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 pedogenesis, 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 ^{14}C 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 pedogenesis 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 fluvial 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 reclamation affected pedo-sedimentary processes during the last 6 kyr.

Keywords

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

1. INTRO

In alluvial settings, the development and burial of soil horizons are intimately related to fluvial activity (Kraus and Bown, 1993) and governed by a variety of factors. High sediment supply and fluvial aggradation inhibit pedogenesis (Demko et al., 2004), whereas low sediment supply and fluvial incision promote soil development in interfluvial areas (Blum and Törnqvist, 2000). Climate influences the amount of sediment delivered to the basins, modulating the magnitude and type of 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 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 the intrinsic dynamics of the fluvial apparatus (Kraus, 1999; Cleveland et al., 2007; Hajek and Wolinsky, 2012). In coastal settings, fluvial dynamics are strongly influenced by sea-level oscillations (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 reconstructed in ancient alluvial successions, where the geometric relationships between paleosols and associated fluvial sediment bodies are superbly exposed (Gibling and Bird, 1994; McCarthy and Plint, 2003; Dubiel and Hasiotis, 2011; Soares et al., 2020). However, the role of distinct controlling factors on paleosol development and burial can hardly be assessed (Wright and Marriott, 1993; 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 (Rasbury et al., 1997; Hofmann et al., 2017). Furthermore, diagenesis may have substantially modified soil properties after burial, hindering comparison with modern soils and the reconstruction 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). However, numerous uncertainties remain in the interpretation of ^{14}C dates from paleosols. Organic 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, 2010).

This study is based on refined pedostratigraphic reconstructions and on a vast ^{14}C dataset which permitted to chronologically constrain the major architectural and facies changes. The primary aim of this research is to understand the causes and mechanisms of soil development and burial in an alluvial-deltaic setting and its relation to fluvial and coastal dynamics. To this purpose, we focused on Upper Pleistocene-Holocene paleosols and on their stratigraphic relationship with older, coeval and younger pedo-sedimentary units. Secondary objectives are to discuss: (i) the relative role of global and local controlling factors on soil development and burial; (ii) the potential use (and major limitations) of radiocarbon to assess the exposure age of paleosols.

2. GEOLOGICAL SETTING

2.1. Structural setting and basin-scale stratigraphy

The Po sedimentary basin is enclosed in a natural amphitheatre bordered by the Alps, to west and to the north, and by the Apennines, to the south. The eastward prolongation of the Po Plain is 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., 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 thrust systems: from west to east, the Monferrato, Emilia and Ferrara-Romagna Arcs (Ghielmi et al., 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 variations in thickness are controlled by lateral changes in subsidence rates (Campo et al., 2020), ranging between zero and 2 mm/y (Bruno 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 units (Ghielmi et al., 2013). Widespread alluvial sedimentation started in the Po Plain around 870 kyr B.P. (Muttoni et al., 2003, 2011; Gunderson et al., 2014). The Po River, flowing in W-E direction, divides the alluvial plain in two sectors (Fig. 1A). In the norther sector, Alpine tributaries flow 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 and paralic succession (Amorosi et al., 2017a, b). Early Holocene estuarine deposits typically onlap onto Upper Pleistocene alluvial strata, testifying to the progressive drowning of the alluvial plain 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 (Amorosi et al., 2017b, 2019).

2.2. Paleosols of the Po Basin

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 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 responds to several local names: ‘Caranto’ in the Venetian Plain (Mozzi et al., 2003), ‘Ringhiera paleosol’ in the Romagna area (Ravazzi et al., 2006), ‘YD paleosol’ in the Po Delta Plain (Amorosi et al., 2017b; Morelli et al., 2017), ‘PH paleosol’ in the Bologna urban area (Amorosi et al., 2014; Bruno et al., 2020b; this name will be used in this paper). Although a wide literature reported the existence 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 (Bruno et al., 2015, 2013).

3. METHODS

This work is based on the sedimentological and stratigraphic analysis of core material. Sixty-nine cores, 30-50 m deep (red circles in Fig. 2), were analysed and sampled for radiocarbon dating. For each core lithology, grainsize, colour and accessory materials (vegetal and shell macrofossils, carbonate concretions, redoximorphic features) were reported. The reaction to HCl was tested as a qualitative indicator of the presence of calcium carbonate. Resistance was measured on fine-grained sediments with a Pocket Penetrometer (PP). One hundred and seventy-two wood, peat, vegetal-remains and bulk-soil samples were dated at the KIGAM AMS laboratory (Daejeon, Republic of Korea) and at the Ion Beam Physics laboratory of the ETH (Zurich, Switzerland, Synal et al., 2007). Samples were collected fresh from the innermost part of the core and desiccated in a 40° oven. The humin fraction was separated through acid-alkali-acid (AAA) pre-treatment and AMS dated. Humic acids, extracted from 9 bulk soil samples were also dated. Conventional ^{14}C ages were calibrated using OxCal 4.4, with the IntCal 20 curve (Reimer et al., 2020).

More than 100 core descriptions from previous works and from the database of the Geological, Seismic and Soil survey of Regione Emilia-Romagna (yellow circles in Fig. 2) were reinterpreted following calibration with new cores. These descriptions provide information on lithology, colour, consistency, reaction to 10% HCl, pocket penetrometer measurement and radiocarbon dates. Additional 204 radiocarbon dates derive from previously published works (Geological Map of Italy to scale 1:50.000; Ravazzi et. al., 2006; Cacciari et al., 2017; Marabini and Vai, 2020).

4. RESULTS

4.1. Sedimentary facies

The sedimentological analysis of the studied cores led to the identification of several facies associations which belong to alluvial, deltaic, estuarine and shallow-marine depositional 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 facies associations, the reader is referred to Amorosi et al. (2017b, 2019) and Bruno et al. (2017).

4.1.1. Fluvial channel-related facies association (FCR)

This facies association is composed of cross-stratified or laminated coarse-to-fine sand with local 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).

4.1.2. Freshwater swamp and marsh (SM)

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/cm^2. 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).

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/cm^2.

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.

4.1.4. Well-drained floodplain (WDFP)

This facies association is composed of firm-to-hard clayey silts and silty clays. Sedimentary structures are absent. By contrast, polygonal structures, desiccation cracks and rootlets are commonly observed. A well-expressed horizonation is given by colour and $CaCO_3$ distribution (Fig. 3). Dark horizons, showing weak or no reaction to HCl alternate with lighter horizons showing strong 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 orange mottles or bands (RF in Fig. 6), are commonly observed at the mesoscale. Plant debris and freshwater gasteropods are seldom encountered. PP values are > 2 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 $CaCO_3$ 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).

4.2. Paleosol stratigraphy

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 and 6), whose cumulative thickness is in general < 5 m. Calibrated ages span from the ^{14}C detection 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 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. 3). Undecomposed organic matter was not observed and carbonate concretions have locally been reported from the base and the top of these horizons. Bk horizons are typified by high $CaCO_3$ content (30-40% of the total volume, Bruno et al., 2020b). Carbonate concretions are abundant as coalescent nodules or whitish sub-horizontal bands (evolutionary stage II of Gile et al., 1981; Machette, 1985). Transition to overlying and underlying horizons is in general gradational. Bw horizons show moderate $CaCO_3$ content (Bruno et al., 2020b; Amorosi et al., this volume) and faint mottles of Fe oxides and-or hydroxides. Evidences of clays mobilization (Bt horizons) were not observed. At distal locations, these paleosols may show less clear horizonation (A-Bw profile) or may 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 of a 3-10 m thick sedimentary unit composed of overbank muds and subordinated channel-related sands (Last Glacial Maximum Sedimentary Unit, LGM SU in Figs. 4, 5 and 6). Textural characters of paleosol PH vary from proximal to distal locations. At inland locations (core EM15, Fig. 3), paleosol 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 bands of secondary calcite (evolutionary stage II); (iii) a Bw horizon showing evidence of incipient alteration. At more distal locations (cores EM1 and EM3, Fig. 3), paleosol PH exhibit a grey (5Y 4/1, 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, 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 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).

4.3. Radiocarbon dating of paleosols

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).

Along single cores, multiple dating of A horizons yielded stratigraphically ordered ages, spanning 3-5 kyr (see cores B3, RW28 and 13, Figs. 4, 5C and 6). Only in core SPR (Fig. 5C) two samples from the base and the top of the same A horizon provided similar ages. In distal cores, plant debris from topmost A horizons yielded ages that are considerably younger (up to 5 kyr) than those from bulk-soil samples and just a few centuries older than those from overlying SM peats (see cores EM1, EM3 and B3, Figs. 3, 4 and 5). Dates from A horizons indicate aggradation at extremely low rates (0,05-0,23 mm yr-1) during soil development. Accumulation rates above or between paleosols are at least 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).

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 yielded ages clustered around three minor peaks (P1, P2 and P3, Fig. 7), at 46-41, 38-32 and 29-25 cal kyr B.P., respectively. The overall decay with time in the number of ^{14}C dates across the Pleistocene simply reflects the fact that older sediments were penetrated by a progressively lower number of cores. A decline in the number of ^{14}C dates from soil samples is also recorded towards younger ages, together with a sharp increase in dates from SM and PDFP deposits. Ages from Holocene soil samples are clustered around two peaks. The older peak reflects, in part, the youngest ages from paleosol PH and in part dates from a distinct paleosol (H1 in Figs 4, 5 and 6) observed at 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 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 was buried in large part of the Po Plain during the Middle Ages (see core CRV, Fig. 5A). Locally, this paleosol is still exposed (Fig. 5). Interestingly, the maximum number of ^{14}C dates from facies FCR is recorded between H2 and H1.

5. DISCUSSION

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

The use and reliability of radiocarbon dates from paleosols have been widely debated because, unlike plant macrofossils, which preserve the particular ^{14}C content of the moment of grow, soils are open systems where organic carbon is continuously incorporated, humified and-or oxidized to CO_2 before burial (Orlova and Panychev, 1993; Martin and Johnson, 1995; Gaudinski et al., 2000). Therefore, ^{14}C ages from paleosols reflect the mean residence time of the organic carbon plus the time of burial (Matthews, 1980; Geyh et al., 1983). The amount of organic carbon added to a soil 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 matter may lead to the redistribution of ^{14}C ratios in different soil fractions. As a result, apparently conflicting ages can be obtained from the same paleosol (Figs. 5 and 6) and from different soil fractions (Table 1). Because of this inherent variability, many authors concluded that radiocarbon 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 contamination (e.g. rootlets), paleosols should yield ages scattered within its formation period. The estimation of the length of this period, which can attain several millennia, is an intriguing goal for Quaternary paleosol stratigraphers.

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

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 et al., 2017a). More complex is the precise assessment of the ages that mark the onset and end of pedogenesis for each paleosol, even if radiocarbon dates are available from the base and top of A horizons. Previous work has demonstrated that, although some organic carbon can remain in soils for millennia (Feng, 2009), the oldest organic fraction accumulated at the onset of pedogenesis may 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 from organic-sediment samples collected from the base of the A horizon may not be representative of the onset of soil formation (Gilet-Blein et al., 1980; Alexandrovskiy and Chichagova, 1998). Several factors influence carbon turnover times in soils (Kleber, 2010; Chen et al., 2013; Pandey et al., 2014). 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 developed in cold climatic conditions, such as paleosols P3, which formed at the onset of Last Glacial Maximum, dates from the base of the A horizon should better approximate the onset of pedogenesis. This assumption is corroborated by dates from core B3 (Figs. 3, 4 and 5), where the 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 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 several soil profiles (Matthews, 1981; Matthews and Dresser, 1983; Alexandrovskiy and Chichagova, 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 than the one introduced in deeper soil horizons via roots exudation (Fontaine et al., 2007). The measured accumulation rates, incompatible with alluvial sedimentation, suggest that litter accumulation accounted for soil aggradation during exposure, with a possible contribution of wind-transported dust. Therefore, paleosols from the Po Plain represent condensation horizons, rather than depositional hiatuses. On the contrary, sedimentation between pedogenetic phases seems a quasi-instantaneous process (see ages from P2 and P3 in core B3, Figs 4 and 5). Paleosols aggradation is also testified by the presence of carbonate concretions in lower A horizons, likely related to the upward migration of the eluvial-illuvial (A-B) boundary, and by vertically stacked sub-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, 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 compound, which may yield anomalously young ages due to contamination from overlying strata (Matthews, 1980). In our case, the oldest ages are provided by humic acids (Table 1). We thus excluded contamination from overlying strata during each phase of pedogenesis.

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

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 coastline migrated in response to the stepwise sea-level fall toward its lowstand position, ~300 km away from the modern one (Pellegrini et al., 2018). Among P paleosols, only P3 developed during a phase of sea-level drop (~30–50m between 30 and 24 cal ka BP, Peltier and Fairbanks, 2006; Fig. 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 processes are controlled by the balance between sediment supply and water discharge (Blum and Törnqvist, 2000). Low sediment supply-discharge ratios result in channel incision and soil development in the adjacent interfluves, whereas high sediment supply-discharge ratios promote fluvial aggradation and paleosol burial (Demko et al., 2004). Bruno et al. (2020b) documented that soil development and burial during the Late Pleistocene was mainly controlled by changes in vegetation driven by climatic oscillations at the scale of Bond cycles (Bond et al., 1993). Soils started 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 erosion (low sediment supply). Soil development was systematically interrupted during the subsequent cold phases, at the culmination of Bond cycles, corresponding to North Atlantic Heinrich 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 to the alluvial plains. Increased sediment supply promoted the filling of fluvial incisions and the 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 fluvial-channel 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).

Above the LGM SU, PH is the most widespread paleosol in the Po Plain. This surface yielded 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 dateable material from the underlying LGM SU. However, available dates indicate that the onset of 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 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 sediment supply related to the progressive afforestation of the area (Pini et al., 2010; Vescovi et al., 2010; Campo et al., 2020), with diachroneity possibly related to particular local conditions (e.g. 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 ^{14}C dates from paleosol PH post-date the Oldest Dryas (OD in Fig. 7). The clustering of dates around the Younger Dryas in the Po delta plain (Figs. 7A and 8), suggested a possible control of this cold event on paleosol formation (Amorosi et al., 2017b, Morelli et al., 2017). The wider dataset shown in this work highlights that pedogenesis started in many sites of the Po Plain at least during the Bølling and Allerød warm phases (BA in Fig. 7A). Unlike cold phases at the 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 sedimentation. Paleosol development recorded after LGM in other alluvial (Mol, 1997; Choi, 2005; van Balen et al., 2010; Janssens et al., 2012) and eolian (Kaiser et al., 2009; Hošek et al., 2017) 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; Helmens, 2014).

Whereas Alpine rivers still flow entrenched in Pleistocene deposits, Apennine incisions were filled during the Early Holocene (see cores RN3 and EM2, Fig. 3C) and paleosol PH was buried soon after. In the coastal plain, where paleosol PH is encountered at greater depths beneath estuarine deposits, burial ages show an overall inverse power law with elevation (Fig. 9) that reflects the progressive drowning of the area during the early Holocene sea-level rise (Bruno et al., 2017). Slight dispersion of data may reflect post-burial deformation of paleosol PH by differential subsidence and neotectonics (Amorosi et al., 2021). In the area under marine influence (shades of blue in Fig. 10), burial ages decrease landwards from ~11.5 to 5.4 kyr B.P.. This overall spatial trend is not recorded in two isolated areas, where paleosol PH and older paleosols are upbended (see also Fig. 5C). In 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 redistribution along the soil profile and relatively low amounts of organic carbon in the A horizons (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 debris in the topsoil, radiocarbon dated to the early Holocene (Figs. 4 and 5), reflects the progressive rise in groundwater table (Bohacs and Suter, 1997) up to the setting of waterlogged conditions and to the accumulation of peat beds (Diessel, 1992; Richardson and Vepraskas, 2001; Stolt and 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 incisions by fine-grained and organic-matter-rich estuarine strata (facies SM, cores RN3 and EM2 in Fig. 5C). The lack of fluvial sands (facies FCR) at the base of fluvial incisions, suggests that valley filling was promoted by eustasy rather than by an upstream increase in sediment supply (Kvale and 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. (Fig. 7), corresponds to a generalized re-organization of the river network at the onset of Neoglaciation (Piovan et al., 2012; Bruno et al., 2015; Amorosi et al., 2017b), which led to the burial of paleosol H1. Slope instability and enhanced sediment production, recorded in the Northern 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 and Nicosia, 2012). The progressive return to stable floodplain conditions was paralleled by improved techniques of land reclamation and management from the late Bronze age to late Antiquity (Bruno et al., 2013).

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 ^{14}C calibrated ages within A horizons revealed that Po Plain paleosols are condensation horizons, which experienced slow aggradation during exposure (0,05-0,23 mm/yr) due to accumulation of plant litter, with a likely contribution of wind-transported dust.

Millennial-scale climate oscillations and glacio-eustasy concurred in driving the pedo-sedimentary evolution of the Po Plain during the Late Pleistocene. In particular, during the MIS 3, pedogenetic events were mainly controlled by changes in vegetation driven by climatic oscillations at the scale of Bond cycles. Paleosols developed in well-drained floodplain environments, dissected 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.

Widespread soil development (paleosol PH) occurred at the end of the Last Glacial Maximum, following glaciers retreat and afforestation of drainage basins. Burial of paleosol PH was diachronous as indicated by radiocarbon dates from vegetal remains preserved at the topsoil. At distal locations, paleosol PH was progressively drowned and 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, reflecting 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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Figures and captions

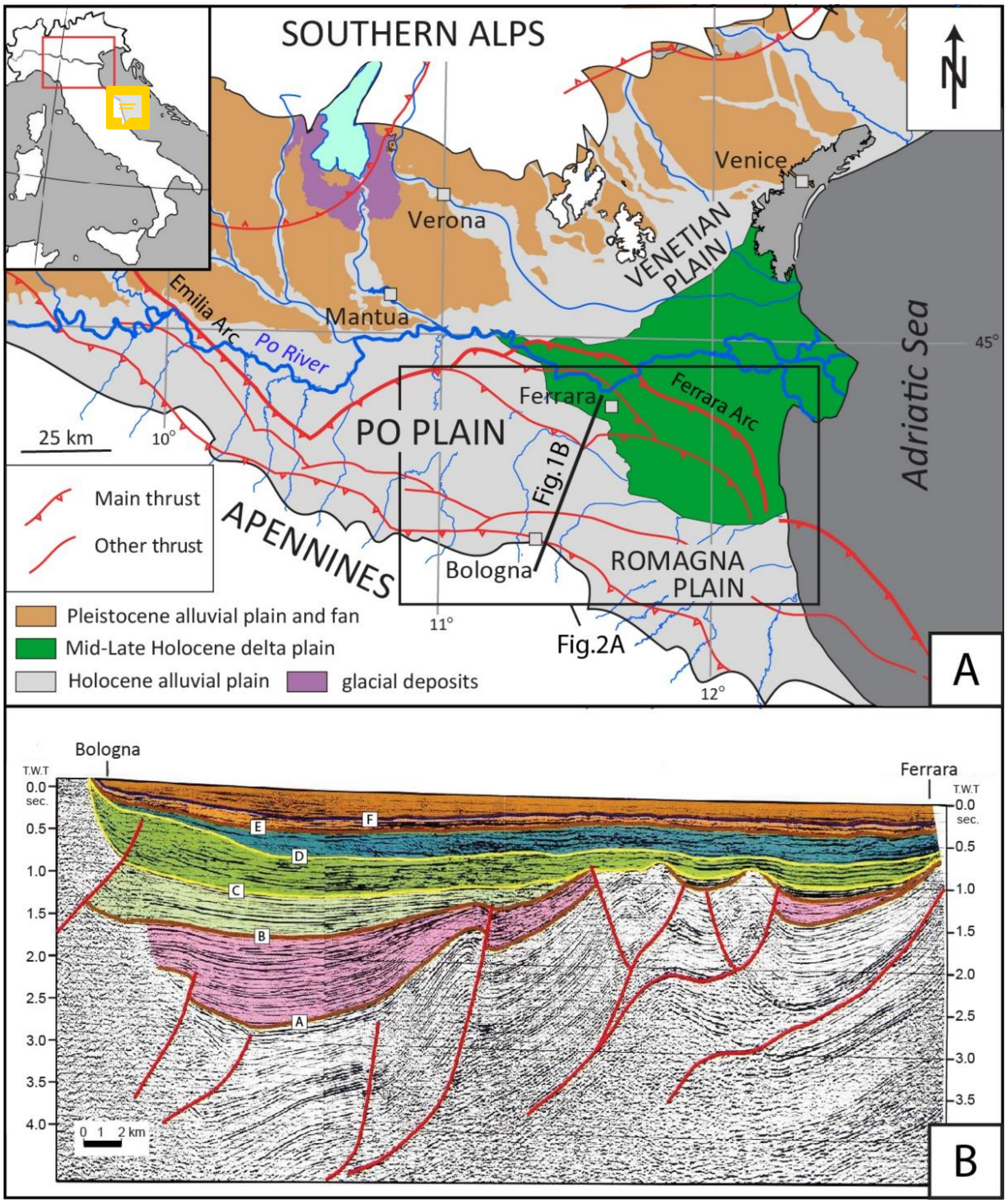


Fig. 1 – A. Geological map of the Po Plain, with location of the study area and of the seismic profile of Fig. 1B (modified after Bruno et al., 2019). The extent of Pleistocene alluvial deposits is from Fontana et al. (2014). Buried structures are modified after Burrato et al., (2003). B. Interpreted 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).

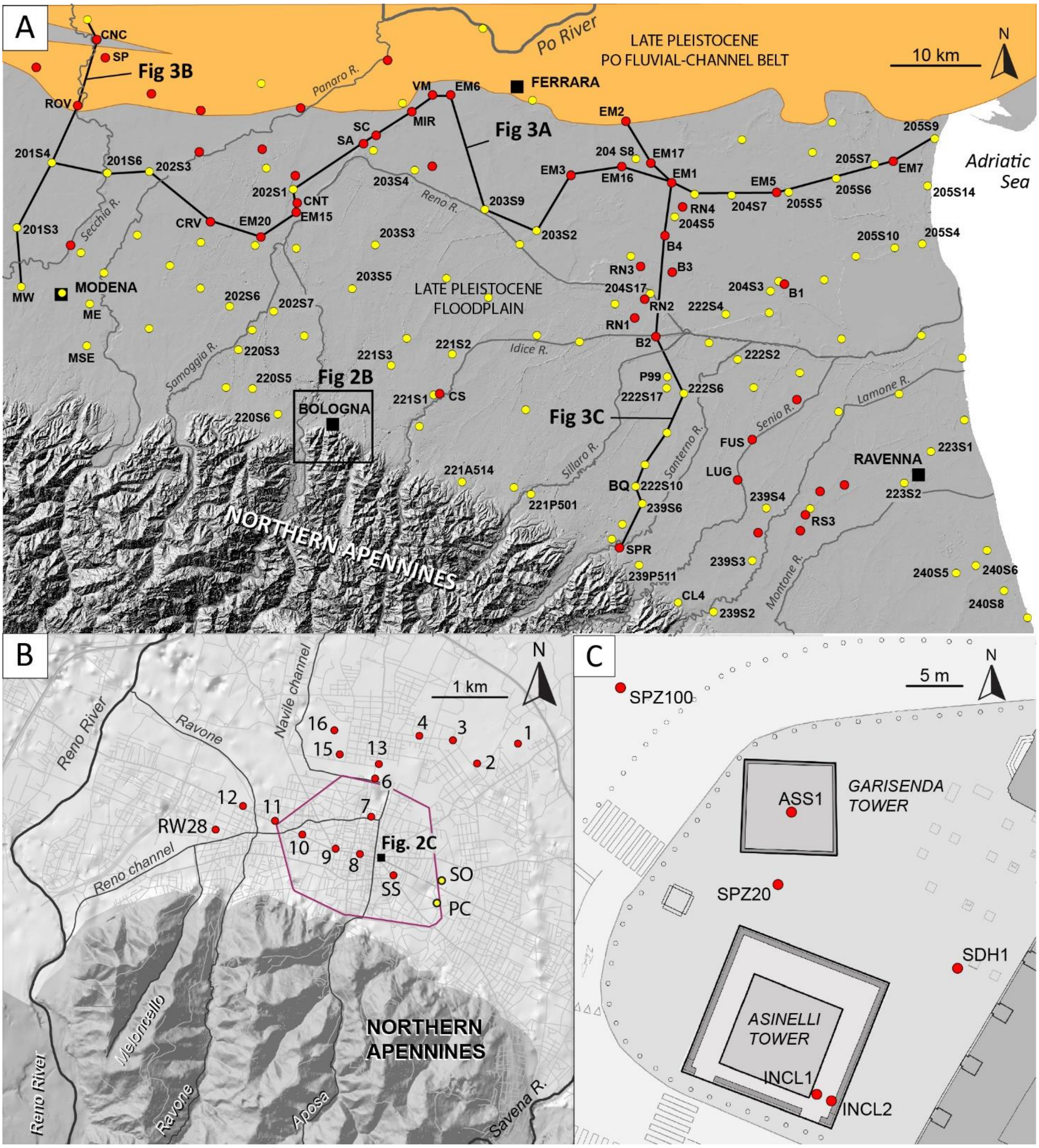


Fig. 2 – A. Study area with location of cores (red circles), core descriptions (yellow circles) and cross-sections of Fig. 5. B. Close-up on the Bologna area, with location of cores shown in Fig. 6. C. Location of cores drilled in the Two-Towers Medieval district (modified after Bruno et al., 2020b).

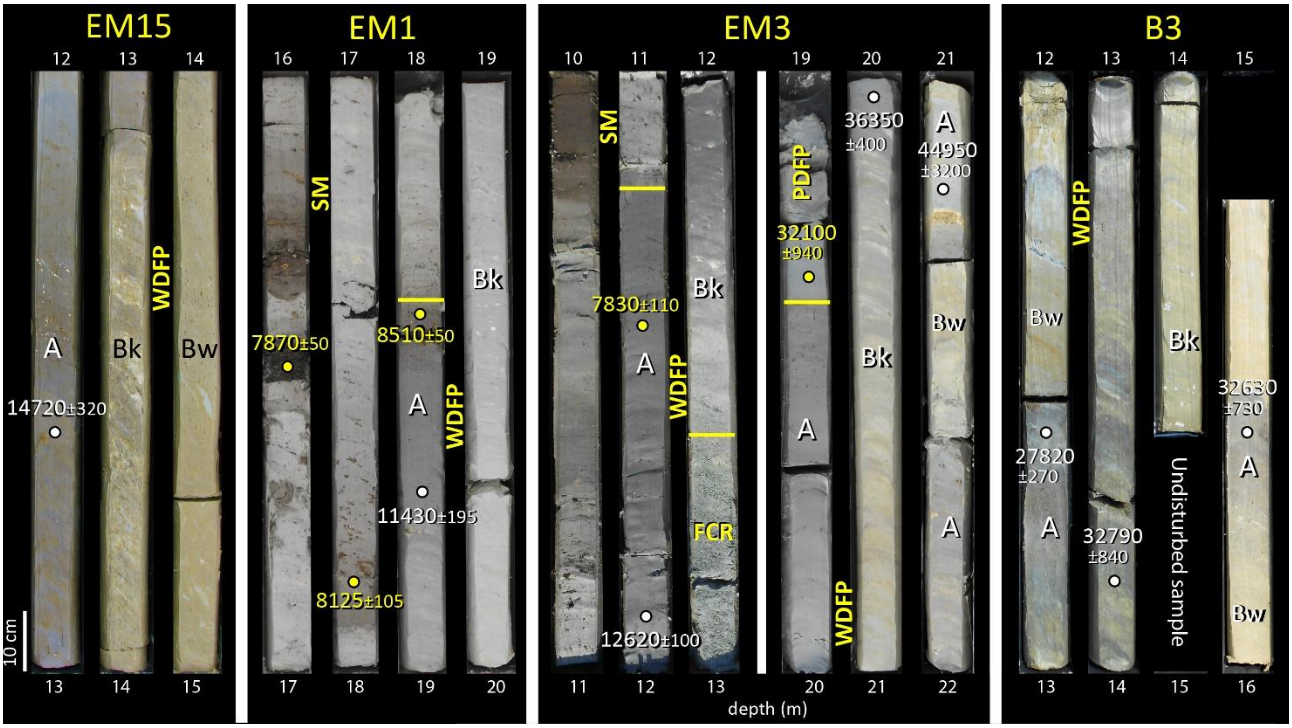


Fig. 3 – Representative photographs of paleosols from cores EM1, EM3 and B3 (location in Fig. 2). WDFP: well-drained floodplain; PDFP: poorly-drained floodplain; SW: swamp; FCR: fluvial-channel-related deposits. For pedological horizons A, Bk and Bw, see text. The white dates (cal kyr B.P.) are from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

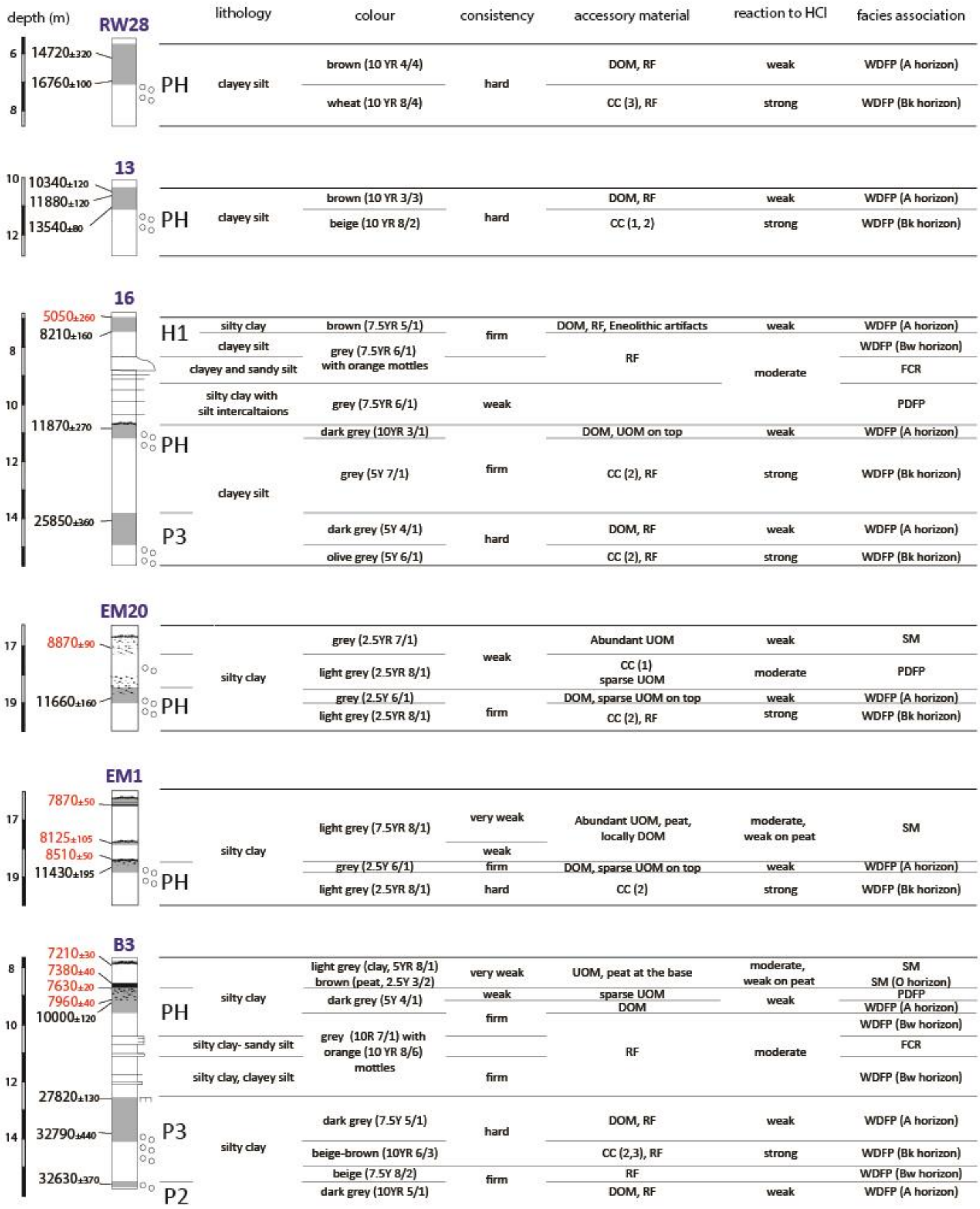


Fig. 4 – Sedimentological characters and radiocarbon ages of paleosols P2, P3, PH and H1 from cores RW28, 16, 17, EM1, EM20 and B3 (for location see Fig. 2). DOM: dissolved organic matter; UOM: undecomposed organic matter; RF: redoximorphic features; CC carbonate concretions (the number indicates the 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.

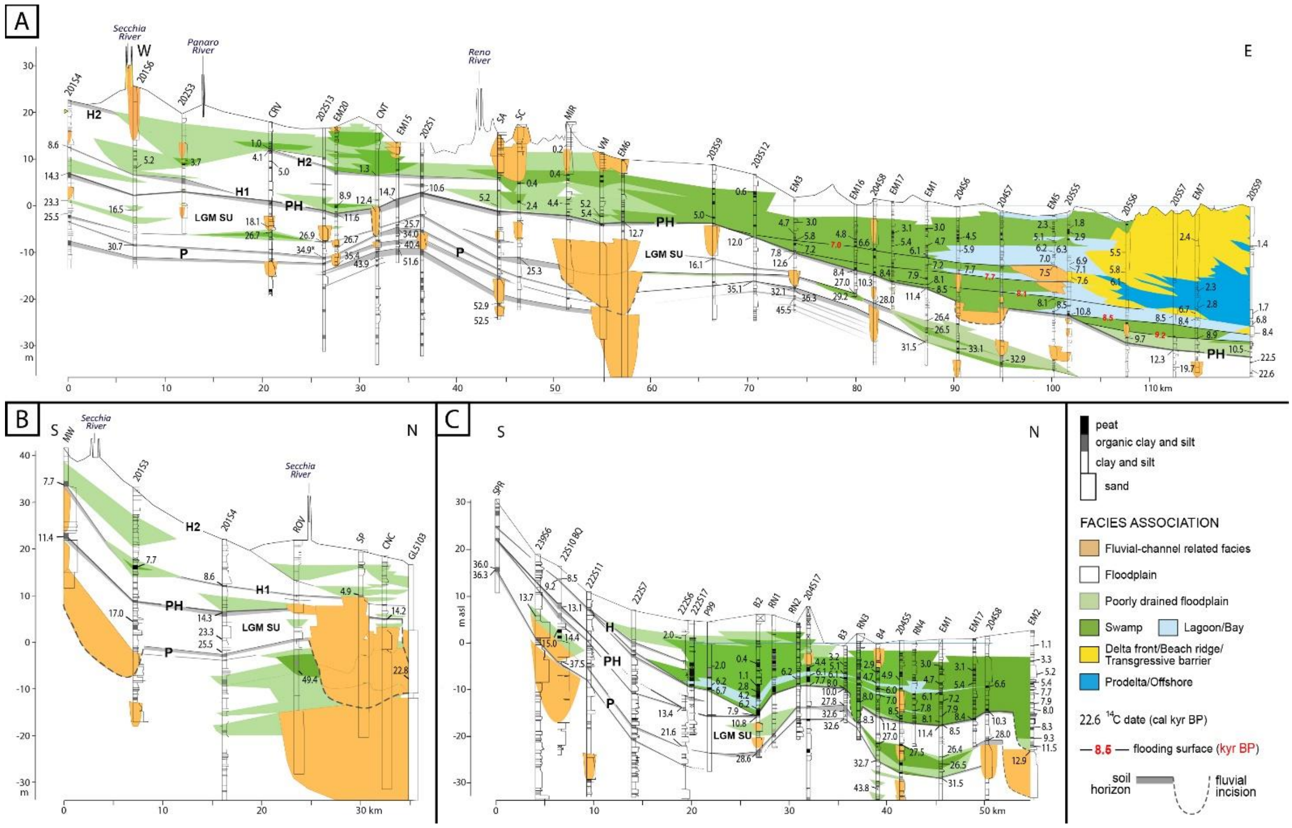


Fig. 5 – Upper Pleistocene and Holocene stratigraphy of the Po Plain depicted in cross-sections with W-E (A) and S-N (B and C) orientations (location in Fig. 2A). LGM SU: Last Glacial Maximum Stratigraphic Unit.

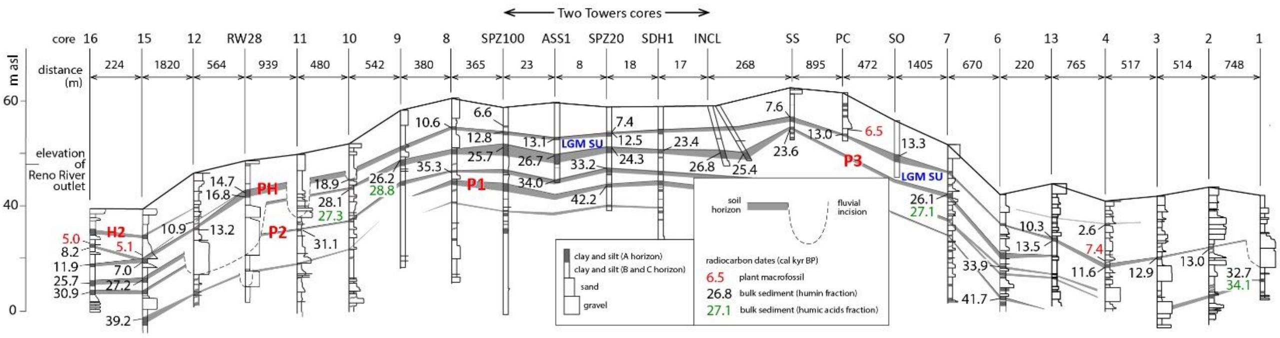


Fig. 6 – Paleosol-based correlations in the Bologna Area. LGM SU: Last Glacial Maximum Stratigraphic Unit. Location of cores in Fig 2.

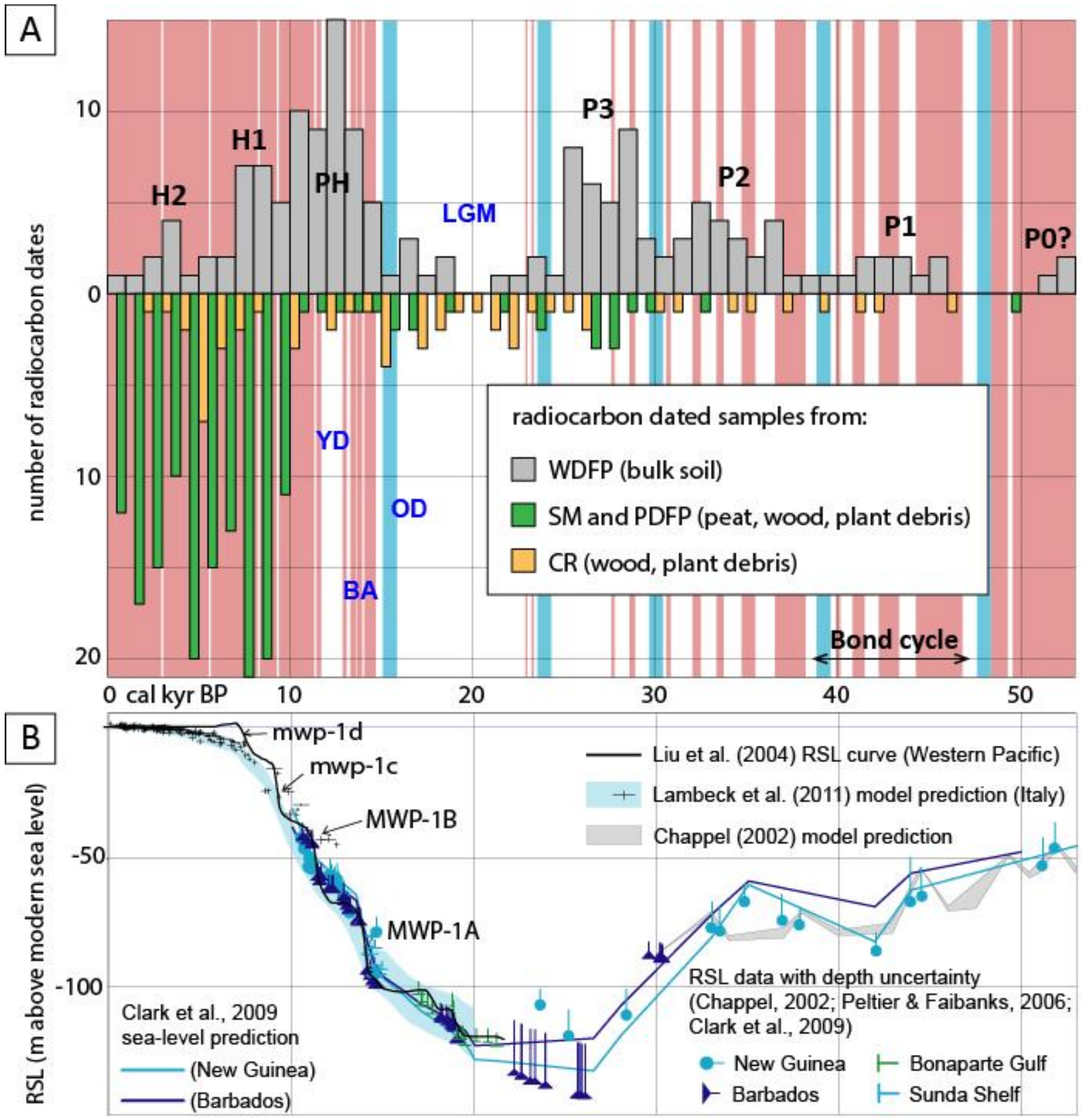


Fig. 7 – A. Distribution of radiocarbon dates from bulk-soil and plant-macrofossil samples. A major peak in dates from bulk-soil samples corresponds to paleosol PH; five minor peaks are related to paleosols P1, P2, P3, H1 and H2. Notable is the lack of radiocarbon dated soils around 20 cal ky B.P. (LGM). 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 sedimentation in the distal Po Plain and the burial of paleosol PH. Red areas correspond to interstadials, according to Bond et al., (1997) and Rasmussen et al. (2014). OD: Oldest Dryas; YD: Younger Dryas. Light blue vertical bars indicate Henrich events (Bond et al., 1992). B. Sea-level oscillations during the last 50 ky.

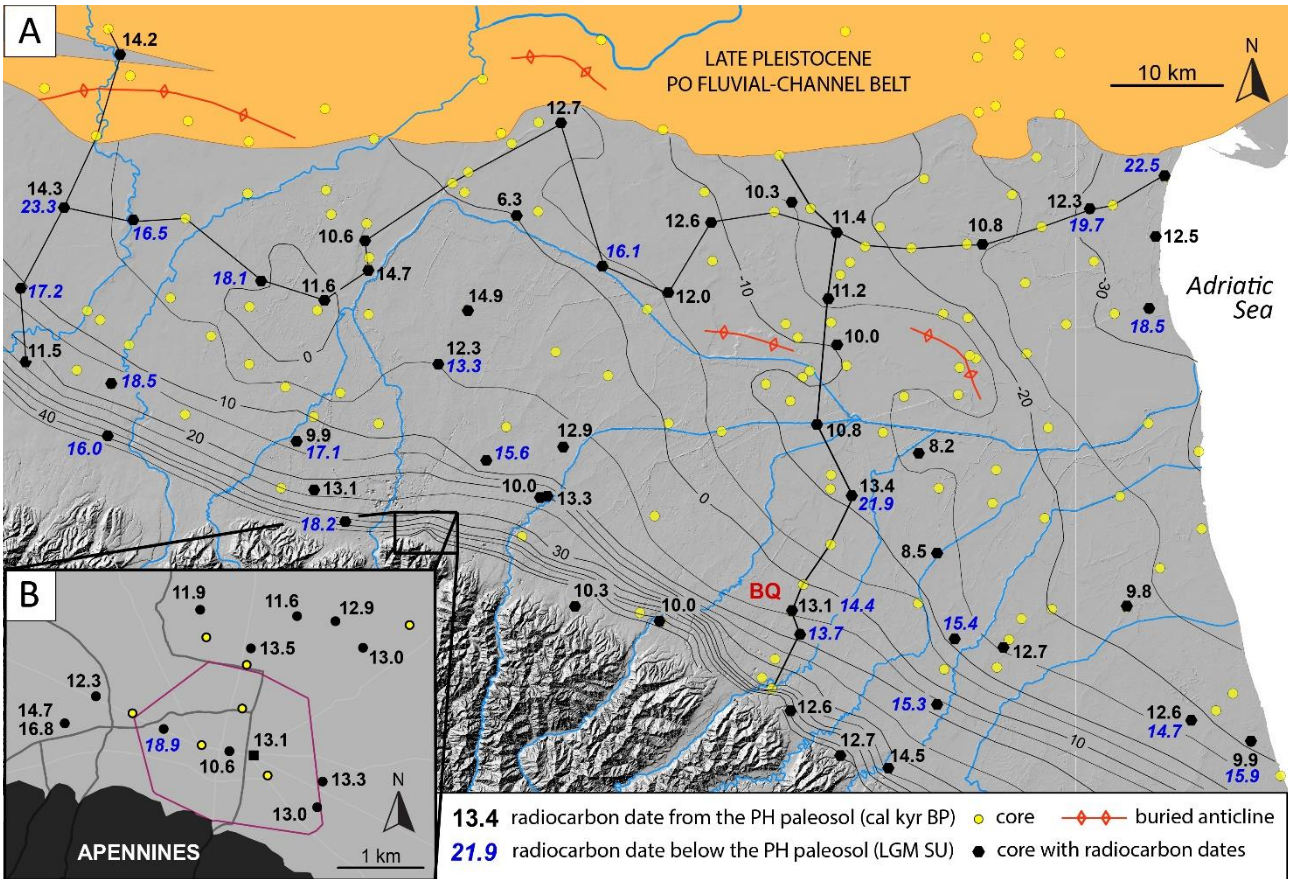


Fig. 8 – Radiocarbon dates (black numbers, cal ky B.P.) from paleosol PH (bulk soil samples) in the Po Plain (A) and in the Bologna urban area (B). Blue numbers are the youngest ages from the underlying LGM stratigraphic unit. The present-day elevation (metres above modern sea-level) of paleosol PH is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

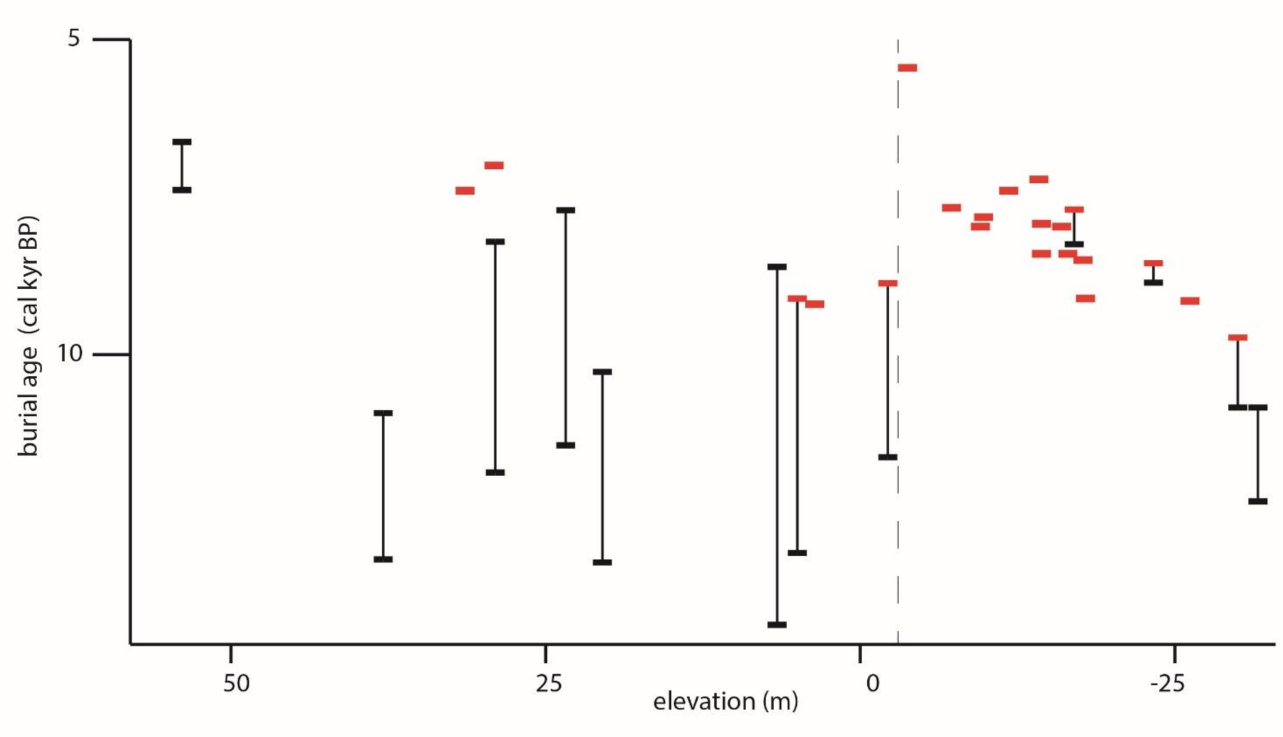


Fig. 9 – Burial ages of paleosol PH plotted against its present-day elevation (metres above modern sea-level). Black rectangles are radiocarbon dates from bulk soil samples. Red rectangles are radiocarbon dates from plant macrofossils. Burial ages from samples at elevation < - 3 m (dashed line) show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

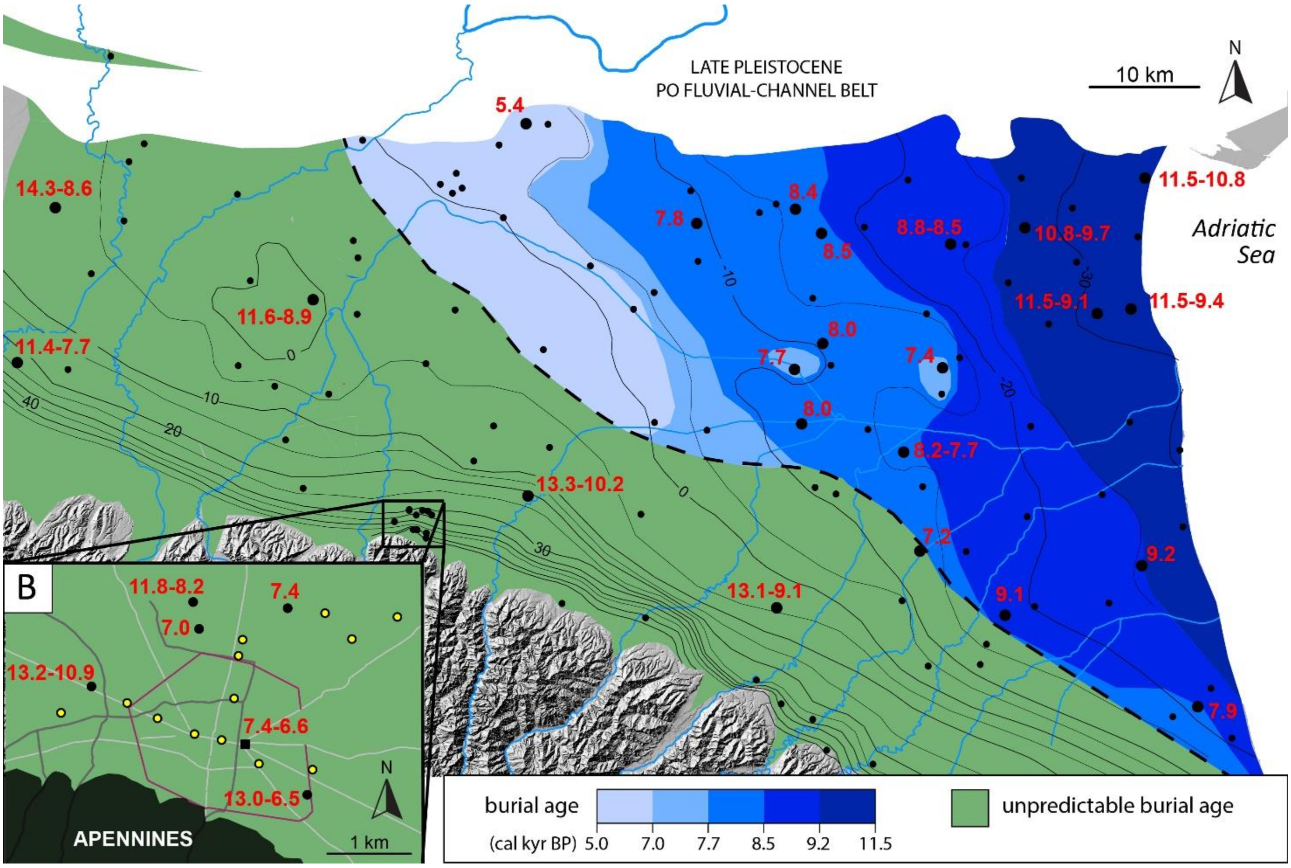
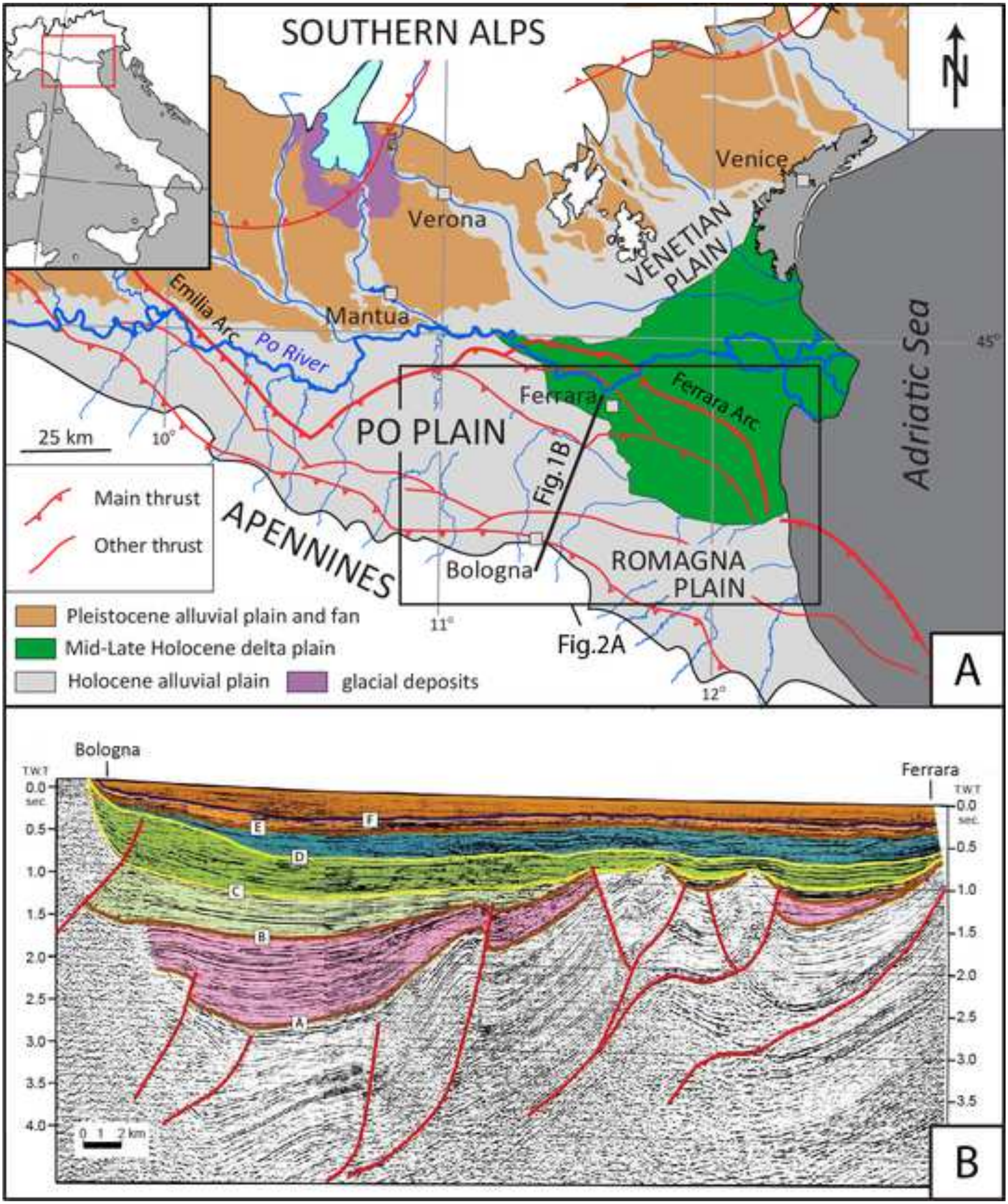


Fig. 10 – Map showing burial ages (cal kyr B.P.) of paleosol PH in the Po Plain (A) and in the Bologna urban area (B). The shades of blue depict progressively younger burial ages upland. In the green area burial ages do not show specific spatial trends. Red numbers represent burial ages or intervals 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.

Figure 1



SOUTHERN ALPS
Verona
Venice
VENETIAN PLAIN
Mantua
Emilia Arc
Po River
Ferrara
Ferrara Arc
PO PLAIN
Fig.1B
Adriatic Sea
45°
25 km
10°
11°
12°
Main thrust
Other thrust
APENNINES
Bologna
ROMAGNA PLAIN
Fig.2A
Pleistocene alluvial plain and fan
Mid-Late Holocene delta plain
Holocene alluvial plain
glacial deposits
A
Bologna
Ferrara
T.W.T
sec.
0 1 2 km
B

Figure 2

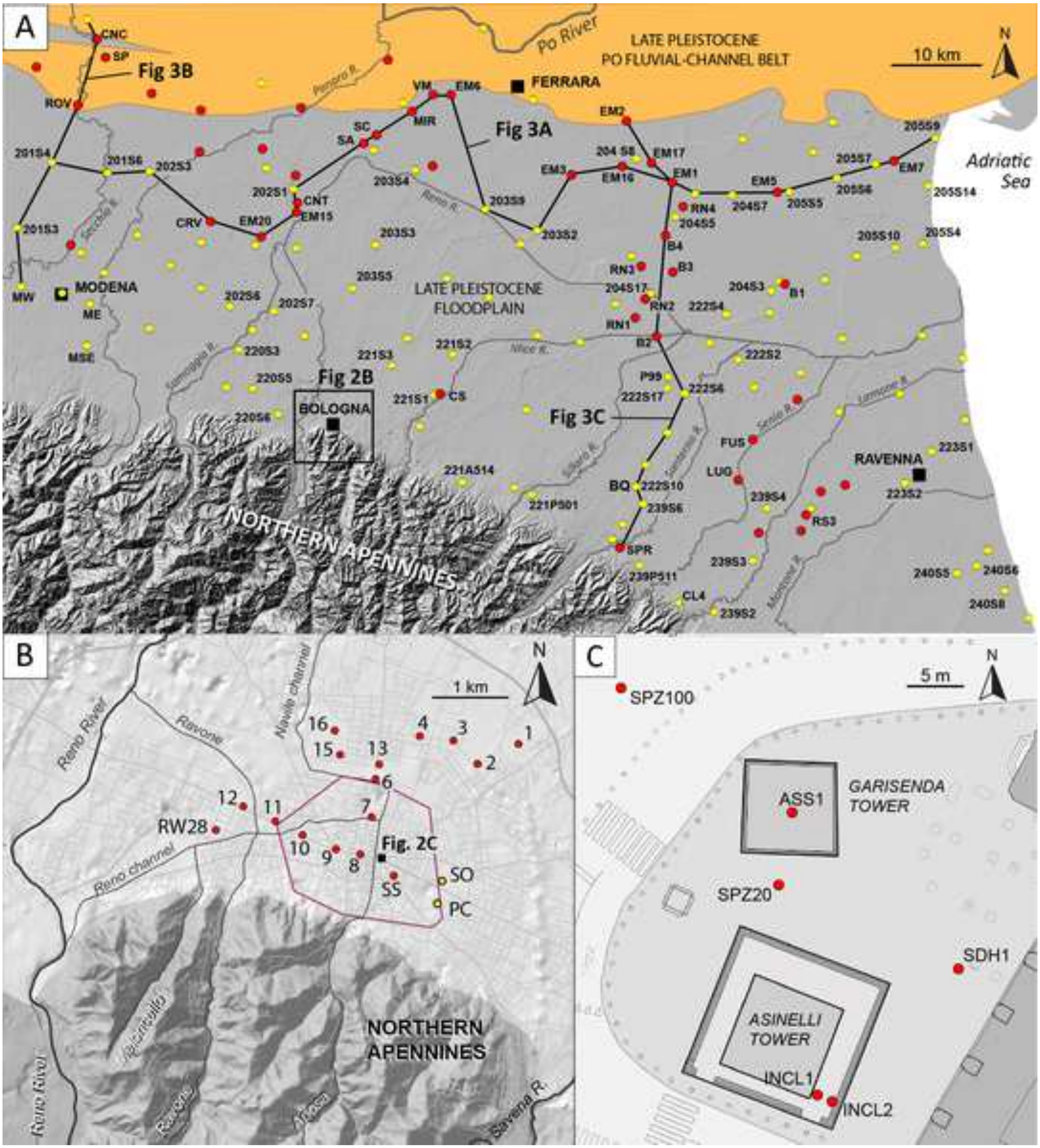
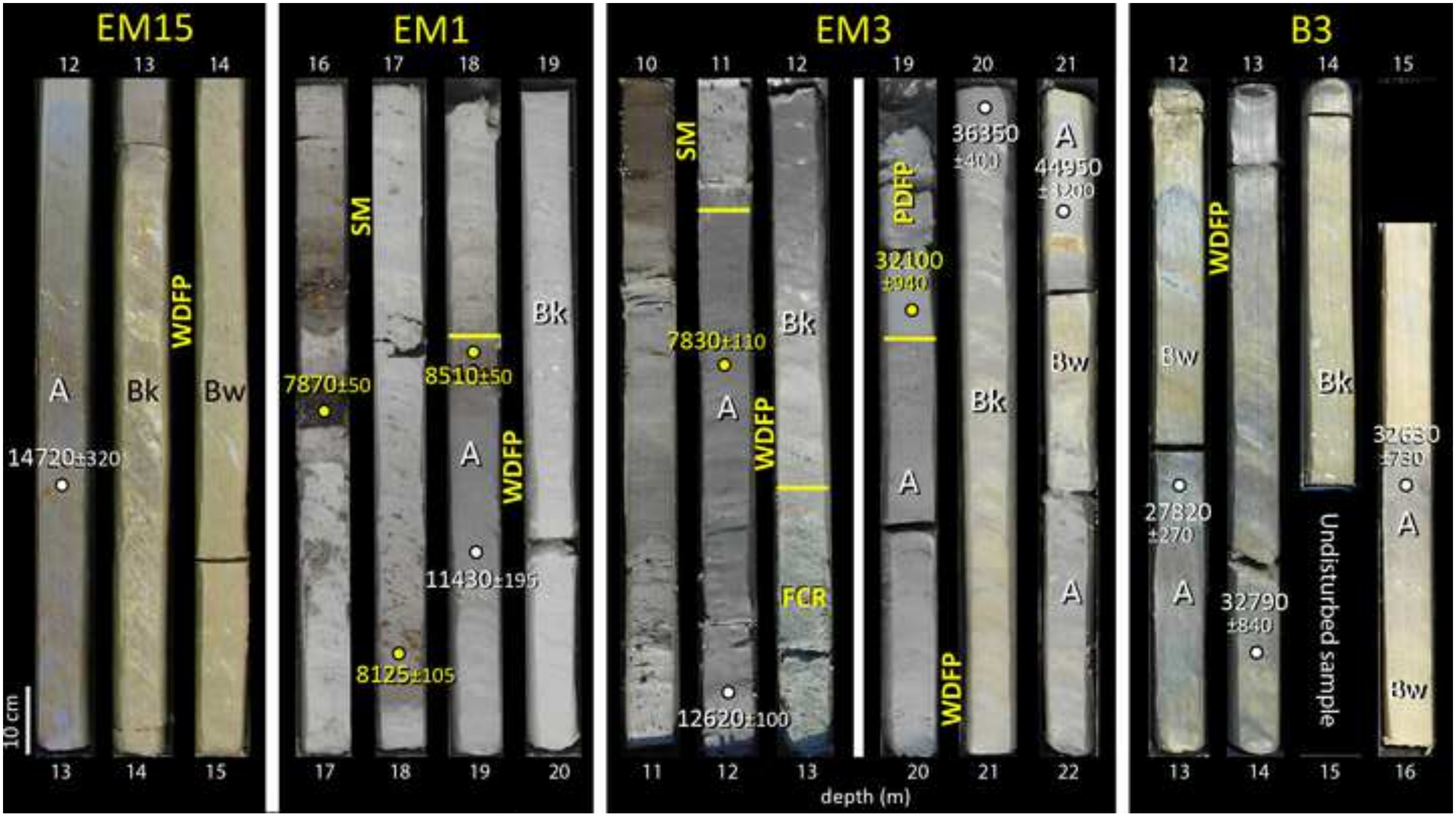

A
LATE PLEISTOCENE
PO FLUVIAL-CHANNEL BELT
Po River
10 km
N
FERRARA
MODENA
BOLOGNA
RAVENNA
Fig 3A
Fig 3B
Fig 3C
Fig 2B
LATE PLEISTOCENE
FLOODPLAIN
NORTHERN APENNINES
Adriatic
Sea
B
1 km
Reno River
Ravone
Navile channel
Reno channel
Fig. 2C
NORTHERN
APENNINES
Savena R.
C
5 m
SPZ100
GARISENDA
TOWER
ASS1
SPZ20
SDH1
ASINELLI
TOWER
INCL1
INCL2

Figure 3



EM15
12 13 14
A
14720±320
Bk
WDFP
Bw
10 cm
13 14 15
EM1
16 17 18 19
SM
7870±50
8510±50
A
WDFP
11430±195
8125±105
Bk
17 18 19 20
EM3
10 11 12 19 20 21
SM
7830±110
A
WDFP
12620±100
Bk
FCR
PDFP
32100
±940
A
WDFP
36350
±400
Bk
A
44950
±3200
Bw
A
11 12 13 20 21 22
depth (m)
B3
12 13 14 15
WDFP
Bw
27820
±270
A
32790
±840
Bk
Undisturbed sample
32630
±730
A
Bw
13 14 15 16

Figure 4

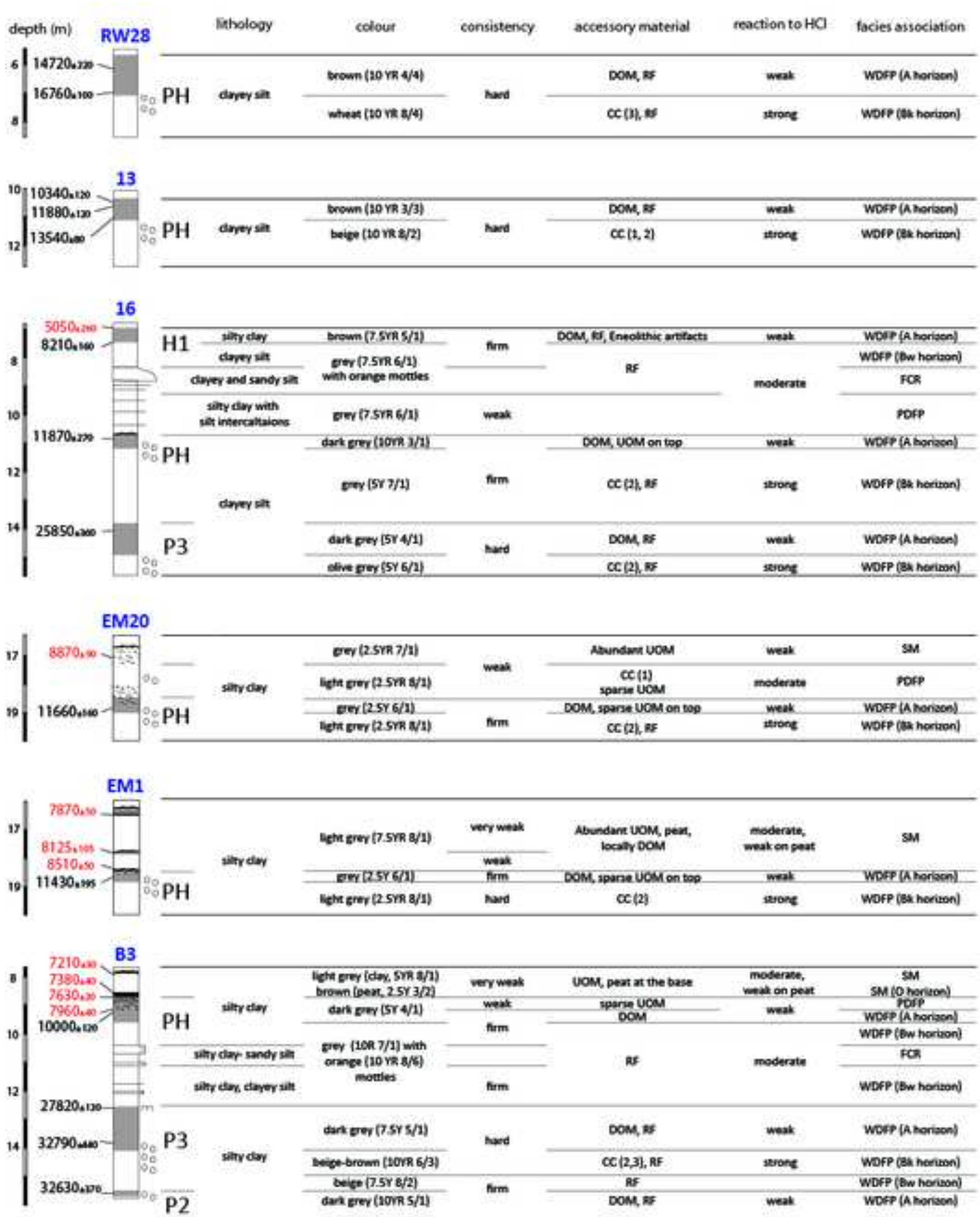
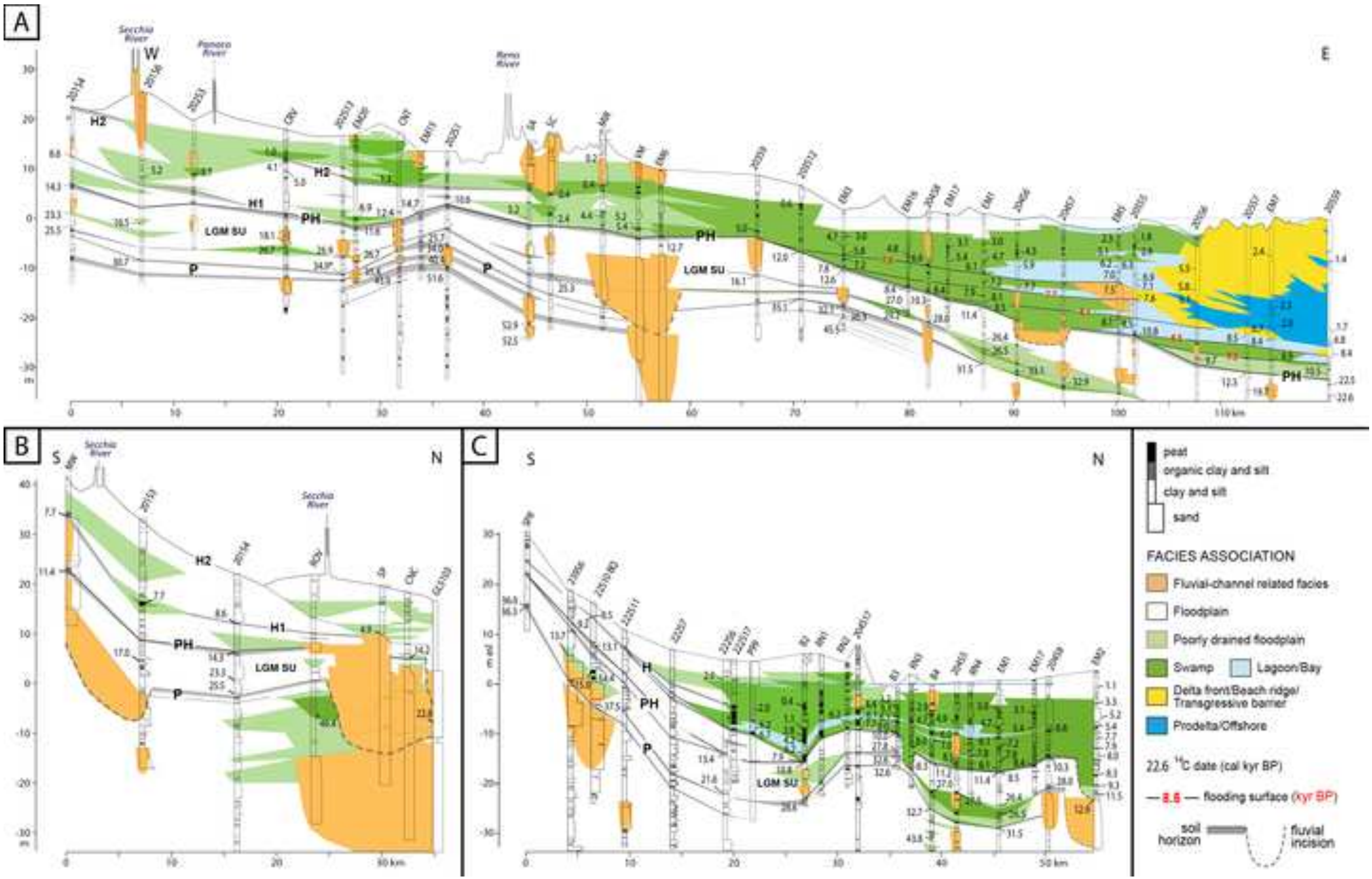

depth (m)
lithology
colour
consistency
accessory material
reaction to HCl
facies association
RW28
13
16
EM20
EM1
B3
PH
H1
P3
P2

Figure 5



A
B
C
W
E
S
N
Secchia River
Panaro River
Reno River
H2
H1
PH
P
H
LGM SU
peat
organic clay and silt
clay and silt
sand
FACIES ASSOCIATION
Fluvial-channel related facies
Floodplain
Poorly drained floodplain
Swamp
Lagoon/Bay
Delta front/Beach ridge/ Transgressive barrier
Prodelta/Offshore
22.6 ^{14}C date (cal kyr BP)
flooding surface (kyr BP)
soil horizon
fluvial incision

Figure 6

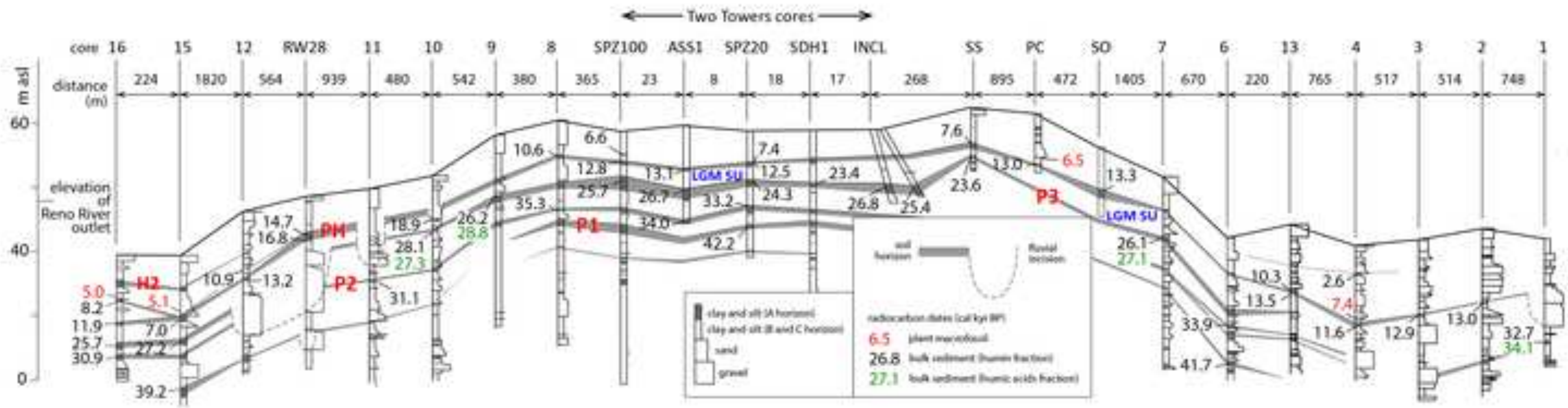
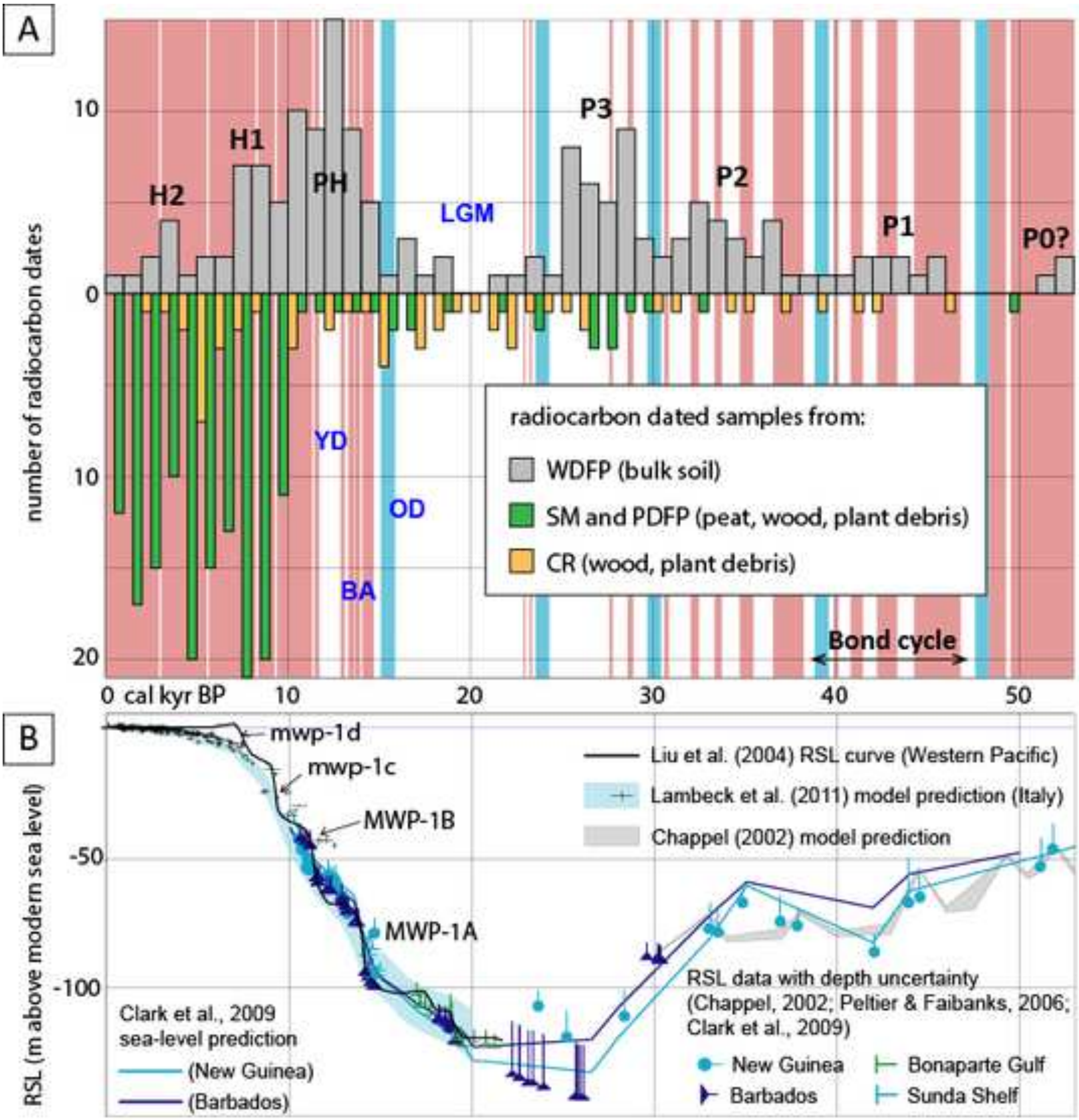
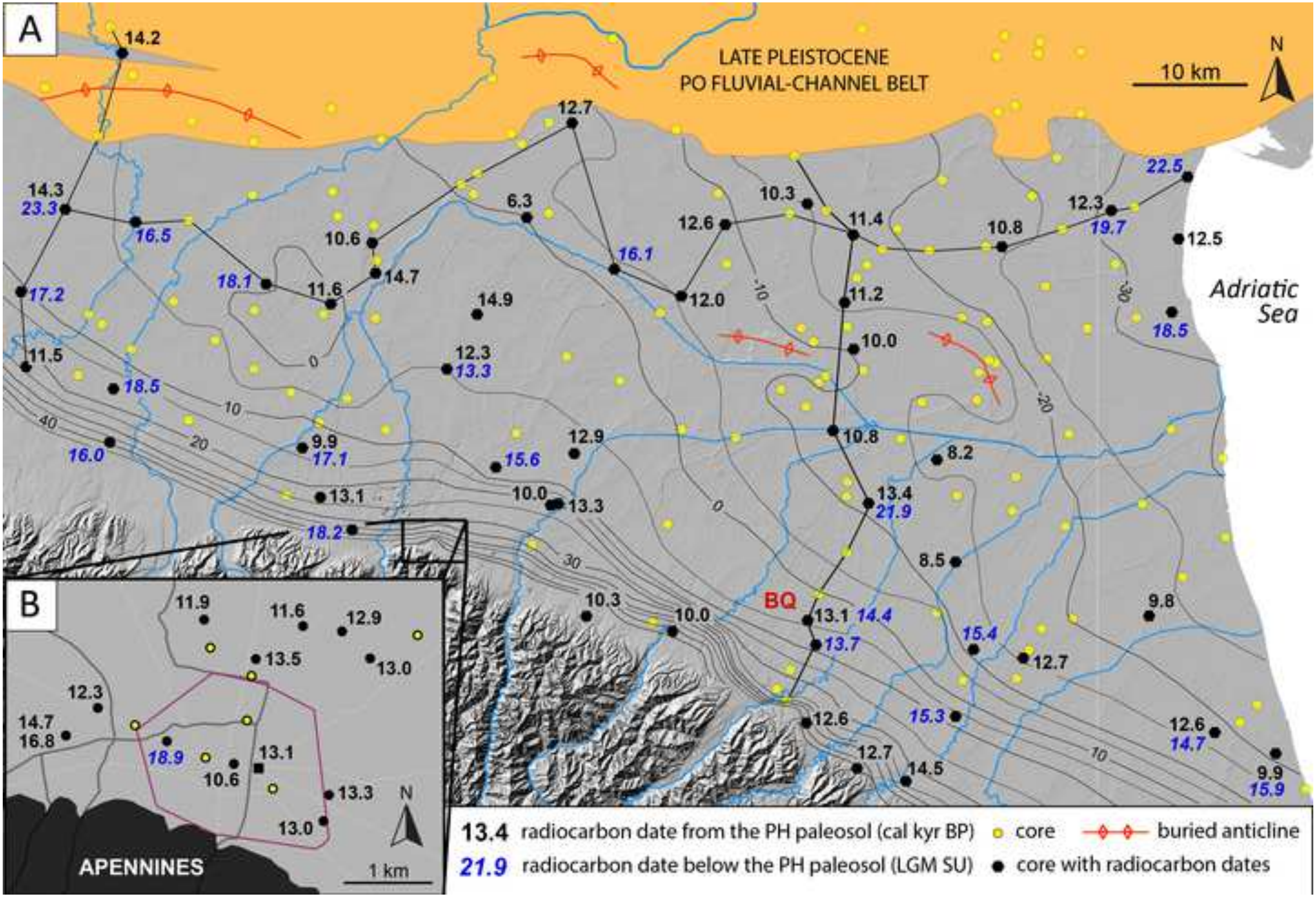

Two Towers cores
core 16 15 12 RW28 11 10 9 8 SPZ100 ASS1 SPZ20 SDH1 INCL SS PC SO 7 6 13 4 3 2 1
distance (m)
224 1820 564 939 480 542 380 365 23 8 18 17 268 895 472 1405 670 220 765 517 514 748
m asl
60
40
0
elevation of Reno River outlet
H2
PH
P2
P1
P3
LGM SU
radiocarbon dates (cal kyr BP)
6.5 plant macrofossil
26.8 bulk sediment (humin fraction)
27.1 bulk sediment (humic acids fraction)
clay and silt (A horizon)
clay and silt (B and C horizon)
sand
gravel
soil horizon
fluvial incision

Figure 7



A
number of radiocarbon dates
H2
H1
PH
LGM
P3
P2
P1
P0?
YD
OD
BA
radiocarbon dated samples from:
WDFP (bulk soil)
SM and PDFP (peat, wood, plant debris)
CR (wood, plant debris)
Bond cycle
0 cal kyr BP
10
20
30
40
50
B
RSL (m above modern sea level)
mwp-1d
mwp-1c
MWP-1B
MWP-1A
Liu et al. (2004) RSL curve (Western Pacific)
Lambeck et al. (2011) model prediction (Italy)
Chappel (2002) model prediction
-50
-100
RSL data with depth uncertainty (Chappel, 2002; Peltier & Faibanks, 2006; Clark et al., 2009)
New Guinea
Barbados
Bonaparte Gulf
Sunda Shelf
Clark et al., 2009 sea-level prediction
(New Guinea)
(Barbados)

Figure 8



A
B
LATE PLEISTOCENE
PO FLUVIAL-CHANNEL BELT
10 km
N
Adriatic
Sea
BQ
APENNINES
1 km
13.4 radiocarbon date from the PH paleosol (cal kyr BP)
21.9 radiocarbon date below the PH paleosol (LGM SU)
core
core with radiocarbon dates
buried anticline

Figure 9

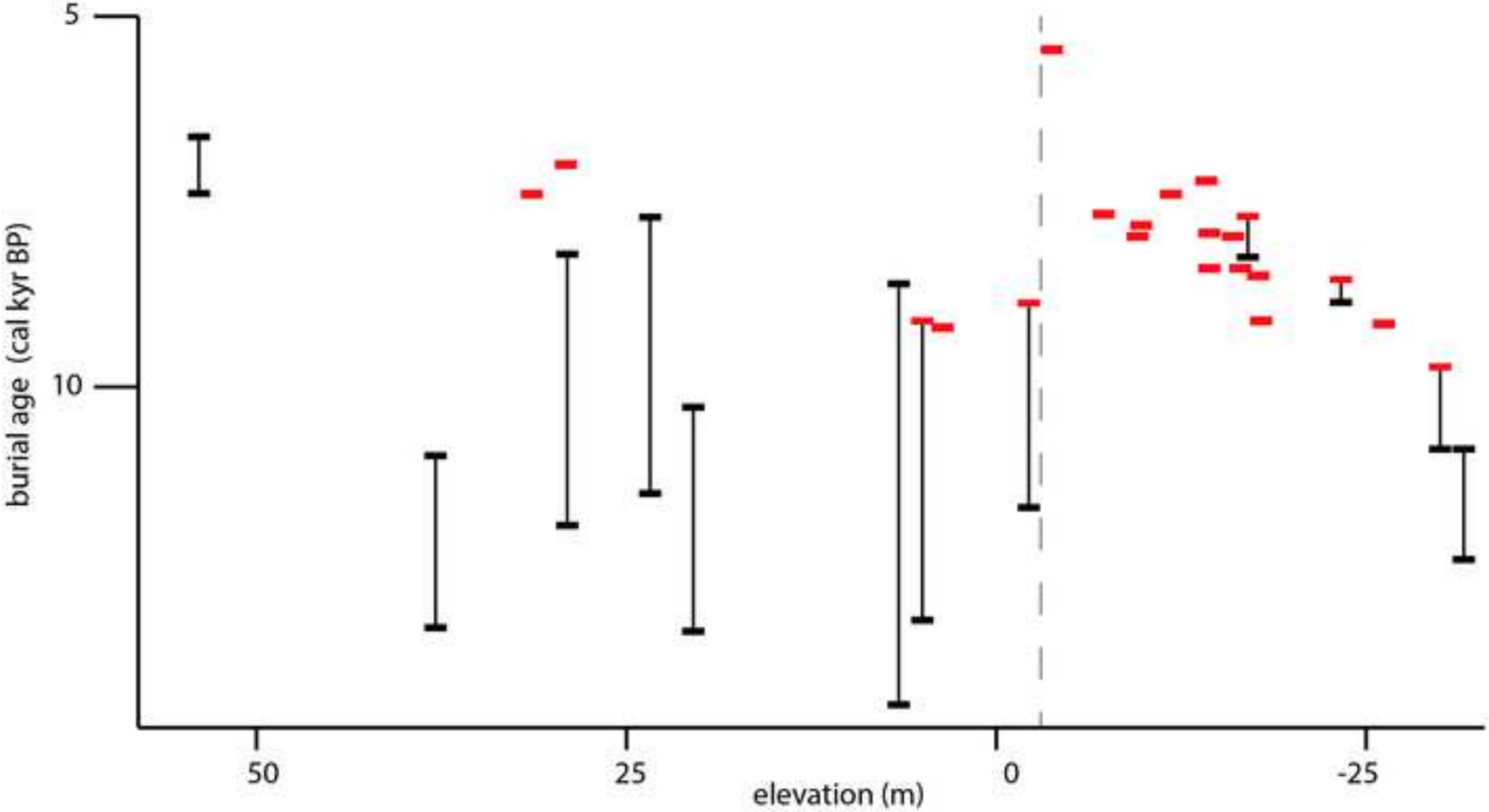
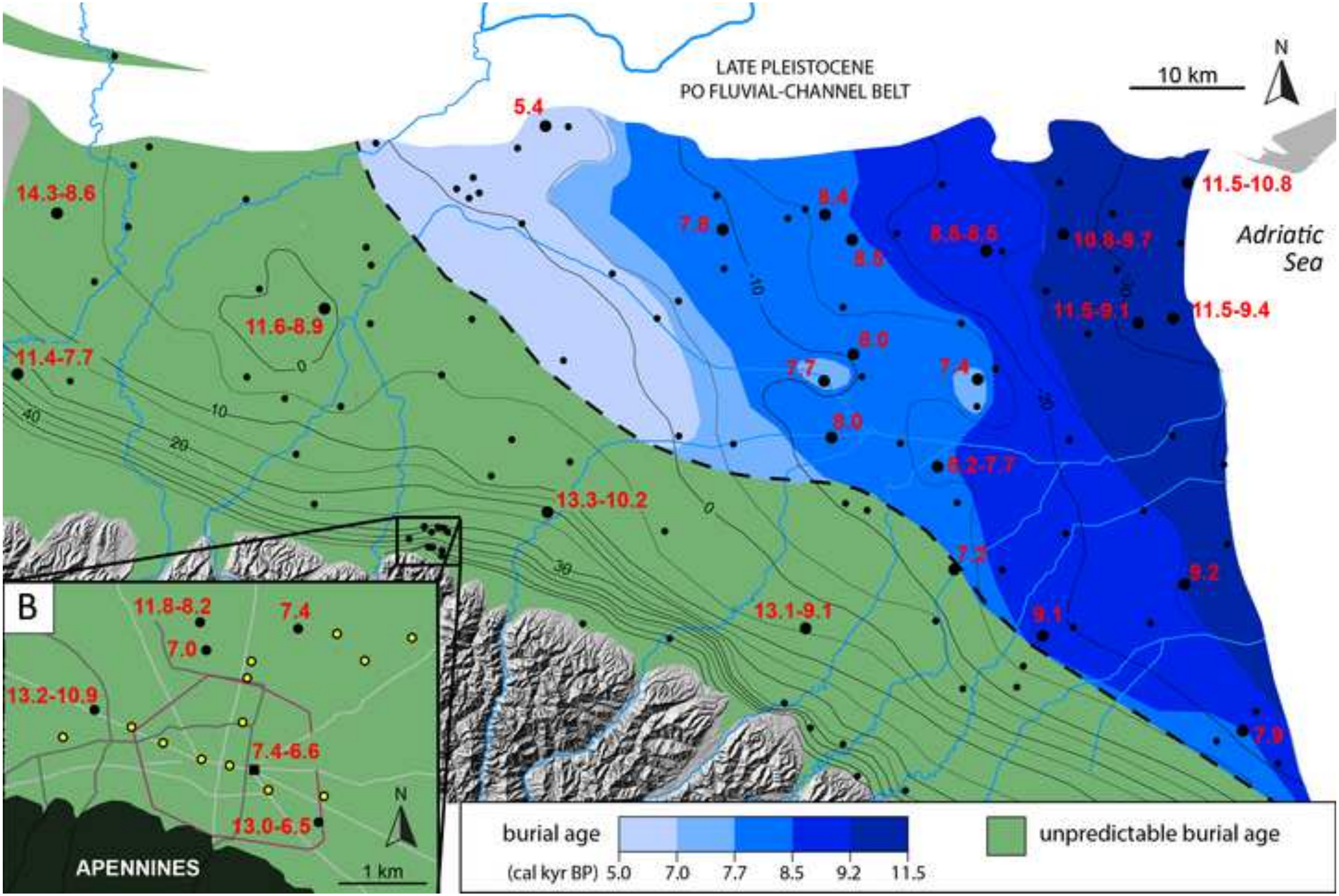


Figure 10



LATE PLEISTOCENE
PO FLUVIAL-CHANNEL BELT
10 km
N
Adriatic Sea
5.4
14.3-8.6
11.6-8.9
11.4-7.7
13.3-10.2
13.1-9.1
11.5-10.8
11.5-9.4
B
11.8-8.2
7.4
7.0
13.2-10.9
7.4-6.6
13.0-6.5
APENNINES
1 km
burial age
(cal kyr BP) 5.0 7.0 7.7 8.5 9.2 11.5
unpredictable burial age

Declaration of interests

☒ 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: