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Abstract: Valley floor sediments from the Pasinler Basin, eastern Turkey, provide evidence for Pleistocene and Holocene floodplain conditions. Three terrace surfaces are present. Evidence for tectonic processes active during the Late Neogene are widespread within the basin but do not appear to have substantially influenced the detail of the Holocene palaeoenvironmental record. Significant changes in hydrology are recorded, with more stable floodplain conditions occurring at around 9,000, 5,500 and 4,000 cal. yr. BP. Incision occurred sometime after approximately 4,000 BP, probably as a response to dual climatic and human controls. Comparisons with key sites in the Anatolian region and beyond suggest these changes are part of a regional climatic pattern, perhaps influenced by changes in the East African Monsoon. Differences in the details of the records across the region reflect the characteristics of the local environment, which, increasingly in the latter Holocene, includes human activity.

Keywords: palaeohydrology, palaeoclimate, human impacts, tectonics, Holocene

Introduction

Fluvial sediment sequences, and associated landforms, have been widely used in assessing past hydrological and broader environmental change over a range of timescales (e.g. Brown, 1997; Gibbard and Lewin, 2002). This study considers a previously unexamined sequence of channel and floodplain deposits from central-eastern Anatolia.

The Late Quaternary in the eastern Mediterranean region has been marked by sometimes-dramatic changes in conditions, induced both through natural climatic variability and, particularly in the Late Holocene, by human activity (e.g. Bruins 1990; Harrison, et al. 1996; Rosen, 1997; Karabiyikoğlu et al*.,* 1999; Landmann & Kempe 2001; Roberts et al., 2001). Despite this recent interest, there are still significant geographical gaps in the distribution of sites. This is of considerable importance as the region is located at a climatic, ecological and archaeological 'cross-roads' (cf. Roberts *et al.,* 1999a p499) between Asia and Europe.

The present study aims to identify the pattern of environmental change preserved in fluvial landforms and sediments in the ENE-trending Pasinler Basin (N39º58' E041º58'), approximately 40 km east of the city of Erzurum (Figure 1). By adopting a multiproxy approach, combining geomorphology, stratigraphy and palaeoenvironmental analyses, and by comparing the results with other studies in the region, potential local and regional controls can then be assessed. A key concern is whether the record is dominated by climatic, land use or tectonic controls.

The Pasinler Basin

The Basin originated during the Tertiary in response to localised extension associated with the activity of major strike-slip faults related to the collision of the Arabian microplate and Eurasia. Extension, and associated volcanism, continued until at least the Pliocene (Bayraktutan, 1983; Keskin, et al., 1998). During the Plio-Pleistocene, the Basin experienced a compressive tectonic regime and this continues to the present. A reflection of this is tilting and subsidence, permitting the accumulation of sediment, in places several hundred metres thick (Bayraktutan, in prep.).

Low magnitude seismicity (<M4) is frequent and the area has been affected by a number of higher magnitude, destructive events, most recently the Pasinler and Horasan earthquakes on 13th September AD1924 and 30th October AD1983 respectively (both Ms 6.8) (Ergin, et al. 1967; Bayraktutan, et al., 1986; Eyigodan, et al*.,* 1999), with a smaller event (Ms 5.9) on 3rd January AD1952. A large number of landforms, including modified fluvial features and landslides, indicate seismo-tectonic processes have been active throughout the Late Pleistocene. Basin margins in particular are characterised by back tilted surfaces associated with thrust structures (Rust, et al.,1999).

The Erzurum-Kars Plateau to the north of the Basin is largely composed of lavas and pyroclastic units (Bayraktutan, 1985; Keskin, et al., 1998) while the southern side of the basin includes Tertiary limestones. Within the basin, sedimentary strata are widespread, including some Plio-Pleistocene lacustrine deposits and extensive Pleistocene and Holocene clastic sediments that underlie much of the present valley floor.

Adjacent areas in eastern Anatolia, particularly near Mount Ararat, have experienced glaciation (Birman, 1968). No conclusive evidence of glaciation has yet been found in the mountains around the Basin, though some possibly glaciated landforms and outcrops of till-like sediment have yet to be properly investigated.

The Basin occurs near the regional drainage divide (Figure 1). To the east, separated from the basin by a low col, are the headwaters of the Euphrates, which drains to the Persian Gulf, while catchments to the north drain into the Black Sea and those to the south are part of the Tigris drainage system. The Basin itself is part of the headwaters of the River Aras, which drains east to the Caspian Sea. In historical times it has been on an important trade and human migration route and it is likely that this has been the case through much of the Holocene.

The region occurs in the transition between true dry steppe to the east and more temperate climate zones to the west, though with marked seasonality associated with continentality. Recent maximum and minimum temperature extremes, recorded at Erzurum, are 36ºC and -38ºC respectively (FMI, 2001). Annual precipitation is typically between 400-600mm, though summer months are mainly dry, with some thunderstorms, while the winters are often characterised by heavy snowfall. This seasonality is reflected in surface hydrology, with soil generally desiccated by late summer and river flows being at their highest in the spring and early summer. Late-lying and even permanent snow occurs at higher elevations.

Land use and vegetation

The natural vegetation of the basin may be woodland with steppe, subalpine and alpine communities at higher elevations (cf. Frey & Kürschner, 1989; van Zeist & Bottema, 1991), but, if so, this has been modified by several millennia of human activity. The archaeological site of Sos Höyük, an ancient *tell* in the modern village of Yiğittası, for example, provides evidence of near continuous occupation since at least the Late Chalcolithic (3,300-3,000 BC; Sagona, 1999-2000, 2000) and several other *tell* mounds are present, for example at Bulemaç.

At present, most of the basin is farmed. Arable crops are widespread with, for example, sunflowers being grown in damper sites on the valley floor while drier sites are used for wheat, with extensive irrigation systems. Steeper slopes and higher elevation areas are used for grazing. Grazing also occurs in favourable areas of the basin floor, for example the seasonally wet grassland near Bulemaç. Diverse meadow communities occur on steep slopes on the basin's northern margin. Forest vegetation is present in regions to the north and south of the basin (van Zeist et al. 1968; van Zeist & Bottema 1991) and there are stands of *Pinus sylvestris*, notably near Erzurum, immediately east of the basin (Yilmaz & Zengin, 2004). The main settlement today is Pasinler, situated in the middle of the basin, which also includes a number of villages (Figure 2).

Geomorphology

A series of geomorphological zones occur within and around the Basin, reflecting both pre-Quaternary volcanism (e.g. the Erzurum-Kars Plateau), and Quaternary tectonism and climate change. These include fault scarps, back-tilted surfaces, mass movements, alluvial fans, terraces and presently active floodplains (Collins *et al.,* in preparation).

Methodology

Valley floor landforms were mapped by a combination of analysis of two Landsat images (taken in AD1993 and AD2000), 1:30,000 air photographs, 1:25,000 topographic maps and field survey. The stratigraphy of the fluvial deposits was established by field examination of natural riverbank exposures, cut faces and coring by a hand-driven gouge.

Sediment samples collected from a shallow pit and using a 5cm gouge, located near Bulemaç (Figure 2), were analysed to provide a multiproxy suite of data:

- Loss on Ignition (LOI) to estimate organic carbon (550°C) and carbonate (850°C) (Figure 3). Although there are potential limitations with this method, particularly the precision of estimation of organic carbon for low carbon samples (Boyle, 2001), the general trends identified remain valid.

- Pollen analysis to shed light on past vegetation. Pollen was extracted from subsamples by successive treatment in 10% NaOH, sieving, 10% HCl, 60% HF, acetolysis, HF and mounting in safranin-stained glycerine jelly. Pollen frequencies proved to be low and this is reflected in typically low counts. Many of the grains identified exhibited some form of deterioration and there is probable selective preservation, leading to apparent over-representation of resistant types, such as Lactuceae. Because of the limitations of the pollen dataset, only variations in the most frequent taxa are considered (Figure 3).

- ^{210}Pb activity in the upper part of the sequence was measured by alpha spectroscopy, following the methodology used by Cundy et al*.* (1998).

- ^{14}C-AMS assays of the organic component of four samples to provide an outline chronology, calibrated using OxCal 3.5 (Table 1, Bronk Romsey, 2000). Analysis is based upon the full two standard deviation range of the calibrated dates. The samples are likely to have contained a diachronous mix of carbon of different ages due to pedogenic effects (cf. Birkeland, 1974; Elliott & Worsley, 1999), so the age estimates produced should be interpreted as 'ball-park' dates.

Drainage network

The Basin's present drainage exhibits an elongated dendritic pattern (Figure 2). An increase in confluence angle west of Pasinler reflects both the steeper gradient of this higher part of the basin floor and local tectonic uplift. The asymmetry of channel position within the basin suggests a tectonic control deflecting drainage towards the north. The central tributary (the Çaykara) is the most significant, collecting water from much of the western half of the basin. The southeastern tributary (the Inçesu) flows from the southern mountains, while the northeastern tributary (referred to here as the Porsuk stream) mainly derives its water from the northern mountains.

The Porsuk stream shows signs of significant tectonic influence. Its upper reaches are currently separated from the Çaykara by a fault-bounded ridge, though a low point at Sos Höyük suggests that localised uplift has blocked the flow and diverted it to join the series of channels originating around Porsuk. In places, it also shows a lower gradient and higher sinuosity compared to the Çaykara.

The three tributaries merge immediately east of Pasinler to form the Hasankale river. There is a marked step in the channel's long profile at Pasinler that corresponds with the confluence of an ephemeral channel draining the northern mountains through the Büyukdere valley. It is possible that the periodic increases in stream power associated with this channel may be responsible for the step, though a tectonic origin or lithological control cannot yet be excluded.

Fluvial landforms and sediments

Four main alluvial terrace surfaces have been identified to date through field survey (Figure 4):

Terrace 3 is the most extensive, covering most of the western part of the basin floor and, in the vicinity of Sos Höyük, around 6-9 m above the present channel. The gradient of this terrace is greater than that of lower, and presumably younger fluvial surfaces, meaning that it appears to pass beneath these surfaces near Pasinler. Limited riverbank exposures show this to be underlain by stacked sequences of sandy gravel (facies types Gm and Gp, Miall, 1990), typical of deposition in a braided channel. At the southern basin margins it grades into a series of alluvial fans. Natural and artificial exposures in these proximal locations show dominantly coarse, clastic sediments typical of deposition by flash floods and debris flows (Figure 4). In places these proximal deposits are disrupted by thrust faults and the terrace surfaces is warped. Locally, silt dominated sequences occur and a silt unit with occasional gravel occurs across the top of the terrace. In one locality (T3-1 on Figure 2), soft sediment deformations, including small load casts, diapirs and some fracturing of bedding, were found. Coarse clastic deposits, almost certainly rapidly deposited, overlay this and deformation is likely to have resulted from sudden loading.

A number of planar surfaces occur east of Pasinler at a variety of elevations above the Hasankale channel. Some of these may be related to Terrace 3 though extensive faulting and warping, together with coalescence with alluvial fan surfaces means that it is not yet possible to assess this.

Terrace 2, typically around 2-3m above the floodplain, is much less extensive than Terrace 3 or Terrace 1, and is more difficult to identify in the field because its bluff with Terrace 3 is degraded and no exposures of sediment have yet been observed. It mainly occurs to the south of the Çaykara stream, near Sos Höyük. No surfaces have yet been correlated with Terrace 2 north of the river in this area.

Terrace 1 occurs along both banks of the Çaykara and Inçesu stream, forming an area of roughly flat ground between the two, approximately 1-2 m above the floodplain.

West of Bulemaç, along the Inçesu, the difference in elevation between Terrace 1 and Terrace 3 reduces until it becomes impractical to differentiate the surfaces. Exposures east of Bulemaç were limited and generally disturbed. East of Pasinler, along the Hasankale, a low terrace is present which seems to be a continuation of Terrace 1. Artificial modifications of the surface near Pasinler and fault-related surface movements, however, mean that it is not yet possible to confidently confirm this and the age of the underlying deposits is not known.

There is a degree of spatial variability in the sequences of sediments preserved beneath Terrace 1, reflecting natural spatial and temporal variations in the position of channels and degrees of floodplain stability (Figure 4). Detailed field survey using near-continuous riverbank exposures and shallow boreholes, however, has allowed the major stratigraphic units to be traced laterally over much of the western basin. These are perhaps best preserved at Bulemaç (*circa* E041º36' N39º58'), where a periodically damp, shallow depression in the valley floor surface produces conditions suitable for the preservation of organic material. Figure 5a shows the sequence exposed in the bank of the Incesu river, south of Bulemaç (T1-1 on Figure 2).

The Hasankale has incised by around 3.5-4.0m since abandoning this terrace, revealing exposures of coarse gravel and sand, up 1.5m thick, overlain by up to 2m of silt, all unconformably overlying weathered bedrock (Pliocene Horasan Formation tuffs).

Palaeoenvironmental summary of the fluvial succession underlying Terrace 1

A summary of the palaeoenvironmental conditions represented by the Terrace 1 sequence is presented in Table 2.

At some point, Terrace 2 was abandoned through incision. It is not yet possible to determine how deep the incision was, though its field relationship to Terrace 1 suggests a minimum of around 3m.

The onset of deposition of the Terrace 1 sequence has not yet been dated, though fine sediments occur at Bulemaç before *circa* 9,000 cal. yr. BP (unit PWb, Figure 3). Of the twenty-four sequences recorded through Terrace 1 deposits in the western part of the basin, seventeen have fine sediment at their base, suggesting this was not just a localised event.

Stratigraphic unit PWb at Bulemaç shows no evidence of significant breaks in sedimentation. In some exposures there are occasional beds of sand, or isolated clasts, suggesting periodic input of coarser sediment. Increases in carbonate content could reflect a slight increase in sediment supply from limestone areas of the catchment. An alternative, and perhaps more probable hypothesis, given the setting, is that carbonate accumulated through subsequent pedogenic processes related to seasonal changes in groundwater and associated dissolution and reprecipitation. Although the very limited pollen data must be considered with caution, the types present suggest a largely open habitat, with Chenopodiaceae reflecting steppe communities (Figure 3)

PWc reflects a significant change in valley floor conditions, with blocky aggregates suggesting pedogenesis. This is likely to have been a diachronous event across the basin, with the onset and intensity of soil formation varying spatially due to slight differences in altitude and hydrology. The PWc palaeosol has been traced laterally for several kilometres suggesting a phase of relative stability and increased pedogenesis on the valley floor in the western part of the Pasinler Basin. An implication of this is that there was minimal channel migration during this interval, with vertical accretion of the floodplain, similar to Brown's (1997) stable bed-aggrading banks model.

At Bulemaç, the organic component of PWc yield dates to around 9000 cal. BP. The increase in organic content (Figure 3) suggests a rise in ground water levels, at least seasonally. An increase in productivity, probably in response to damper conditions, and a reduction in mineral sediment supply, suggests fewer, or lower energy, floods. Poor preservation and domination by Lactuceae suggests a significant post-depositional distortion of the pollen assemblage. The taxa which are present suggest a largely tree free environment.

Sediments overlying the early-mid Holocene soil indicate valley floor aggradation. The nature of this deposition appears to have been spatially variable, with fine-grained sediments widespread but local deposition of sand and gravel. As with older sediments, taphonomic processes have affected the pollen assemblages from Bulemaç and a detailed vegetation reconstruction is not possible. The assemblages do show a significant increase in *Pinus* (up to >50%). In the context of a floodplain deposit, where pollen may have been reworked or preferentially sorted or preserved, this could be an example of severe over-representation. The higher frequency of *Pinus*, however, occurs in a number of samples across a range of depths, and is markedly greater than in the lower parts of the core, suggesting a real increase in its abundance in the area. Horizons with blocky ped structures possibly indicate periods of stability on the valley floor. Although episodes of pedogenesis are indicated by soil aggregate structures in a number of riverbank exposures, no clear spatial continuity between them has been identified and it seems probable that they are localised episodes. In the Bulemaç core, two of these episodes are dated in the mid-Holocene, at around 5,500 and 4,000 cal BP.

Following the mid-Holocene pedogenic episodes, a significant change in fluvial behaviour took place, with evidence for both lateral and vertical erosion. This was accompanied by the deposition of coarse channel deposits and finer (sand-silt) overbank sediments. Available radiocarbon dates suggest this happened after circa

4,000 cal BP. Near the Bulemaç *tell* settlement, the channel appears to have been used as a midden during the Byzantine period (based upon broken pot fragments; a precise date on the pot has proved difficult to determine). The midden was buried by fluvial gravel at some point after this (i.e. in approximately the last 1000 years). Only background levels of ^{210}Pb activity occurred at depths greater than 3-4cm, suggesting Terrace 1 ceased to be an aggrading surface more than 150 years ago.

Floodplain. This is present along much of the Çaykara and Porsuk channels, east of Pasinler, with intermittent floodplain surfaces along the Inçesu, which mainly has its banks formed of Terrace 1 deposits.

In many places the floodplain occurs only on one side of the active channel, though there is no indication yet that there is a systematic asymmetry – rather it simply reflects the meandering of the channel. Meander cutoffs are common. East of Pasinler, the floodplain of the Hasankale channel is wider (up to 300-400m) and sinuous (sinuosity 0.4-1.4). This may be a consequence of vertical incision being hindered by resistant bedrock, exposed in the channel bed in places, causing enhanced lateral movement.

In contrast, most channels in the western sector have low sinuosity and feature gravel bedforms. A significant exception to this is immediately west of Pasinler, in the vicinity of the Bulemaç wetland, where the Çaykara and especially the Porsuk channels become highly sinuous (sinuosity ~1.9). This is clearly related to locally reduced gradients. Tectonically induced subsidence is a possible factor as the available map survey data suggests that the northernmost channel is up to ten metres lower than nearby channels to the south.

All river bank exposures of sub-floodplain deposits in the basin showed a similar general stratigraphy, with basal gravel and sand overlain by silt and sand, into which the modern soil is forming (Figures 4 and 5b). Such a sequence is typical of a laterally migrating channel, leaving a record of coarse within-channel deposits buried beneath fine overbank sediment. A number of sinuous palaeochannels on the floodplain east and west of Pasinler support this hypothesis.

Although not widespread, deformed sedimentary units have been found within the sub-floodplain deposits at a shallow depth in several places (near Sos Höyük, at Bulemaç and between the villages of Altınbaşak and Eğirmez; Figure 5). Those present near Sos Höyük occur in an area that has experienced gravel extraction, with the associated heavy vehicle movements, and they may be anthropogenic. At the other two sites, the deformations include load casts and diapirs, with folds and sometimes fracturing of bedding structures. They mainly affect sands, though some gravel has been moved. In places small clasts of fine sediment, which have been broken and moved from their original position, are present. The thickness of the deformed horizon is between 15-30cm. These structures are identical to many which have been attributed to seismically induced liquefaction (e.g. Obermeier and Pond, 1998; Stewart et al., 2001; Upadhyay, 2003). Their shallow depth, and burial by overbank deposits precludes an origin as a result of sudden loading. Similarly, instability due to localised over saturation caused by modern irrigation can be ruled out as the sites were some distance from irrigation ditches. The age of the event, or events, is difficult to determine as they reflect post-depositional, *in situ* deformation. It is likely that deformation occurred after deposition of much of the floodplain silt as this would act as a confining layer above the saturated gravels and sands (cf. Obermeier and Pond, 1998).

A piece of plastic was found lying horizontally in undisturbed floodplain silts at a depth of 36cm in section F1-1 (Figure 2), adjacent to and stratigraphically overlying deformed sediments (Figure 5b), indicating deposition of the silt confining layer in that locality within the last 50 years. This places the event responsible for deformation at this site on the Hasankale in the latter 20th Century, with the 1983, Ms 6.8 event being the most likely candidate. It is possible that the deformations at Bulemaç date to another event.

Discussion

Influence of ground movements

Tectonic processes have clearly had a significant impact on drainage in the Pasinler area. Reverse faults and thrust sheets occur to the north and south, steepening marginal slopes and also directly modifying channel networks, as in the case of the Porsuk stream at Sos Höyük, while the difference in river channel elevations, falling from south to north in the area west of Pasinler, suggests localised subsidence at the base of the northern mountain front. This subsidence has provided the accommodation space for a considerable body of alluvial sediments.

The reasons for the presence of the damp grassland at Bulemaç are unclear. It does however occur near the confluence of two river channels within an inter-channel depression. Whether this results simply from slightly greater deposition near the channels (i.e. levées) or if tectonic processes have been involved has yet to be determined. Elsewhere in the basin, with the exception of irrigated fields, the dampest areas are fed by tectonically controlled springs and seepages. What is apparent is that, while environmental conditions at this site have varied, broadly in line with the rest of the western basin, the preservation of some organic material (Figure 3) indicates that it has been damper than other parts of the valley at a number of times during the Holocene.

Soft sediment deformation in floodplain exposures appears to reflect recent seismicity. Their limited extent probably reflects the combination of factors at these particular sites: sufficient ground acceleration, water-saturated sediments with limited internal strength, and a strong fine-grained cap that prevented early release of pore water pressure.

What is surprising, given the recent frequent seismicity, is the absence of similar deformation structures in any of the pre-20th Century Holocene or earlier sediments yet examined. It is unlikely that this is a reflection of a sudden increase in the occurrence and magnitude of earthquakes.

For the Terrace 3 deposits, the lack of identified soft sediment deformation may be a reflection of only limited data. Furthermore, the typically coarse nature of the deposits and their likely rapid rate of emplacement is likely to mean that any deformation is rare and widely spaced.

The absence of deformation in any of the Terrace 1 sections studied in detail, or indeed in other exposures observed but not formally recorded, is more problematic as these sections include numerous sequences similar to the present floodplain. No conclusive answer is yet apparent. It may be that chance is involved. The two major seismic events of the last 100 years have occurred in the early Autumn when groundwater levels are rising but the fine cap of floodplain overbank deposits still maintains desiccation-induced strength. Conditions at other times of year may be less favourable for liquefaction. Vegetation may also be involved, with local variations in root depth, and hence possible stabilisation of liquefaction-susceptible layers. This may be added to by spatial and temporal variations in land use and related degrees of disturbance of the confining layer. Furthermore, deformation may be more likely in young, unconsolidated sediments near to channels and these are the most likely to be destroyed by subsequent lateral erosion. An alternative explanation would be that the historical seismic regime of the Pasinler area is different from that of the previous few thousand years but it seems unlikely that this is the case given the abundance of fresh deformation structures in the Basin.

The Pasinler terrace sequence

Terrace 3 is almost certainly a diachronous surface, though it clearly represents the end of the last major phase of aggradation in the basin. The continuous nature of the surface (i.e. no significant degradation across much of the basin) suggests it is not very old, and was perhaps abandoned by the Çaykara, Incesu and Porsuk streams as recently as the final stages of the last cold stage.

Some channels flow across Terrace 3, rather than being incised into it, suggesting either that, since it was formed, these streams have lacked the energy to erode into it or, in some areas, fluvial deposition on the Terrace 3 surface is continuing.

The succession of sedimentary units underlying Terrace 1 provides a record of fluvial and floodplain conditions, with vertical aggradation dominating for much of the early and middle Holocene. This aggradation was clearly not continuous, with at least one hiatus marked by soil formation, and at least one episode of minor erosion. Variations in the nature of deposition and surface stability appear to reflect hydrological conditions. Prior to approximately 9,000 cal. yr BP, mineral deposition dominated. The abundance of fine sediments, represented by PWb, suggests extensive erosion of soils. This accords with the limited pollen evidence for a largely tree-free environment.

The organic soil formation episode at Bulemaç, dated to around 9,000 cal. yr. BP, corresponds stratigraphically with pedogenesis (and by inference some degree of stability) recorded in sub-Terrace 1 sediment sequences adjacent to both the Çaykara and Incesu channels. This suggests a change in hydrological conditions affecting at least the southern and western parts of the basin; perhaps less variable flows, with lower power floods.

Sediments overlying the buried soil are spatially and vertically variable, ranging from silt to gravel. This suggests a further shift in basin conditions in the mid- and late-Holocene, with an increase in sediment supply and stream power, resulting in the preservation of both within-channel and overbank deposits.

After deposition of PWg, fluvial behaviour clearly underwent a significant change, with energy within the stream exceeding the critical stream power threshold (cf. Bull, 1979), leading to incision. Such a change implies an increase in peak flow and/or a decrease in sediment supply at some point during approximately the last 1,000 years (based upon the Byzantine pot found at Bulemaç) and continuing to the present.

Four forcing mechanisms may be involved in the change since the Byzantine Period:

a) a shift to drier climatic conditions, with less vegetation and a more variable, flashy hydraulic regime;

b) variations in sediment supply as a result of seismically-triggered landslides;

b) land use change resulting in changes in catchment soil characteristics and runoff;

c) tectonic processes effectively 'rejuvenating' the channel in this part of the valley floor.

In reality it is likely that all four were involved. The Lake Van record suggests drier conditions in the last two thousand years (Wick et al., 2003). Bruins (1990) reports some changes in erosion and runoff at the time of the Arab Conquest in the Levant. Perhaps, something similar occurred in the Pasinler area at the end of the Byzantine period though Sagona (pers. comm., 2000) identifies significant prehistoric cultural change in the area which does not seem to have a strongly expressed fluvial signature. Finally, there is a ten metre difference in elevation between the southeastern tributary channel and that of the north eastern channel, which is less than three kilometres away, probably reflecting ongoing tectonic movements, though whether this would have a major impact on the fluvial system over a timescale of a few centuries is uncertain.

The pollen record from Bulemaç must be interpreted with considerable caution, as selective taphonomic processes prior to burial and differential preservation are likely to have significantly affected the assemblages. There are also gaps in the record from depths where very little or no pollen was found. Some general conclusions can be made however. The lowermost samples (units PWb and PWc) contain low frequencies of arboreal pollen, with herbaceous taxa dominant. This suggests only limited woodland in the catchment area with much of the area covered by a variety of grasslands. These probably included steppe-type communities (based upon the presence of Chenopodiaceae), in better-drained sites and damper grassland near channels and springs. The significant relative expansion of *Pinus* above 134 cm depth may indicate a climatic shift, perhaps less severe winters. Given the archaeological evidence for occupation in the basin (Sagona, 1999-2000, 2000), the decline in *Pinus* after circa

5,500 cal. yr BP is likely to be a reflection of increased human activity in the basin, though climatic change could also be involved.

Regional comparisons

Limitations of the floodplain record - Fluvial deposits are notoriously hard to interpret, even within a single catchment, with some sequences producing records that are unlike any others nearby due to spatial variability in fluvial processes. This is particularly the case in sites dominated by within-channel deposits. Floodplain sequences formed by vertical accretion of sediment delivered by overbank flow provide a better opportunity for considering past catchment changes. Caveats must be invoked, however, regarding, for example, changes in the balance between erosion and deposition along a river's long profile; a river may be actively destroying its floodplain in one area and accreting in another. The Pasinler sequence, identified from multiple, near continuous exposures extending over several kilometres and two cores, does exhibit a number of spatially consistent features which suggest catchment-scale controls. Consequently comparison can be made with sequences elsewhere.

Inter-site comparisons - The Pasinler wetland record outlined above reveals a broadly similar sequence of conditions to those found at other Holocene sites in the Anatolian region (van Zeist & Bottema, 1991; Wilkinson, 1999; Fontugne et al., 1999; Roberts et al*.,* 1999b; Wick et al., 2003) (Figure 7).

During the Quaternary cold stages, the Eastern Mediterranean region clearly experienced cooler temperatures than today (e.g. Birman, 1968; Roberts, 1983). At higher elevations, periglacial slope processes and seasonal melting may have occurred, leading to rapid runoff and erosion, reflected in coarse slope deposits studied elsewhere (e.g. Nemec and Kazanci, 1999) and the clastic deposits of Pasinler Terrace 3. As with other parts of the world, cold stage climates are likely to have oscillated, leading to repeated destabilisation of slope-channel systems, maintaining high fluvial sediment loads.

It is not yet possible to fully assess the impact of the start of the Holocene on river behaviour in the Pasinler Basin, though evidence from Lake Van (Landmann et al*.,* 1996) suggests a rapid change in sediment deposition there (Figure 6). This may relate to the extensive fine deposits found at the base of the Pasinler Terrace 1 sequence.

At around 9000 cal. yr. BP more humid conditions appear to have been prevalent, with marshes forming in central Anatolia (Fontugne *et al.,* 1999), damp floodplain environments in eastern Anatolia (this study) and possibly a lake present in the Amuq basin of southern Anatolia (Wilkinson, 1999). This evidence, coupled with $\delta^{18}O$ and $\delta^{13}C$ records from speleothems in Israel (Bar-Matthews & Ayalon, 1997) dated to between 10,000-7,000 cal. yr BP, an increase in some arboreal taxa at Lake Van after c. 10,460 cal BP (Wick et al., 2003), and probably also Sapropel 1 in the eastern Mediterranean (Rossignol-Strick, 1999), suggest that conditions in the Near East as a whole at this time were more humid than for most of the rest of the Holocene.

This was followed by regionally extensive dry conditions at around 8,000 cal. yr BP, recorded in the Konya Plain, Dead Sea and Lake Van. Variable river flow occurred in the Euphrates valley (Wilkinson, 1999), while a period of limited erosion followed by the onset of mineral aggradation at Pasinler may also reflect this.

A second phase of more humid conditions existed in the interval between 7,000 and 5,000 cal. yr BP, though available radiocarbon dates suggest that the onset of marsh formation in central Anatolia slightly ante-dated damper valley floor conditions in Pasinler and Hatay and stable perennial flow in the Euphrates (Rosen, 1997; Wilkinson, 1999). Whether this is an artefact of the limitations of ^{14}C dating in these contexts or represents a real difference is unclear at present and requires further investigation. Using the $\delta^{18}O$-derived proxy humidity record from Lake Van in southeastern Anatolia, Lemke and Sturm (1997) suggest that the period between 7,000 and 5,000 cal. yr BP featured some of the most humid climatic conditions in eastern Anatolia of at least the last 8000 years.

After 5,000 cal. yr BP dry conditions prevailed in the Euphrates valley, while mineral aggradation at Pasinler (sub units PWf/g) suggests a reversion to more variable flow. Rosen (1997) reports a similar type of erosive phase at the end of the 3rd Millennium BC in southern Anatolia. In contrast, Fontugne et al. (1999) propose a period of marsh development in central Anatolia at this time - i.e. locally humid conditions, though the authors suggest this could reflect land use and water management rather than a shift to a wetter climate.

The Dead Sea has experienced low levels, albeit with a number of oscillations, since around 4,000 BP while the Lake Van proxy humidity curve suggests dry conditions between 3,000 cal. yr. BP and approximately 1500 BP, with conditions broadly similar to present over the last 1500 years. Wilkinson (1999) suggests the Euphrates in southern Anatolia and northwestern Syria has been characterised by a braided-meandering planform for the last 1000 years. In contrast, a higher water level has been proposed for Lake Zeribar, in northwestern Iran, though this may have been a result of alluvial fan growth that more effectively dammed the lake (Megard, 1967). From the western Pasinler basin there is evidence for a change in flow regime at the wetland site some time during the last 1000 years, leading to incision and then adoption of braiding and meandering planforms in different locations.

The evidence for broadly synchronous sequences of events during the early and middle Holocene at widely spaced sites across much of the Near East in different morphological and tectonic settings, and identified using different proxy datasets, indicates regional scale climatic oscillations, with periods when moist air masses were pushed further north and east than at present. Direct impact from a migration of the East African Monsoon may not have occurred, though such a migration would have had knock-on effects in the Mediterranean (Harrison et al., 1996).

Differences between site records across the Eastern Mediterranean in the later Holocene may reflect intra-regional climatic variations. The abandonment of Terrace 1 in Pasinler, for example, is currently difficult to correlate with a regional signal, with perhaps the exception of the fall in the level of Lake Van. This may be due to a lack of evidence for this time period, though it seems likely that the incision may be a local effect associated with catchment scale changes in land use, for example water harvesting and irrigation. It seems probable that geographical differences in the increase of population density and land use across the Eastern Mediterranean produced spatial variations in how the landscape as a whole, and floodplain-channel systems in particular, behaved (cf. Robinson & Lambrick, 1984).

Conclusions

A record of Late Quaternary environmental change is recorded in the fluvial landforms and deposits of the Pasinler Basin. The record is not continuous and a number of terraces, unconformities and at least one palaeosol suggest deposition occurred in pulses in response to climatic and land use change and possibly seismo-tectonic movements.

Although the data available are limited, the palaeoenvironmental record from the Pasinler Basin, eastern Anatolia demonstrates the potential use of floodplain sedimentary successions for assessing palaeoenvironmental variations in regions where lakes are rare or absent, or where the chronological and palaeoenvironmental reliability of lacustrine sequences needs validating. Furthermore, the sensitivity of fluvial regimes potentially makes them useful recorders of natural and human-induced change, even given some acknowledged limitations associated with dating and reworking of material.

The environmental and climatic history provided by floodplain sediments is clearly of some significance to understanding the pattern of settlement and land use in Anatolia. The humid phase at around 9,000 cal. yr BP coincides with, and perhaps facilitated, the development of Neolithic settlements in the region, while incision after the Byzantine period (circa in the last 1,000 years) will have exacerbated the impact of the lowering of water tables caused by a shift to drier conditions (cf. Rosen, 1997; Leopold & Vita-Finzi, 1998).

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Collins et al. Pasinler Basin paper: Table captions

Table 1. Radiocarbon dates from core 2 at Bulemaç. [a - relative to PDB-1 international standard. b - Following convention of Stuiver and Polach 1978, normalized to $\delta^{13}C$ of -25‰. c - calibrated 2 standard deviation range using OxCal 3.5 (Bronk-Romsey 2000). Calibrated dates are given in calendar years and years before present (0 BP= 1950AD)].

Table 2: Stratigraphy of the valley floor sedimentary succession (terrace 1 and floodplain) west of Pasinler and interpreted palaeoenvironments. NB depth ranges for PWa-f are taken from Bulemaç core 1, PWg and PWh are taken from nearby river bank exposures

Collins et al. Pasinler Basin paper: Figure captions

1. Turkey and adjoining countries, showing major rivers and the location of Pasinler (marked by large 'P', east of Erzurum)
2. The Pasinler Basin, showing major drainage lines, spot heights, and the locations of recorded sections and cores. Sections mentioned in the text or on other figures are highlighted. Inset: long profiles of the main channels
3. A: Exposure of sediments underlying Terrace 1, in the Inçesu stream bank near Bulemaç (~section T1-1). B: Deformed floodplain sediments (Section F-2), showing folding and breakage of the original bedding. Note the horizontal root which occurs near the base of the floodplain silts and is at approximately the same depth below the surface as a horizontally lying piece of plastic in an adjacent exposure. The four holes are *Riparia riparia* (Sand martin) burrows.
4. Palaeoenvironmental results for Bulemaç core T1-4.
5. Schematic morphostratigraphy of the Late Pleistocene and Holocene fluvial landforms and sediments of the Pasinler Basin, showing the subunits underlying Terrace 1 and the floodplain.
6. Comparison of palaeoenvironmental histories for selected sites across the Eastern Mediterranean region: Pasinler - this study; 2) Konya - Roberts *et al.* 1999, Reid *et al* 1999, Fontugne *et al.* 1999, Karabiyikoğlu *et al*. 1999; 3) Hatay and Euphrates valley - Wilkinson 1999; Lake Van (Landmann *et al.* 2002) and Dead Sea (Heim *et al*. 1997; Enzel *et al*. 2003), showing changes between high and low levels; Soreq Cave, Israel (Bar-Matthews & Ayalon 1997), showing changes between wetter and drier climates.

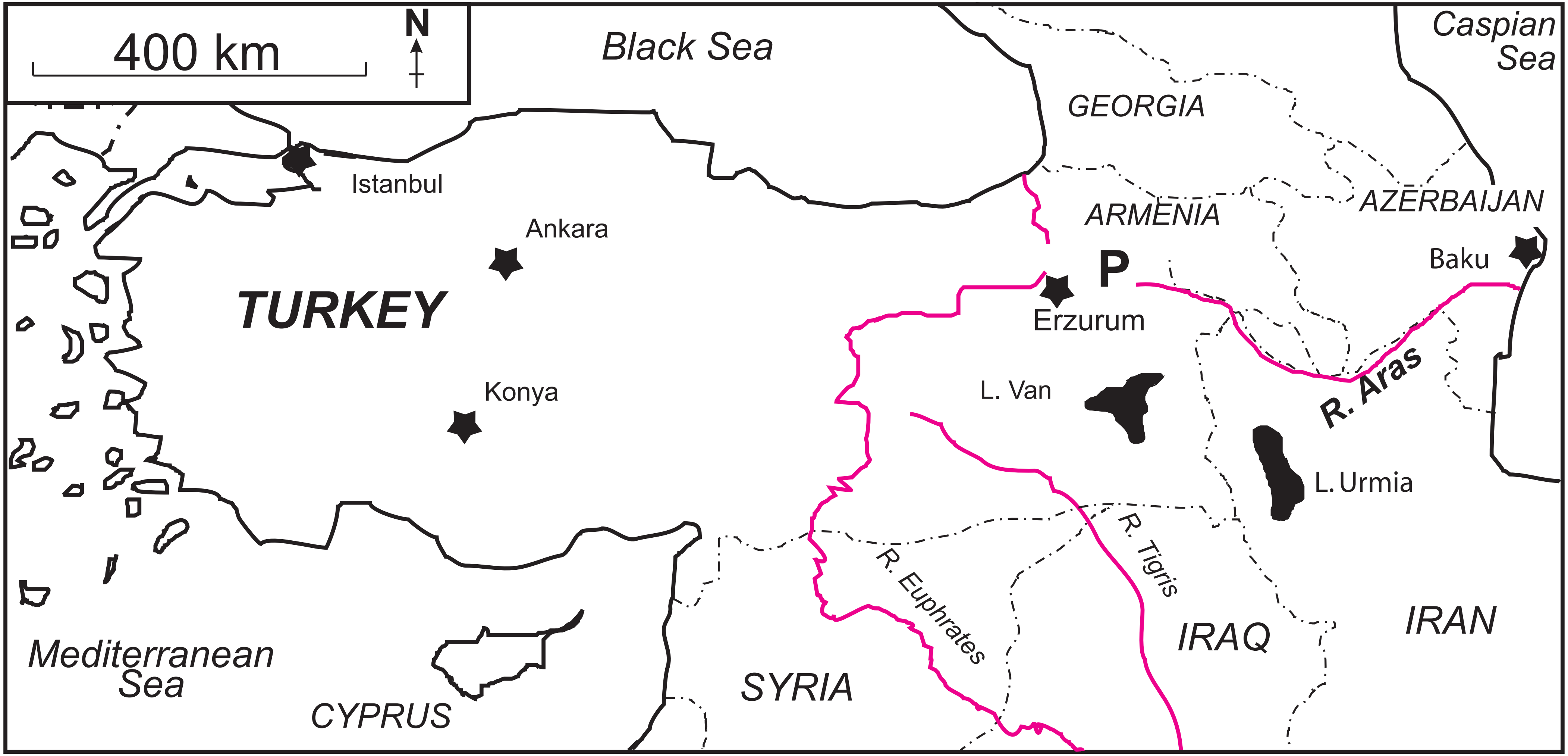
Table 1. Radiocarbon dates from core 2 at Bulemaç. [a - relative to PDB-1 international standard. b - Following convention of Stuiver and Polach 1978, normalized to $\delta^{13}C$ of -25‰. c - calibrated 2 standard deviation range using OxCal 3.5 (Bronk-Romsey 2000). Calibrated dates are given in calendar years and years before present (0 BP= 1950AD)].

Laboratory code	Depth	Material	Treatment	$\delta^{13}C$ [a]	Conventional ^{14}C date [b]	Calibrated date range [c]
Beta-132230	44-45 cm	Organic sediment	acid washes	-25.5%	3680 ± 40	4150-3890 BP [2200-1940 BC]
Beta-143324	74-75 cm	Organic sediment	acid washes	-25.7 ‰	4860 ± 60	5730-5460 BP [3780-3510 BC]
Beta-128202	136-137 cm	Organic sediment	acid washes	-25.2‰	8000 ± 40	9020-8650 BP [7070-6700 BC]
Beta-132231	155-156 cm	Organic sediment	acid washes	-26.5‰	8150 ± 50	9280-9000 BP [7330-7050 BC]

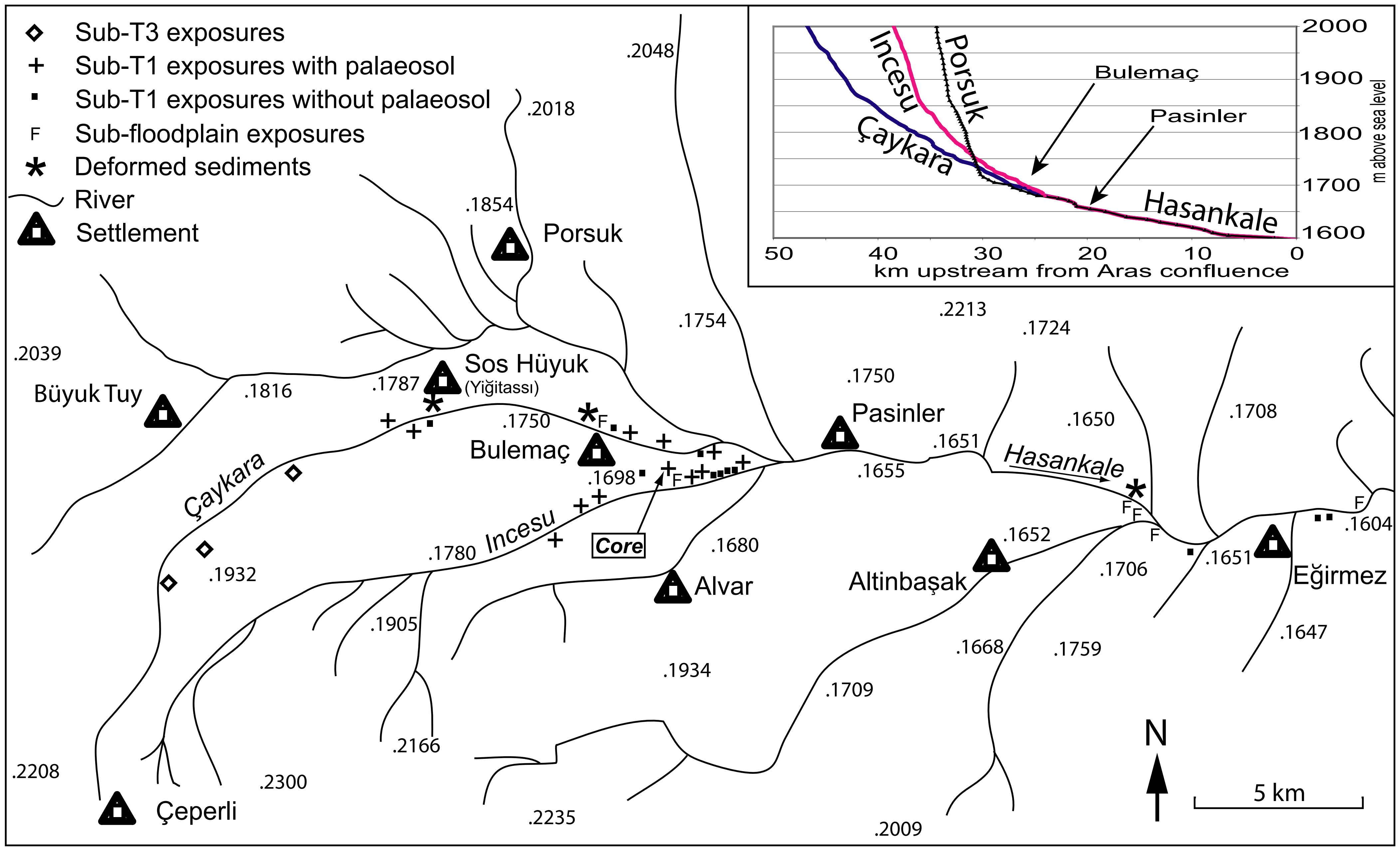
Table 2: Stratigraphy of the valley floor sedimentary succession (terrace 1 and floodplain) west of Pasinler and interpreted palaeoenvironments. NB depth ranges for PWa-f are taken from Bulemaç core 1, PWg and PWh are taken from nearby river bank exposures

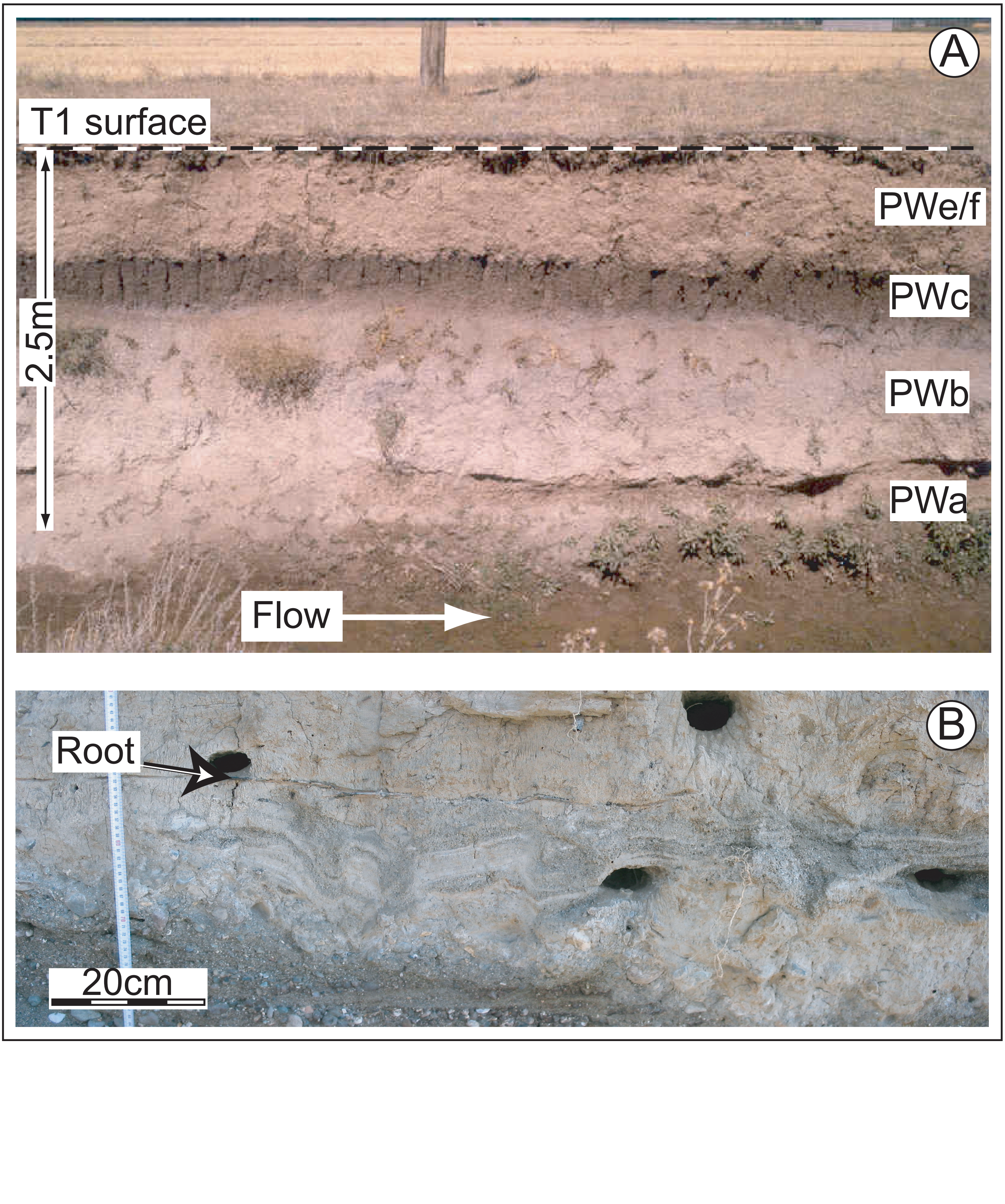
Unit	Depth range cm	Sedimentary characteristics	Depositional environment	Environmental & inferred climate	Estimated age
PWh	present floodplain	Silt, sand and gravel, some deformations, unconformably overlying	Laterally migrating channel and floodplain, incised through PWg-b	arable and pastoral farming, semi arid-continental climate	20th Century-present
PWg	(230)100-0	current-bedded sand and gravel, unconformably overlying	braided river channel cut into PWf, infilled during flood events	periodic flooding, variable flow	Post-Byzantine period (perhaps less than 1000 BP)
PWf	37-0	medium/fine sand, more organic near surface, small broken gastropods and fine obsidian-rich sand at 16-18 cm depth paraconformably overlying	Seasonally damp floodplain with overbank deposition	some woodland, steppe grassland and arable humid-continental climate	younger than 3800 BP
PWe	124-37	organic silt-clay, sand bed at 71-72cm, some gravel at 79-80 cm and 84-85 cm, shell fragments at 38-40 cm merging into	Seasonally damp floodplain, rare overbank flooding	*Pinus* dominated woodland nearby humid climate	*circa* before 5700 BP to 4100 BP
PWd	134-124	silty fine sand, shell fragments at 132-134 cm, becoming less organic towards top unconformably overlying	overbank deposition	very limited woodland damp floodplain storm or thaw flooding	after *circa* 8600 BP, before 5700 BP
PWc	182-134	organic silt-clay, merging into	Seasonally damp floodplain	some woodland damp floodplain humid	before 9300 BP to *circa* 8600 BP

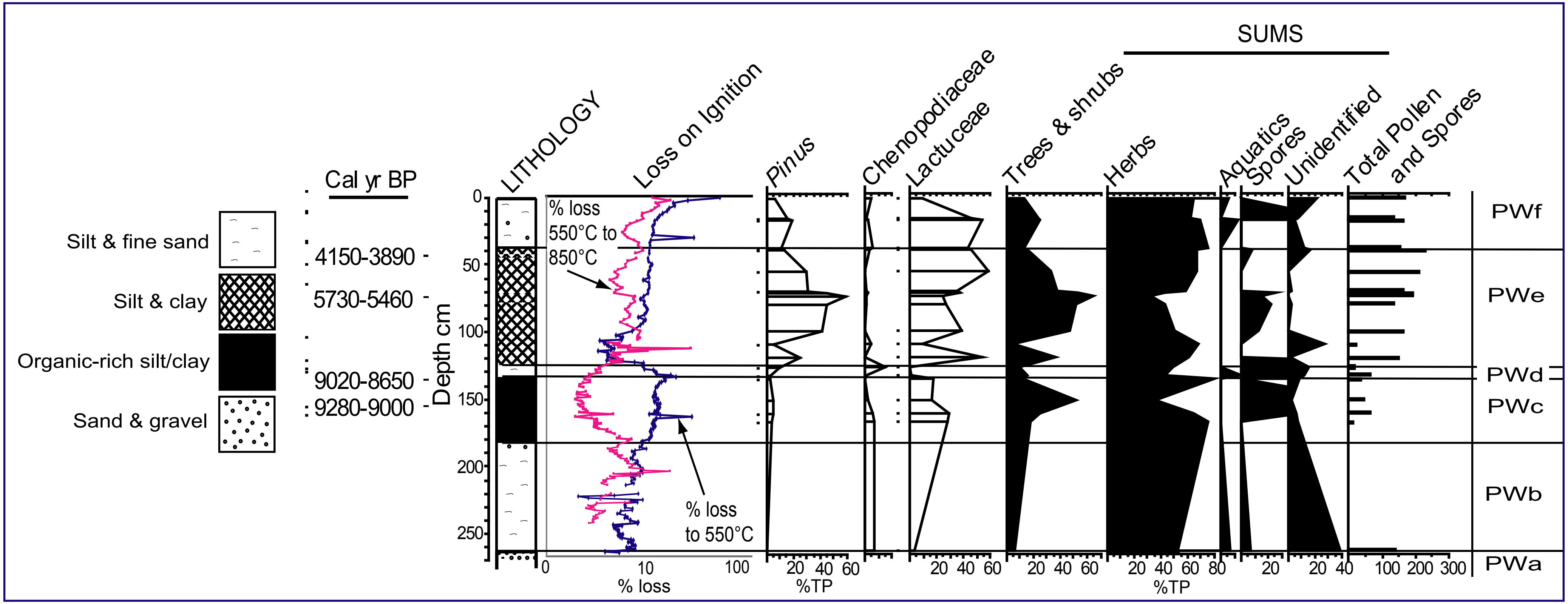
PWb	263-182	fine sand, silt-clay, some gravel ?unconformably overlying	overbank deposition	few trees, steppe in catchment	before 9300 BP
PWa	below 263	medium sand and fine gravel	river channel deposition	no data	before 9300 BP



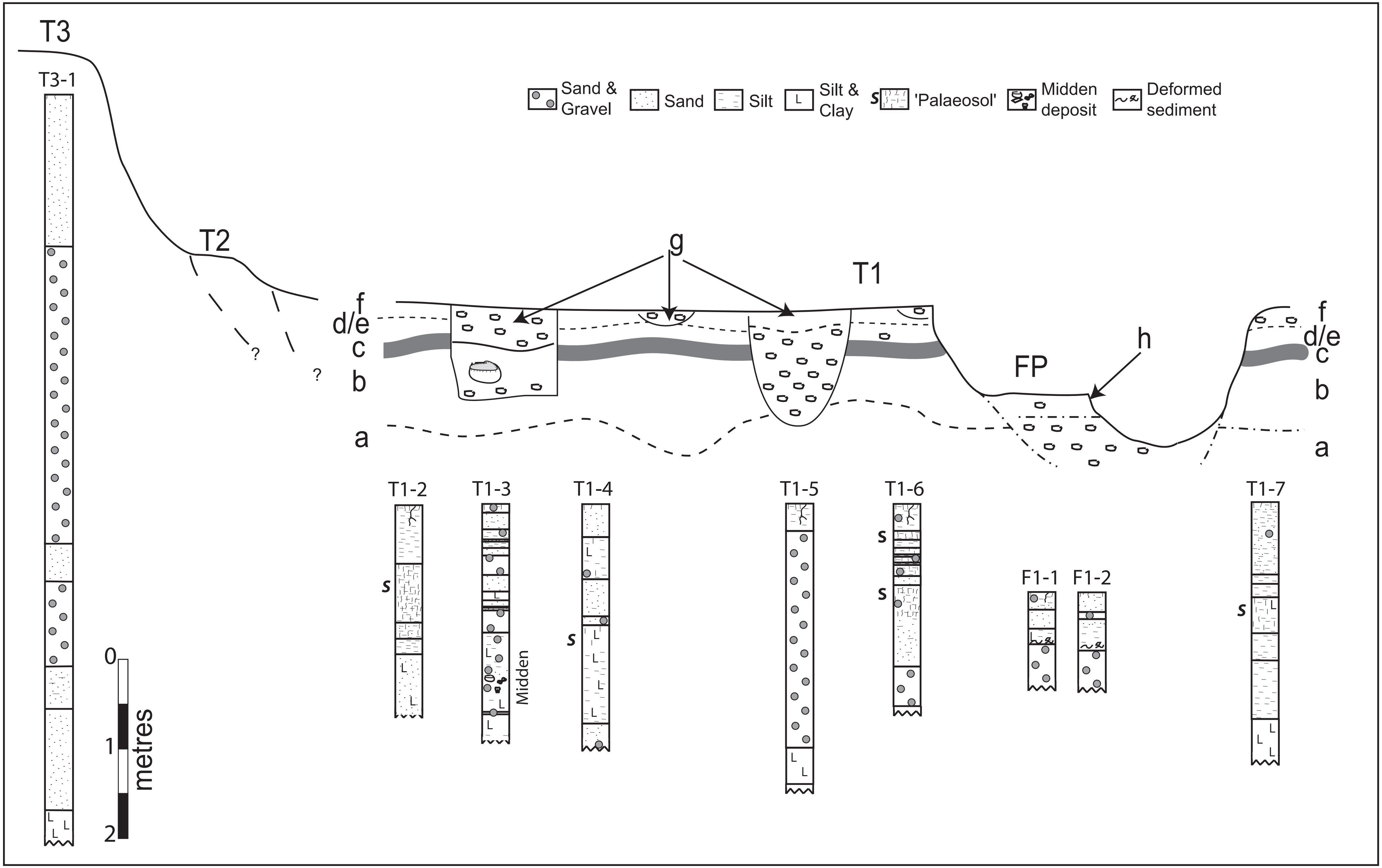
400 km
N
Black Sea
Caspian Sea
GEORGIA
ARMENIA
AZERBAIJAN
Istanbul
Ankara
TURKEY
Konya
P
Erzurum
Baku
L. Van
R. Aras
L. Urmia
R. Tigris
R. Euphrates
IRAQ
IRAN
SYRIA
Mediterranean Sea
CYPRUS


Sub-T3 exposures
Sub-T1 exposures with palaeosol
Sub-T1 exposures without palaeosol
Sub-floodplain exposures
Deformed sediments
River
Settlement
Porsuk
Sos Hüyuk
(Yiğitassı)
Büyuk Tuy
Bulemaç
Pasinler
Hasankale
Çaykara
Incesu
Core
Alvar
Altinbaşak
Eğirmez
Çeperli
.2048
.2018
.1854
.1754
.2213
.1724
.2039
.1816
.1787
.1750
.1750
.1698
.1650
.1708
.1651
.1655
.1932
.1780
.1680
.1652
.1706
.1651
.1604
.1905
.1934
.1668
.1759
.1647
.1709
.2166
.2208
.2300
.2235
.2009
N
5 km
2000
1900
1800
1700
1600
m above sea level
50
40
30
20
10
0
km upstream from Aras confluence

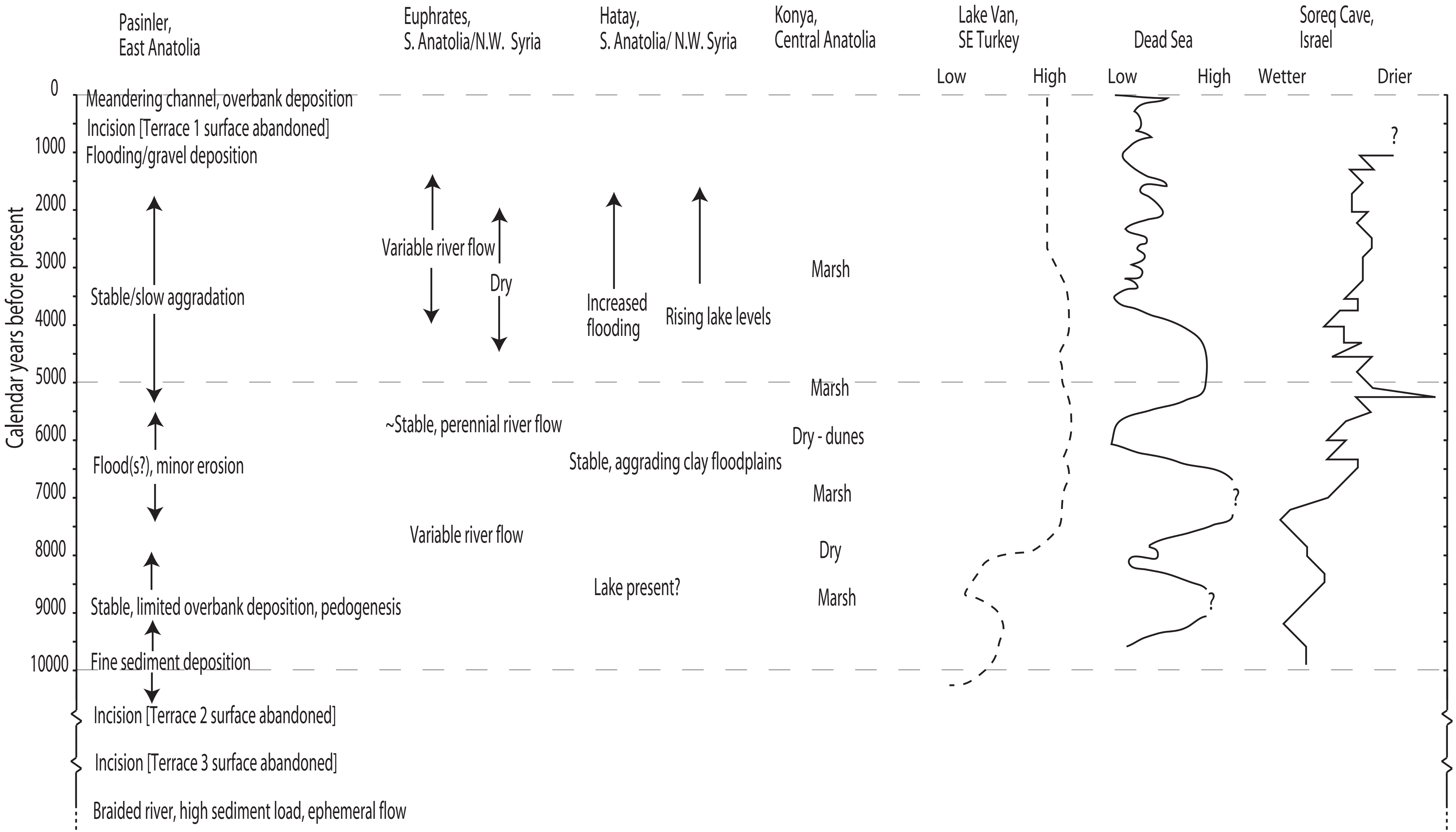

A
T1 surface
2.5m
PWe/f
PWc
PWb
PWa
Flow
B
Root
20cm



Silt & fine sand
Silt & clay
Organic-rich silt/clay
Sand & gravel
Cal yr BP
4150-3890
5730-5460
9020-8650
9280-9000
Depth cm
LITHOLOGY
Loss on Ignition
% loss 550°C to 850°C
% loss to 550°C
% loss
Pinus
%TP
Chenopodiaceae
Lactuceae
SUMS
Trees & shrubs
Herbs
%TP
Aquatics
Spores
Unidentified
Total Pollen and Spores
PWf
PWe
PWd
PWc
PWb
PWa



T3
T3-1
T2
g
T1
f
d/e
c
b
a
FP
h
Sand & Gravel
Sand
Silt
Silt & Clay
s 'Palaeosol'
Midden deposit
Deformed sediment
T1-2
T1-3
T1-4
T1-5
T1-6
F1-1
F1-2
T1-7
Midden
metres
0
1
2


Pasinler, East Anatolia
Euphrates, S. Anatolia/N.W. Syria
Hatay, S. Anatolia/ N.W. Syria
Konya, Central Anatolia
Lake Van, SE Turkey
Dead Sea
Soreq Cave, Israel
Low
High
Low
High
Wetter
Drier
Calendar years before present
0
1000
2000
3000
4000
5000
6000
7000
8000
9000
10000
Meandering channel, overbank deposition
Incision [Terrace 1 surface abandoned]
Flooding/gravel deposition
Stable/slow aggradation
Flood(s?), minor erosion
Stable, limited overbank deposition, pedogenesis
Fine sediment deposition
Incision [Terrace 2 surface abandoned]
Incision [Terrace 3 surface abandoned]
Braided river, high sediment load, ephemeral flow
Variable river flow
Dry
~Stable, perennial river flow
Variable river flow
Increased flooding
Rising lake levels
Stable, aggrading clay floodplains
Lake present?
Marsh
Marsh
Dry - dunes
Marsh
Dry
Marsh
?
?
?