 375 the onset of pedogenesis.

Sediment samples collected from the top of A horizons may be older than burial time, because a certain amount of time is required for litter humification and mineralization (Chichagova and Cherkinsky, 1993). At inland locations of our study area, where O horizons are generally not preserved, burial ages cannot be assessed precisely. Available dates permit to assert, for example, that paleosol PH was buried after 10.3 kyr B.P. in core 13 (Figs. 4 and 6) and between 14.3 and 8.6 kyr B.P. in core 201S4 (Fig. 5). Conversely, at distal locations, plant macrofossils preserved in the topsoil provides ages that approximate the paleosol burial age with an error of a few centuries. For example, in EM1, where estuarine sedimentation was already established around 8.1 kyr B.P., burial of paleosol PH occurred around 8.5 kyr B.P. (Figs. 3, 4 and 5).

Age/depth gradients, like those observed in the Po Plain paleosols, have been reported from 385 386 several soil profiles (Matthews, 1981; Matthews and Dresser, 1983; Alexandrovskiy and Chichagova, 387 1998; Becker-Heidmann et al., 2002) and can possibly be explained by the fact that the amount of carbon added to the topsoil through litter accumulation is in general order of magnitudes greater 388 than the one introduced in deeper soil horizons via roots exudation (Fontaine et al., 2007). The 389 measured accumulation rates, incompatible with alluvial sedimentation, suggest that litter 390 accumulation accounted for soil aggradation during exposure, with a possible contribution of wind-391 392 transported dust. Therefore, paleosols from the Po Plain represent condensation horizons, rather than depositional hiatuses. On the contrary, sedimentation between pedogenetic phases seems a 393 quasi-instantaneous process (see ages from P2 and P3 in core B3, Figs 4 and 5). Paleosols 394 395 aggradation is also testified by the presence of carbonate concretions in lower A horizons, likely 396 related to the upward migration of the eluvial-illuvial (A-B) boundary, and by vertically stacked sub-397 horizontal bands of secondary calcite in Bk horizons (Fig. 3). Carbonate concretions in upper A horizons, on the other hand, reflect post-burial illuviation from overlying paleosols. For this reason, 398 399 we avoided radiocarbon dating of carbonates and removed secondary calcite in acid steps of the AAA sample preparation, before AMS analysis. Humic acids are also considered as a mobile 400 compound, which may yield anomalously young ages due to contamination from overlying strata 401 402 (Matthews, 1980). In our case, the oldest ages are provided by humic acids (Table 1). We thus 403 excluded contamination from overlying strata during each phase of pedogenesis.

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5.2. Pedo-sedimentary evolution of the Po Plain and controlling factors

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With the chronological constraint of 376 radiocarbon dates, the alluvial succession of the southern Po Plain provides an unprecedent record of pedo-sedimentary events across the last 50 kyr.

Aggradationally stacked paleosols 'P' developed during the Late Pleistocene, when the Adriatic 411 coastline migrated in response to the stepwise sea-level fall toward its lowstand position, ~300 km 412 away from the modern one (Pellegrini et al., 2018). Among P paleosols, only P3 developed during a 413 phase of sea-level drop (~30–50m between 30 and 24 cal BP, Peltier and Fairbanks, 2006; Fig. 414 415 7B). Conversely, P1 and P2 developed in a period characterized by low-amplitude sea-level oscillations (Chappell, 2002; Fig. 7B). Far from the influence of sea-level change, pedo-alluvial 416 processes are controlled by the balance between sediment supply and water discharge (Blum and 417 Törnqvist, 2000). Low sediment supply-discharge ratios result in channel incision and soil 418 development in the adjacent interfluves, whereas high sediment supply-discharge ratios promote 419 420 fluvial aggradation and paleosol burial (Demko et al., 2004). Bruno et al. (2020b) documented that 421 soil development and burial during the Late Pleistocene was mainly controlled by changes in 422 vegetation driven by climatic oscillations at the scale of Bond cycles (Bond et al., 1993). Soils started 423 to form at the onset of Bond cycles, when warm climate conditions favoured the spread of dense forests (Magri, 1999; Fletcher et al., 2010), which protected the Apennine drainage basins from 424 erosion (low sediment supply). Soil development was systematically interrupted during the 425 426 subsequent cold phases, at the culmination of Bond cycles, corresponding to North Atlantic Heinrich 427 events (light blue vertical lines in Fig. 7B). These periods record generalized forest retreats (Fletcher et al., 2010; Helmens, 2014), which enhanced erosion in the drainage basins and sediment transfer 428 to the alluvial plains. Increased sediment supply promoted the filling of fluvial incisions and the 429

burial of formerly exposed interfluves by overbank sedimentation. Thin sediment packages were
deposited between two successive phases of subaerial exposure (Figs. 3 to 6). These deposits (and,
partly, the underlying paleosols) were modified during the successive pedogenetic phase. A similar
response to millennial-scale climate oscillations has been observed in loess paleosol-bearing
successions (Rousseau et al., 2011, 2007; Moine et al., 2017; Zeeden et al., 2018).

The Last Glacial Maximum was characterized by deposition of thick and widespread fluvialchannel gravel-sand bodies in the Po Plain (Ravazzi et al., 2012; Campo et al., 2016) and in other alluvial systems worldwide (Lewis et al., 2001; Briant et al., 2005). Alpine glaciers, which extended down to the Po Plain margin, supplied huge amounts of sediments to the related fluvio-glacial systems (Fontana et al. 2014). In the Apennine domain, high sediment supply, favoured by a predominantly herbaceous vegetation cover of drainage basins (Vescovi et al., 2010), resulted in the deposition of the LGM SU, which bears poor evidence of pedogenesis (Figs. 4, 5 and 6).

442 Above the LGM SU, PH is the most widespread paleosol in the Po Plain. This surface yielded 443 radiocarbon dates spanning between ~17 and 6 kyr B.P. (Fig. 8), but mostly clustered around 14 and 10 kyr B.P. (Fig.7). The precise assessment of the age of PH formation is hampered by paucity of 444 445 dateable material from the underlying LGM SU. However, available dates indicate that the onset of 446 post-LGM pedogenesis was diachronous across the Po Plain, without a specific spatial trend (Fig. 8). Whereas in the Bologna area PH started to develop at least 16.8 kyr B.P. (see core RW28, Figs 6 and 447 448 8), in the Romagna plain (Bubano Quarry - BQ, in Figs 5C and 8) pedogenesis started after 14.4 kyr B.P. (Ravazzi et al., 2006). We argue that post-LGM pedogenesis was favoured by decreased 449 sediment supply related to the progressive afforestation of the area (Pini et al., 2010; Vescovi et al., 450 2010; Campo et al., 2020), with diachroneity possibly related to particular local conditions (e.g. 451 452 distance from active channels). North of the Po River, in the Alpine tributary system, fluvial incision

occurred soon after glacial retreat (~17 kyr B.P., Monegato et al., 2017), due to sediment trapping
in lakes which formed in place of abandoned glacial amphitheatres (Fontana et al., 2014).

Interestingly most of the ¹⁴C dates from paleosol PH post-date the Oldest Dryas (OD in Fig. 7). 455 The clustering of dates around the Younger Dryas in the Po delta plain (Figs. 7A and 8), suggested a 456 possible control of this cold event on paleosol formation (Amorosi et al., 2017b, Morelli et al., 2017). 457 458 The wider dataset shown in this work highlights that pedogenesis started in many sites of the Po 459 Plain at least during the Bølling and Allerød warm phases (BA in Fig. 7A). Unlike cold phases at the 460 culmination of pre-LGM Bond cycles, the Younger Dryas cold spell did not lead to paleosol burial. This event was probably too short-lived to promote the filling of fluvial incisions and overbank 461 sedimentation. Paleosol development recorded after LGM in other alluvial (Mol, 1997; Choi, 2005; 462 van Balen et al., 2010; Janssens et al., 2012) and eolian (Kaiser et al., 2009; Hošek et al., 2017) 463 464 successions, indicates a likely climate control at a supra-regional spatial scale. Progressively warmer conditions lead to widespread afforestation and reduced sediment supply (Fletcher et al., 2010; 465 Helmens, 2014). 466

Whereas Alpine rivers still flow entrenched in Pleistocene deposits, Apennine incisions were filled 467 468 during the Early Holocene (see cores RN3 and EM2, Fig. 3C) and paleosol PH was buried soon after. 469 In the coastal plain, where paleosol PH is encountered at greater depths beneath estuarine deposits, 470 burial ages show an overall inverse power law with elevation (Fig. 9) that reflects the progressive 471 drowning of the area during the early Holocene sea-level rise (Bruno et al., 2017). Slight dispersion 472 of data may reflect post-burial deformation of paleosol PH by differential subsidence and 473 neotectonics (Amorosi et al., 2021). In the area under marine influence (shades of blue in Fig. 10), 474 burial ages decrease landwards from ~11.5 to 5.4 kyr B.P.. This overall spatial trend is not recorded 475 in two isolated areas, where paleosol PH and older paleosols are upbended (see also Fig. 5C). In 476 cross-sections, the progressive drowning of paleosol PH is testified by onlap terminations of Holocene flooding surfaces (Fig. 5A). These horizons, associated to prominent facies backstepping, are coeval to the main Holocene melt water pulses (MWP-1b, 1c, 1d; Fig. 7B), which forced the landward migration of the Po estuary across the study area. The effects of MWP-1A are not recorded in the study area, which was too far from the coastline at that time (Storms et al., 2008).

Pedo-sedimentary features of paleosol PH reflect this evolutionary history. Carbonate 481 482 redistribution along the soil profile and relatively low amounts of organic carbon in the A horizons 483 (Amorosi et al., this volume) are indicative of a low groundwater table (Kalisz et al., 2010), enhanced by fluvial incision during the early phases of soil development (Tebbens et al., 1999). Preserved plant 484 debris in the topsoil, radiocarbon dated to the early Holocene (Figs. 4 and 5), reflects the progressive 485 rise in groundwater table (Bohacs and Suter, 1997) up to the setting of waterlogged conditions and 486 to the accumulation of peat beds (Diessel, 1992; Richardson and Vepraskas, 2001; Stolt and 487 488 Rabenhorst, 2011). The rise of the groundwater table, forced by sea-level rise (Kosters and Suter, 1993; Törnqvist, 1993; Wadsworth et al., 2003), led to the progressive filling of Apennine fluvial 489 incisions by fine-grained and organic-matter-rich estuarine strata (facies SM, cores RN3 and EM2 in 490 491 Fig. 5C). The lack of fluvial sands (facies FCR) at the base of fluvial incisions, suggests that valley 492 filling was promoted by eustasy rather than by an upstream increase in sediment supply (Kvale and 493 Archer, 2007).

Beyond the area of marine influence, at elevations > -3 m, the burial age of paleosol PH is substantially unpredictable (Figs. 9 and 10). The dispersion of burial dates in this upstream sector reflects the dynamics of Holocene sedimentation, dominated by avulsion processes and by the deposition of poorly extended alluvial units (Figs. 5 and 6). This aspect is particularly evident in the Bologna area (Fig. 6): in core 16, for example, paleosol PH was already buried around 8.2 kyr B.P., when paleosol H1 was still developing (see also Fig. 4); in the nearby core 15, at a distance of just 224 m, the burial of paleosol PH occurred only around 7 kyr B.P.. The transition to an avulsive sedimentation pattern at the onset of the Holocene, recorded in several worldwide fluvial systems (Phillips, 2011; Stouthamer et al., 2011), is an indirect consequence of the landward migration of the coastline, which dramatically reduced the area available for sediment storing and lowered rivers longitudinal gradients. Decreased stream power, and capability to erode banks and create space for sediment storage, led to river aggradation, crevassing and avulsions (Blum et al., 2013).

The discontinuity of Holocene paleosols (Figs. 5 and 6) reflects the limited areal extent of the sedimentary units on which they developed. Avulsive sedimentation patterns and crevassing also determined laterally variable maturity of Holocene paleosols. Indeed, soil development was interrupted only in restricted areas affected by alluvial deposition. In sediment-starved areas, soils continued to evolve. Lateral changes in soil maturity are observable in modern soils, which started to develop during the Late Bronze age (areas where H2 is still exposed – Fig. 5), and after the Middle Ages (areas where H2 is buried - Fig. 5).

The peak in radiocarbon dates from fluvial channel-related deposits, centred around 5-6 kyr B.P. 513 (Fig. 7), corresponds to a generalized re-organization of the river network at the onset of 514 Neoglaciation (Piovan et al., 2012; Bruno et al., 2015; Amorosi et al., 2017b), which led to the burial 515 516 of paleosol H1. Slope instability and enhanced sediment production, recorded in the Northern 517 Apennines since 5.5 kyr B.P. (Soldati et al., 2006; Picotti and Pazzaglia, 2008), likely reflect widespread deforestation of the Apennine drainage basins during the early Eneolithic (Cremaschi 518 519 and Nicosia, 2012). The progressive return to stable floodplain conditions was paralleled by 520 improved techniques of land reclamation and management from the late Bronze age to late Antiquity (Bruno et al., 2013). 521

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524 6. CONCLUSIONS

Based on the sedimentological analysis and correlation of more than 170 core data, with the chronological constraint of 376 radiocarbon dates, this study provided new data on: (i) the complex relationships between pedogenetic, fluvial and costal processes in alluvial and paralic environments; (ii) the relative influence of global (i.e. climate and eustasy) and local (e.g. river avulsions) factors on soil development and burial; (ii) the use and limitations of radiocarbon to assess the exposure age of paleosols.

In the Upper Pleistocene-Holocene mud-prone succession of the Po Basin, vertically stacked, weakly developed paleosols testify to intermittent pedogenesis, periodically interrupted by overbank sedimentation. Radiocarbon analyses permitted to define the duration of the main phases of pedogenesis and to compare them to climate and eustatic curves.

Radiocarbon dates from paleosols (bulk-soil and plant-debris samples) and from underlying and overlying sedimentary units indicate exposure periods encompassing 4-6 millennia. This estimate is consistent with pedological features (e.g. type of secondary calcite, lack of illuvial clay, limited rubefaction), which denote incipient pedogenesis. Stratigraphically ordered ¹⁴C calibrated ages within A horizons revealed that Po Plain paleosols are condensation horizons, which experienced slow aggradation during exposure (0,00-0,23 mm/yr) due to accumulation of plant litter, with a likely contribution of wind-transported dust.

543 Millennial-scale climate oscillations and glacio-eustasy concurred in driving the pedo-544 sedimentary evolution of the Po Plain during the Late Pleistocene. In particular, during the MIS 3, 545 pedogenetic events were mainly controlled by changes in vegetation driven by climatic oscillations 546 at the scale of Bond cycles. Paleosols developed in well-drained floodplain environments, dissected 547 by fluvial incision, during relatively warm periods. Overbank sedimentation and paleosol burial occurred during colder phases. During the Last Glacial Maximum, high-sediment-supply hindered
 pedogenesis and led to the accumulation of 3-10 m-thick overbank strata.

550 Widespread soil development (paleosol PH) occurred at the end of the Last Glacial Maximum, 551 following glaciers retreat and afforestation of drainage basins. Burial of paleosol PH was 552 diachronous as indicated by radiocarbon dates from vegetal remains preserved at the topsoil. At 553 distal locations, paleosol PH was progressively drowned and buried under estuarine sediments 554 during the Holocene phases of post-glacial sea-level rise. Beyond the area of marine influence, burial 555 ages of paleosol PH change from a place to another without specific spatial trends, reflecting 556 upstream fluvial sedimentation dominated by avulsions.

Holocene (H) paleosols, observed at inland locations, grade seaward into coeval estuarine and
deltaic deposits. These pedogenized horizons show a poor correlation potential and laterally
variable degree of maturity that reflect avulsive sedimentation patterns and widespread crevassing.
Human activities (e.g. deforestation and land reclamation) became an increasingly important factor
affecting pedo-sedimentary processes during the last 6 kyr.

This study highlights the potential for paleoenviromental reconstructions of upper Quaternary sedimentary successions. The pedo-stratigraphic approach adopted here, relying upon a robust chronological framework, permitted to assess the temporal and spatial scale of pedogenetic events and to discern the relative contribution of distinct controlling factors that typically act contemporaneously.

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1010 Figures and captions



Fig. 1 – A. Geological map of the Po Plain, with location of the study area and of the seismic profile
of Fig. 1B (modified after Bruno et al., 2019). The extent of Pleistocene alluvial deposits is from
Fontana et al. (2014). Buried structures are modified after Burrato et al., (2003). B. Interpreted deep
seismic profile (from Regione Emilia-Romagna & Eni-Agip, 1998, location in Figure 1A), with
identification of the major blind thrusts (red lines) and main stratigraphic unconformities (lines from
A–F) within the Po Basin fill (coloured area).



Fig. 2 – A. Study area with location of cores (red circles), core descriptions (yellow circles) and crosssections of Fig. 5. B. Close-up on the Bologna area, with location of cores shown in Fig. 6. C. Location

- 1023 of cores drilled in the Two-Towers Medieval district (modified after Bruno et al., 2020b).
- 1024



Fig. 3 – Representative photographs of paleosols from cores EM1, EM3 and B3 (location in Fig. 2).
WDFP: well-drained floodplain; PDFP: poorly-drained floodplain; SW: swamp; FCR: fluvial-channelrelated deposits. For pedological horizons A, Bk and Bw, see text. The white dates (cal kyr B.P.) are
from bulk-sediment samples, whereas the yellow ones are from plant macrofossils.





Fig. 4 – Sedimentological characters and radiocarbon ages of paleosols P2, P3, PH and H1 from cores
RW28, 16, 17, EM1, EM20 and B3 (for location see Fig. 2). DOM: dissolved organic matter; UOM:
undecomposed organic matter; RF: redoximorphic features; CC carbonate concretions (the number
indicates the evolutionary stage, from Gile et al., 1981; Machette et al., 1985). WDFP: well-drained

floodplain; PDFP: poorly-drained floodplain; SW: swamp; FCR: fluvial-channel-related deposits. The
black dates (cal kyr B.P.) are from bulk-sediment samples, whereas the red ones are from plant
macrofossils.

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Fig. 5 – Upper Pleistocene and Holocene stratigraphy of the Po Plain depicted in cross-sections with
W-E (A) and S-N (B and C) orientations (location in Fig. 2A). LGM SU: Last Glacial Maximum
Stratigraphic Unit.

1044



Fig. 6 – Paleosol-based correlations in the Bologna Area. LGM SU: Last Glacial Maximum
Stratigraphic Unit. Location of cores in Fig 2.



Fig. 7 – A. Distribution of radiocarbon dates from bulk-soil and plant-macrofossil samples. A major 1053 peak in dates from bulk-soil samples corresponds to paleosol PH; five minor peaks are related to 1054 1055 paleosols P1, P2, P3, H1 and H2. Notable is the lack of radiocarbon dated soils around 20 cal ky B.P. 1056 (LGM). The sharp increase in radiocarbon dates from plant-macrofossils after 10 ky B.P., paralleled by a decline in the number of dates from bulk-soil samples, reflects the onset of estuarine-deltaic 1057 sedimentation in the distal Po Plain and the burial of paleosol PH. Red areas correspond to 1058 interstadials, according to Bond et al., (1997) and Rasmussen et al. (2014). OD: Oldest Dryas; YD: 1059 Younger Dryas. Light blue vertical bars indicate Henrich events (Bond et al., 1992). B. Sea-level 1060 oscillations during the last 50 ky. 1061



Fig. 8 – Radiocarbon dates (black numbers, cal ky B.P.) from paleosol PH (bulk soil samples) in the Po Plain (A) and in the Bologna urban area (B). Blue numbers are the youngest ages from the underlying LGM stratigraphic unit. The present-day elevation (metres above modern sea-level) of paleosol PH is depicted through 5 m-spaced isolines. BQ: Bubano Quarry.



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Fig. 9 – Burial ages of paleosol PH plotted against its present-day elevation (metres above modern sea-level). Black rectangles are radiocarbon dates from bulk soil samples. Red rectangles are radiocarbon dates from plant macrofossils. Burial ages from samples at elevation < - 3 m (dashed line) show an overall inverse proportionality with elevation. On the contrary, no specific trends are observable in burial-age distribution at elevation > - 3 m.



Fig. 10 – Map showing burial ages (cal kyr B.P.) of paleosol PH in the Po Plain (A) and in the Bologna urban area (B). The shades of blue depict progressively younger burial ages upland. In the green area burial ages do not show specific spatial trends. Red numbers represent burial ages or intervals measured along single cores (black circles). The map also depicts the present-day elevation (metres above modern sea-level) of paleosol PH through 5 m-spaces isolines. Black dots are the countered data points. The dashed line represents the maximum landward extent of Holocene estuarine and delta plain deposits.











ì	50504240	LI1 silty day	brown (7.5YR 5/1)	-	DOM, RF, Eneolithic artifacts	weak	WDFP (A horizon)
	02104100	clayey sit	grey (7.5YR 6/1) with orange mottles	firm	1152 Se	moderate	WDFP (Bw horizon)
		dayey and sandy silt					FOR
10	11870.239 -	silty clay with silt intercaltaions	grey (7.5YR 6/1)	weak			POFP
		2ª DH	dark grey (10YR 3/1)		DOM, UOM on top	weak	WDFP (A horizon)
12		∿o PFI dayeysit	gray (5Y 7/1)	firm	CC (2), 85	strong	WDFP (8k horizon)
14	25850	P3	dark grey (5Y 4/1)	hard	DOM, RF	weak	WOFP (A horizon)
1		00	olive grey (5Y 6/1)	1000	CC (2), RF	strong	WOFP (8k horizon)

	EM20							
17	8870.00			grey (2.5YR 7/1)	NOTE OF C	Abundant UOM	weak	SM
	1455	0.0	silty day	light grey (2.5YR 8/1)	weak	CC (1) sparse UOM	moderate	PDFP
19	11660.100	PH		grey (2.5Y 6/1) light grey (2.5YR 8/1)	firm	DOM, sparse UOM on top CC (2), RF	weak strong	WDFP (A horizon) WDFP (8k horizon)

	2	EM1						
17	7870410			light grey (7.5YR 8/1)	very weak	Abundant UOM, peat, locally DOM	moderate, weak on peat	SM
	8510.00		silty day		weak			
19	11430.895	00 000		grey (2.5Y 6/1)	firm	DOM, sparse UOM on top	weak	WDFP (A horizon)
		PHee		light grey (2.5YR 8/1)	hard	CC (2)	strong	WDFP (8k horizon)

7380.40			light grey (clay, SYR 8/1) brown (peat, 2.5Y 3/2)	very weak	UOM, peat at the base	moderate, weak on peat	SM SM (O horizon)
7960.00-		PH silty clay silty clay- sandy silt silty clay, clayery silt	dark grey (SY 4/1) -	firm	sparse UOM	weak	WDEP (A bosisse)
10000+120	PH		grey (10R 7/1) with orange (10 YR 8/6) mottles		RF	moderate	WDFP (Bw horizon)
. 8							FOR
27826	. 14			firm			WDFP (Bw horizon)
32790+440~	P3	silty day	dark grey (7.5Y 5/1)	hard	DOM, RF	weak	WOFP (A horizon)
	00		beige-brown (10YR 6/3)		CC (2,3), RF	strong	WDFP (8k horizon)
32630.00	10.		beige (7.5Y 8/2)	firm	RF	100000	WDFP (Bw horizon)
34030400-	50 0 DO		dark grey (10VR 5/1)		DOM, RF	weak	WDFP (A horizon)















Declaration of interests

 \boxtimes The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

□The authors declare the following financial interests/personal relationships which may be considered as potential competing interests: