The Early and Middle Pleistocene archaeological record of Greece : current status and future prospects

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6 – Quaternary landscape evolution and the preservation of Pleistocene sediments

6.1 INTRODUCTION

The landscape of Greece has long been used as a natural laboratory where prominent scholars from various disciplines of Earth Sciences and Humanities applied and tested their models, developed theoretical frameworks and elaborated on different methodological approaches. The Aegean Sea and its surrounding areas comprise one of the most rapidly deforming parts of the Alpine-Himalayan belt, and as an active tectonic setting it has contributed profoundly to resolving fundamental issues in structural geology and plate tectonics, hydrogeology, geomorphology, and many other subfields of geology (*e.g.* McKenzie 1978; Le Pichon and Angelier 1979; Leeder and Jackson 1993; Jackson 1994; Bell *et al.* 2009). Tectonic activity restricted the development of broad alluvial reaches in Greece (Macklin *et al*. 1995). Coupled with a markedly seasonal climate, this configuration resulted in the development of a landscape which does not promote extensive ecological zonation. The prevalence of mosaic environments, with a striking diversity and variety of ecological resources over short distances, has attracted the interest of ecologists and biogeographers (*e.g.* Tzedakis *et al*. 2002b; Medail and Diadema 2009). Major researchers working in the field of Palaeolithic studies and/or Landscape Archaeology were soon to appreciate the opportunities that this highly 'broken-up' geographical setting offers for the unraveling of key aspects in human-environment relationships. For example, Higgs and Vita-Finzi developed the method of site-catchment analysis during their work in the rugged relief of Epirus, initiating a long-lasting tradition of ecological/landscape approaches in the study of hunter-gatherer economy, which draws much attention to the topographical and geomorphological attributes of the landscape (Higgs and Vita-Finzi 1966; King and Bailey 1985; Bailey *et al.* 1993; Bailey 1997). Despite the major contributions from geological and geographical investigations, and notwithstanding this rather early interest by archaeologists in the role of the landscape, the latter was for a long time conceived essentially as a static, inexorable background that needs to be solely reconstructed in order to become the setting for the archaeological narrative. In this respect, it is only recently that researchers have been encompassing a more integrated and holistic perspective of landscape development in the frames of Palaeolithic investigations (*e.g*. Runnels and van Andel 2003).

Although the role of climate, erosion and tectonic movements was stressed already by the first pioneering researchers (*e.g.* Higgs and Vita-Finzi 1966; King and Bailey 1985), it was only later that an emphasis was given to such factors as agents of bias in the formation of the geoarchaeological archive (*e.g*. Bailey *et al*. 1992; Runnels and van Andel 2003). Inevitably, such a discourse was bound to be focused on and restricted in the spatial and temporal frames defined by each project's objectives. It was basically a combination of research biases (*e.g*. research models targeting primarily caves and rockshelters until about the 1980's; Runnels 2003a), the lack of robust environmental data sets and the limited evidence from excavated sites that hindered the development of broader syntheses with respect to Quaternary landscape evolution.

In this light, the following chapter aims to contribute to the understanding of landscape evolution in Greece during the Quaternary and how this might have influenced the geological and geomorphological opportunities for preservation of Lower Palaeolithic material. Needless to say, the temporal and spatial scale for such an endeavor does not allow for proper modeling. It does allow, though, for a critical overview of the main aspects of climatic fluctuations, tectonic activity, sea-level changes and slope processes, as well as the associated geomorphic controls imposed by those four principal agents of landscape change. The explorative and/or speculative nature of some parts of this treatment is believed to be justified by the apprehension that the aim has first and foremost an archaeological origin: in this regard, rather than high-resolution patterns, it is highly robust ones that are sought.

All in all, the geomorphological perspective advanced here in order to assess past, actual, and potential effects of geomorphic processes upon archaeological preservation and visibility, serves primarily as a starting point for:

1. evaluating the existing status of the Greek Lower Palaeolithic record, with regard to the issue of 'absence of evidence' or 'evidence of absence',
2. understanding and anticipating geomorphic biases, and
3. developing analytical tools and models for future investigations

The critical examination of the Greek 'Lower Palaeolithic' evidence (chapter 4) demonstrated that the record of Greece is in marked contrast to those of other circum-Mediterranean areas (chapter 3), in its main quantitative and qualitative characteristics: there are very few sites and there is an overall lack of stratified material. The fact that a large portion of the evidence lacks a stratigraphic context and/or is associated with secondary contexts emerges as a wider pattern that cannot be attributed to inappropriate research designs, the intensity of investigations or a lack of specialists in the field, as was discussed above in chapter 5. The exploration presented below assesses whether this general 'absence of (stratified) evidence' could be ultimately regarded as 'evidence of absence' for archaeologically visible hominin activities. The remains of these activities are likely to have been preserved, accessible/visible and stratified until the present only in areas where the relevant geological record is equally complete enough and has remained largely undisturbed. Disturbance versus preservation, erosion versus deposition, and deposition/preservation versus archaeological visibility/accessibility, are all conditioned mainly by geomorphic factors. These factors and their potentially biasing effects upon the archaeological record are tightly interrelated but are examined here as separate as possible, into four major groupings: climate, tectonism, sea-level changes and surface (slope) processes. When viewed in conjunction (section 6.6), a conclusion can be drawn on how geomorphic processes have shaped the available geological opportunities, which in turn configured the nature and extent of preservation of the archaeological record.

On this basis, we can arrive at both a quantitative and qualitative assessment of the current picture: how much of the archaeological record may have been lost compared to the geological record at our disposal, how much of it is likely to have escaped the biasing geomorphic agents and what kind of geoarchaeological contexts are we facing today and we should expect to deal with in the future. In this sense, the results of this exploration do not only touch upon the evaluation of the evidence at hand, but also anticipate future research and the methodological toolkit that we need to develop in dealing with geomorphic biases. Despite the fact that the landscape of Greece has attracted early on the interest of earth-scientists and archaeologists, the literature so far lacks a synthesis of landscape evolution in Greece during the Quaternary; it is hoped that the following lines will contribute in filling this vacuum, even if the perspective here remains essentially an archaeological one.

6.2 CLIMATIC CONTROLS

"If tectonics and lithology favor erosion, the main determinant of when it happens is weather" (Grove and Rackham 2001)

6.2.1 The climate of Greece

Greece has a Mediterranean climate,[42] namely one with hot, dry summers and mild, humid winters, where winter rainfall is at least three times more than

42. As a result of the interactions within a mosaic of environmental processes and ecological responses between biotic and abiotic factors at a wide range of spatial and temporal scales, the climate of the Mediterranean displays a vast diversity of features and hence a variety of climate sub-types (Allen 2001). There are, however, general characteristics which are common in the entire basin. In that sense, the general term 'Mediterranean climate' is

that of the summer. As the Mediterranean basin is situated within the boundary between subtropical and mid-latitude atmospheric patterns, its climate is particularly sensitive even to minor changes of the general circulation (Berger 1986), for instance shifts in the location of the mid-latitude storm tracks or sub-tropical high pressure cells (Giorgi and Lionello 2008). Such shifts are thought to be partly responsible for the Quaternary climatic fluctuations and the related changes in the seasonality and geographic distribution of precipitation (Macklin *et al.* 1995). Pressure conditions are markedly contrasted between the western and eastern parts, with the latter being affected mainly by the South Asian monsoon and the Siberian High Pressure System (Xoplaki *et al*. 2003). Thus, in terms of precipitation patterns, there is a broad contrast between double maxima of autumn and spring rainfall in the northern parts and a winter maximum in the southern, whereas the amount of rainfall and the duration of the rainy season decrease from west to east and north to south, with summer drought increasing in the same direction in both duration and intensity (Macklin *et al*. 1995). Interactions between different depression regimes result in unequal distribution of rainfall throughout winter, concentrated into a few days per month or season (Allen 2001). Summer drought lasts longer in the south-eastern parts, extending for up to five consecutive months. Drought is intensified by desiccating regional winds of continental tropical origin, such as those coming from Algeria and the Levant, whereas occasional monsoon air masses may promote summer rainfall.

The establishment of the Mediterranean climate (notably, with a seasonal precipitation mode and a predominantly sclerophyllous vegetation) occurred progressively around the end of the Tertiary and is associated with two major climatic changes. The first occurred at *ca*. 3.2 Ma and introduced a dry summer season together with an increase in sclerophyllous taxa, whereas the second refers to the onset of Northern Hemisphere glaciation and global cooling (Suc 1984). Pollen records from southern Italy and Sicily indicate that significant latitudinal vegetational and climatic (*e.g*. thermal) gradients existed in the Mediterranean Basin already in the (late) Pliocene (Bertoldi *et al*. 1989); a longitudinal gradient was superimposed on the latter ones, reflecting the influence of the Asian monsoon (Suc and Popescu 2005). Other lines of evidence suggest that the Mediterranean climate may have been established intermittently during the course of the Tertiary, or even much before (Tzedakis 2007). In this light, 'establishment' does not mean 'permanence' (ibid, 2059) and the bi-seasonality of the climate has not been consistent since that time, due to the Quaternary climatic fluctuations (Allen 2001). Hence, "Mediterranean conditions would appear during interglacials (reaching their maximum expression during boreal summer insolation maxima), but would not persist during glacials" (Tzedakis 2007, 2059).

The climatic signal in Greece is modulated inter-annually and inter-seasonally and at small spatial scales by the interactions of a complex topography and steep relief with altitude, latitude, orography, vegetational belts and proximity to the marine littoral. In general, western Greece is influenced by low pressures in the western Mediterranean and experiences an annual rainfall between 780 to 1280 mm, whilst eastern Greece is under the influence of the Siberian anticyclone with rainfall amounts ranging between 380-640 mm per year (Kosmas *et al*. 1998a, 71). If we consider also the highest altitudes, such as the uplands of Epirus where precipitation may be >2500 mm, then the decrease in precipitation moving southeast from northwest may be up to tenfold (Fig. 6.1). Most rainfalls occur between October and March, whereas from May to October, potential evapotranspiration exceeds rainfall, creating a large water deficit for plant growth. Average air temperature ranges from 16.5° to 17.8° C. During the cold period, temperature increases with decreasing latitude, while in the warm period temperature increases from the coast to the mainland and especially the plains (ibid). According to the bioclimatic classification of the xer-

used in this chapter to describe patterns that are also applicable to Greece. Similarly, descriptions referring to the 'south-eastern Mediterranean' should be considered as denoting principals that again apply to the climate of Greece, if the latter is not explicitly stated. As the focus is on Greece and the Mediterranean basin, 'Mediterranean climate' refers to that of regions inside the Basin and not to Mediterranean-type climates of other places in the globe (often called 'Mediterranoids').

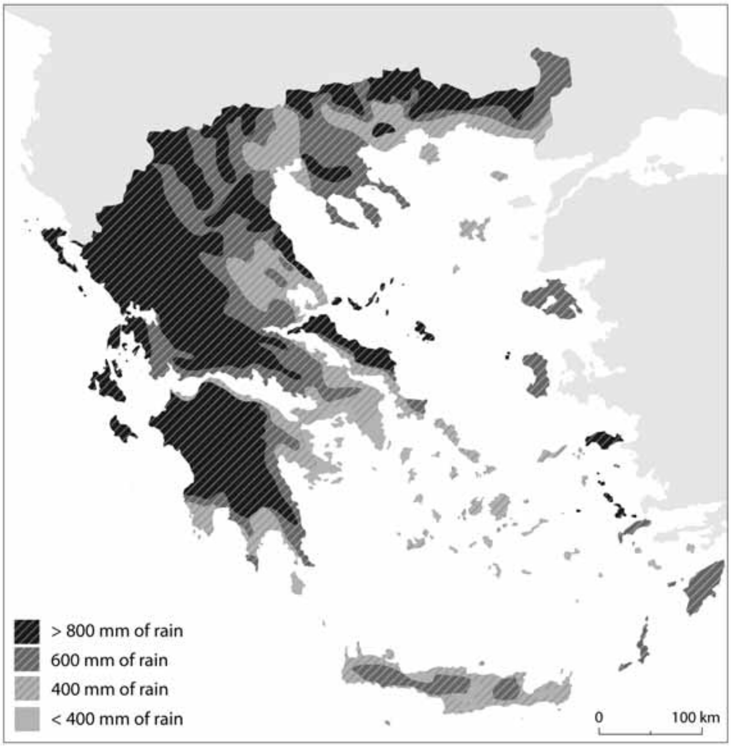

Fig. 6.1 Distribution of annual precipitation in Greece. Modified after Kosmas *et al*. 1998a: fig. 5.3

othermic index[43], most parts of Greece have a mesomediterranean climate, attenuated (*i.e.* shorter dry season) from the Albanian coast inland, over the middle latitudes of north Greece and the Peloponnesus, and accentuated (*i.e.* longer dry season) in the lowland and coastal areas (Tselepidakis and Theocharatos 1989)

6.2.2 Climate, weathering and surface processes

Langbein and Schumm (1958) studied the climatic control on fluvial denudation rates by comparing the sediment yield of drainage basins located in a variety of climates; they found that sediment yield increases with effective precipitation to a peak (at *ca*. 300 mm) followed by decline. The relationship between precipitation and erosion becomes complex and non-linear due to the influence of vegetation (Bloom 2002, 339; Jiongxin 2005). Erosion intensity depends (*inter alia*) on rainfall erosivity and the erosion-resistance capacity of the land. The latter is largely controlled by soil physico-chemical characteristics and land cover properties. Apart from lithology and grain size, there are many soil properties (*e.g.* aggregate cohesion and stability, moisture and organic matter content, porosity, bulk density, etc) and pedogenic processes that are in turn influenced by vegetation (Rettallack 2001). Vegetation cover protects the soil from erosion by the combined effects of various mechanisms: it protects the soil from rain-splash impact and crusting; canopy and litter intercept raindrops, reducing rainfall kinetic energy; organic matter builds up in the soil increasing soil moisture, which in turn enhances aggregate stability; the plant root system binds the soil together; vegetation adds to surface roughness, reducing overland flow velocity (Kirkby 1999; Casermeiro *et al*. 2004). Vegetation patterns are closely associated with the amount and temporal distribution of annual precipitation; generally, higher annual precipitation results in denser vegetation cover and higher vegetation biomass. Hence, in arid/semi-arid conditions, erosion increases as precipitation increases up to a threshold, because precipitation is still inadequate to maintain an effective vegetation cover; beyond that threshold, further increase in precipitation increases vegetation cover to the degree that the latter is now able to enhance

43. This index is the sum of the calculated indices for the dry months and provides the number of biologically dry days during the drought season (UNESCO-FAO 1963).

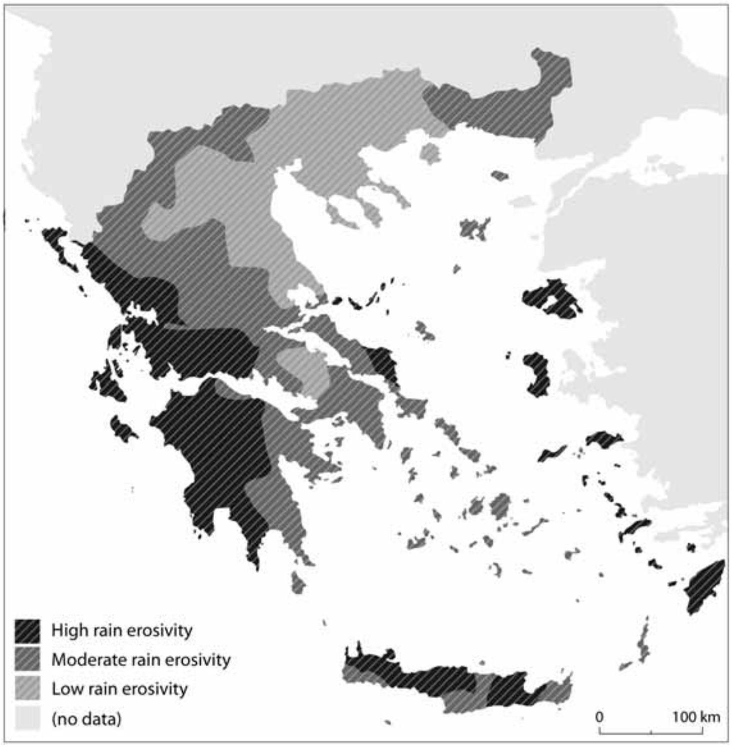

Fig. 6.2 Rain erosivity map of Greece, modified after Kosmas *et al*. 1998b: fig. 15.2. Areas of 'high rain erosivity' experience rainfalls that are unevenly distributed throughout the year; they commonly come in the form of intense storms of short duration and occur during dry seasons

soil stability and the erosion-resistance capacity of the land surface.

For a wide range of environments, both runoff and sediment loss decrease exponentially as the percentage of vegetation cover increases, and if the latter falls below a value of 40%, then, accelerated erosion dominates in sloping lands[44] (Kosmas *et al.* 1999b, 26). For the Mediterranean, the main climatic attributes controlling the degradation of landscapes, especially in semi-arid and arid zones, are the uneven annual and interannual distribution of rainfall, the extreme rainfall events and the out of phase of rainy and vegetative seasons (ibid, 19; Fig. 6.2).

Grove and Rackham argue (2001, 247) that water-related erosion in the Mediterranean depends chiefly on deluges rather than ordinary rainfall. They stress that a single deluge can be expected to have erosional consequences greater than that of ten separate falls of 10-mm-rain. With regard to fluvial erosion, a minor deluge is able to increase a river's sediment load to at least twice as much material, occasionally twenty times as much in the month of the deluge as in the rest of the year (ibid). Importantly, when such extreme events occur early in the season and plant cover is minimal, gullying and sheet erosion is promoted, whereas, late in the season, soils may have now acquired the saturation levels necessary to trigger slumping. Hence, rather than the quantity of the rain, it is pulses of high intensity within storms, which determine the erosional effect[45] (Grove and Rackham 2001, 251). Such intense blasts of rain generate high turbulent surface runoff and are influenced by topography and wind gustiness. Consequently, notwithstanding the importance of a number of feedback-mechanisms between climatic erosivity and soil erodibility, the dominant influence of erosion is the precipitation input, particularly as a result of extreme events (Mulligan 1998).

44. Macklin and colleagues place this threshold of vegetation cover at 70% (1995, 12).

45. Results showed that falls of more than 40 mm accounted for up to two thirds of the erosion, even though they comprised less than 5% of rainfall events and provided little more than 20% of the total rain (Grove and Rackham 2001, 251).

The effects of soil texture and depth, parent material, topography and climate on vegetation performance and degree of erosion were studied at the island of Lesvos (NE Aegean Sea, Greece), along three climate zones: 'semi-arid', 'transitional' and 'dry subhumid' (Kosmas *et al*. 2000). Soils of the same parent material become deeper from the semi-arid zone towards the sub-humid zone, and for all climate zones, vegetation cover increases with soil depth, whereas both vegetation and soil depth are positively correlated with increasing rainfall (but see also Mulligan 1998, 76). Under an annual rainfall of between 550-800 mm, runoff from grass or bare surface may be 120 mm greater than from forest, but trees may have less effect on the torrential rains that do most of the erosion, and many studies concur that maquis, tall undershrubs and grassland are at least as effective as forest (Grove and Rackham 2001). There is evidence that olive trees minimize the speed and amount of runoff, even under extreme rainfall events (ibid, 263), whilst scrublands are the most common plant communities in eroded areas of the Mediterranean (Casermeiro *et al*. 2004).

In another comparative project, the effects of land use and precipitation on runoff and sediment loss were studied at eight Mediterranean sites (Kosmas *et al*. 1997). The results showed that the sites with olives grown under semi-natural conditions gave the lowest rates of runoff and sediment loss. Interestingly, under shrubland vegetation cover, both sediment loss and runoff increase with decreasing precipitation as long as the latter exceeds *ca*. 300 mm per year; below this threshold, erosion decreases with increasing aridity (ibid, 57). Noteworthy, the value of 300 mm precipitation corresponds well to the threshold of the Langbein-Schumm (1958) curve and to that reported by Lavee *et al*. (1998), whereas Inbar (1992) also showed that, although it has been questioned by other studies, the general trend predicted by Langbein and Schumm is valid for basins with a Mediterranean-type climate.

6.2.3 Quaternary climate changes in Greece

There are many interrelated factors that hinder precision in reconstructing Quaternary climate fluctuations and the responses of terrestrial ecosystems to those changes. Some of the main issues concern the following points:

1. Relationships between climatic manifestations such as insolation, temperature and ice volume appear to be non-linear and/or disproportional, due to a number of feedbacks, leads and lags in the earth system; this recent appreciation could even challenge the Milankovitch theory of astronomical climate forcing (Roucoux *et al*. 2008; Maslin and Ridgwell 2005). Equally complex is the task of deciphering the imprint of climate oscillations upon terrestrial records, especially as regards the highly dynamic ecosystems of the Mediterranean; for instance, steep environmental gradients and spatial heterogeneity may result in disparate records of ecosystem change, *e.g*. from pollen-cores deriving from catchments in close proximity to each other, thus impeding a straightforward interpretation of pollen diagrams (Allen 2003). Comparisons may be similarly problematic between different land-records.
2. Continuous terrestrial sedimentary sequences spanning consequent glacial-interglacial cycles are rare due to hiatuses in sedimentation, sedimentary rate changes and lateral variability (Ehlers and Gibbard 2003).
3. Different sets of data record climatic signals of a variety of time-transgressive processes that operate in a range of spatial and temporal scales. As a result, it is difficult to disentangle causal, amplitude- and phase-relationships between the marine, ice-core and continental records (*e.g.* Tzedakis *et al.* 2001; Kukla *et al.* 2002; Tzedakis 2005).
4. All of the above associations are further complicated by divergences in the sensitivity of records and quality of resolution, which are coupled by disparities in the precision of chronological controls and the positioning of chronological anchor points. Whereas there is a relative plethora of evidence for the climate changes of the Late Pleistocene, the events preceding the Last Interglacial can be only broadly reconstructed (*e.g.* Macklin *et al.* 2002), and the discrepancies between continental and deep-sea chronologies increase for the older parts of the Quaternary (Ehlers and Gibbard 2003).

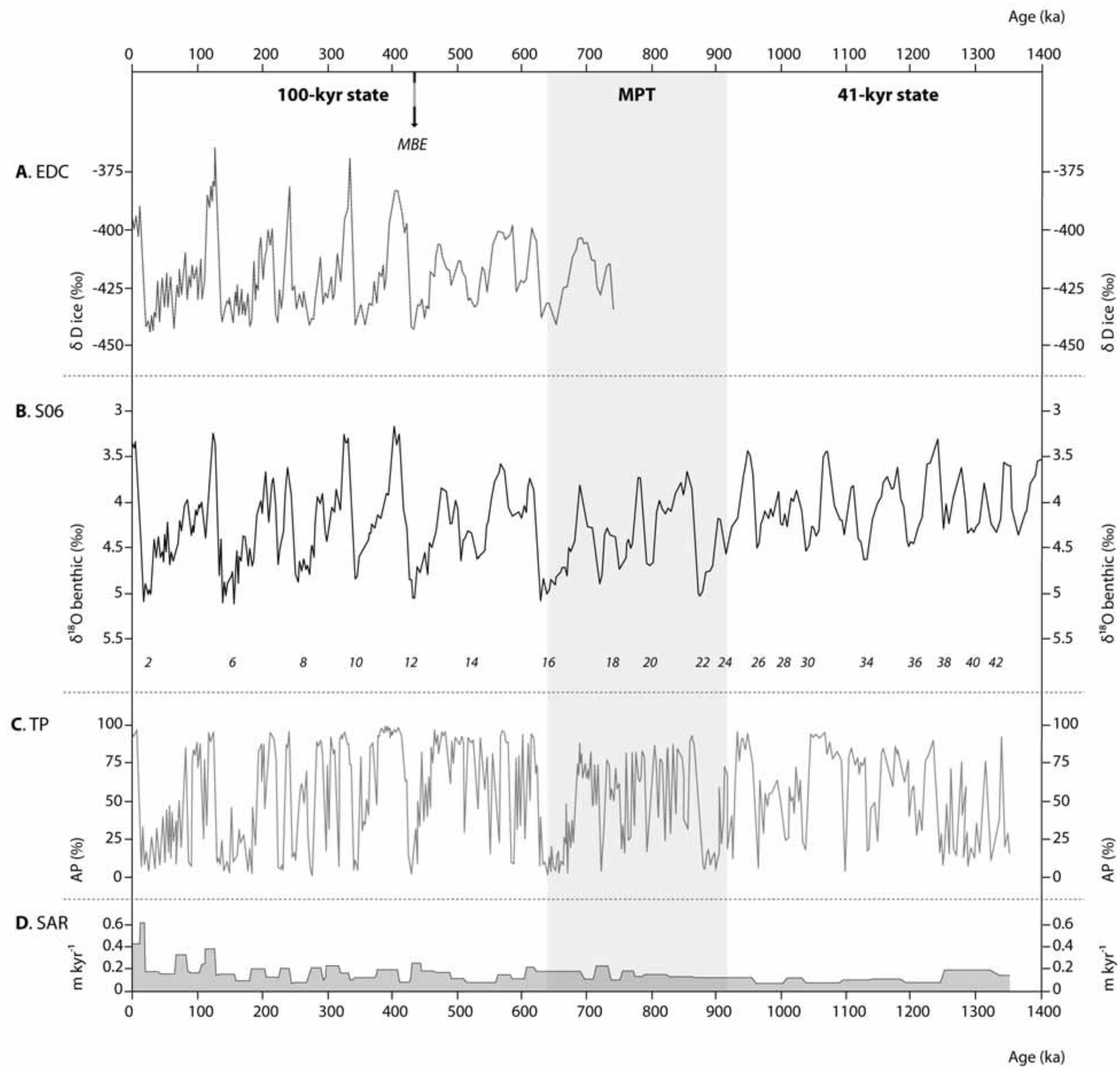

Fig. 6.3 The Tenaghi Philippon Arboreal Pollen (AP) curve (C) plotted on the pollen-orbital derived timescale and compared to (B) the S06 $\delta^{18}O$ benthic composite record from sites in the equatorial East Pacific and (A) the EPICA Dome C Deuterium (δD) record from Antartica. Glacial marine isotopic stages, sediment accumulation rates (D), the mid-Pleistocene transition (MPT) and the position of the mid-Brunhes event (MBE) are also indicated. Modified after Tzedakis *et al*. 2006: fig. 5

Early Pleistocene glacial cycles follow a pace of 41,000 yr-duration attributed to the earth's obliquity, whilst Middle and Late Pleistocene cycles have a 100-ka timescale accredited to orbital precession (Maslin and Ridgwell 2005). The change in the mode of climatic variability is known as the 'mid-Pleistocene transition/revolution' (Fig. 6.3) and is thought to be significant in terms of differences between the two periodicities in the magnitude and amplitude of climatic effects (*e.g.* EPICA 2004; Head and Gibbard 2005), although the nature and mode of the transition has recently been questioned (Huyberts

2007). It is clear though that this change marks a significant increase in global ice volume, the onset of the most extensive glaciations in the Quaternary (beginning with MIS 22) as well as a transition from linear to non-linear forcing of the climate system (Head and Gibbard 2005). Whilst this shift is generally centered at around 800-900 ka, another distinct climatic change, the 'mid-Brunhes event', roughly corresponds to the transition between MIS 12-11 at *ca*. 430 ka (EPICA 2004). This latter transition is viewed as resembling the one into the present interglacial (although longer and with marked differences in the pattern of change) and MIS 11 not only defines a boundary between two different patterns of climate (ibid), but is also considered a unique and exceptionally long interglacial, which may be the best analogue for the present climate (Loutre and Berger 2003; Raynaud *et al*. 2005; but see also Helmke *et al*. 2008).

Whereas climatic variability on orbital frequencies is now well-known, research on the last glacial period has shown that rapid climatic fluctuations occurred on suborbital (millennial-centennial) timescales (*e.g.* McManus *et al*. 1999; Alley *et al*. 2003). High-resolution lake sediment data from Italy and from a marine core in the Ionian Sea demonstrate that the North Atlantic climate variability extended its influence as far as the Mediterranean, also with regard to those high-frequency oscillations, to which vegetation communities responded equally rapidly (Allen *et al*. 1999). Correlations of terrestrial records to marine data sets, either directly through joint pollen studies and oxygen isotope analyses on foraminifera from the same marine core (Roucoux *et al*. 2006; Desprat *et al.* 2009); or indirectly, when assigning the marine timescale to terrestrial sequences by assuming synchronicity of certain events (and using glacial-to-interglacial transitions as tie-points; Tzedakis *et al*. 1997), provide evidence for a close connection between continental and marine records in terms of both orbital and suborbital variability (Tzedakis *et al*. 2006). For the linking of the records, another approach is the pollen-orbital tuning procedure, where palynological changes detected in Mediterranean cores (including Ioannina and Tenaghi Philippon, Greece; see below) are compared directly with astronomical curves (Magri and Tzedakis 2000; Tzedakis *et al*. 2006). For instance, comparison of the Tenaghi Philippon pollen curve with marine sequences (*e.g.* Fig. 6.3) showed that, on orbital frequencies, ice volume extent correlates well with tree population size, whilst on suborbital scales the land-record shows similar frequencies of peaks in steppe vegetation and North Atlantic ice-rafting events (Tzedakis *et al*. 2003b; 2006). Overall, a broad equivalence of terrestrial and marine signals has been confirmed, and the marine isotope stratigraphy can be seen as an appropriate framework also for viewing the continental record (Tzedakis *et al.* 2001, 1585). It is in this respect -and by acknowledging that marine and terrestrial boundaries may not be precisely synchronous- that the marine nomenclature ('MIS') is retained here even when referring to terrestrial stages and/or biogeographical events (*e.g.* forest expansion/contraction).

As indicated by pollen data from South European sites with sufficient moisture, an idealized scheme of vegetation phases within a glacial-interglacial cycle entails the following stages (Tzedakis 2007): a pretemperate phase of open woodland, with expansion of pioneer taxa (*Juniperus*, *Pinus*, *Betula* and *Quercus*); a temperate phase with the development of Mediterranean forest/scrub communities (warm and dry conditions), deciduous forest (warm and wetter) and montane/coniferous forest (beginning of cooling); then, a post-temperate phase of open woodland (initial cooling and drying), followed by the onset of glacial steppe vegetation (cold and dry conditions). Although temperature is an important parameter (mostly for upland and northern areas), the critical climate factor behind these shifts in vegetation composition is considered to be changes in precipitation (Woodward *et al*. 1995).

Long and continuous Quaternary sedimentary sequences in Greece are typically to be found in intermontane basins, usually of tectonic origin (Mountrakis 1985). Thus far, the best-studied polleniferous lacustrine sediments have been retrieved from three such basins (see Fig. 4.1 for their locations): Ioannina, in north-western Greece, provided a high-resolution record (*i.e.* the latest core, I-284, with a mean sampling interval of 200 years) extending back to *ca*. 450 ka (Tzedakis 1994); Tenaghi Philippon, in north-eastern Greece, has the longest continuous European pollen record, with the base of the sequence extending back to 1.35 Ma (Tzedakis *et al*. 2006); and Kopais, in central-eastern Greece, contains a record that extends into the last interglacial and up to MIS 11 (Okuda *et al*. 2001).

Interglacials and interstadials

During the past one million years, interglacials had an average duration of about half a precession cycle, namely *ca*. 10.5 ka (strictly speaking, that is for their warmest and least variable parts; Tzedakis 2007; Kukla *et al*. 2002). Evidence from the Greek and other Southern European records suggests that the onset of interglacial forest expansion is more closely associated with the timing of summer insolation peak and is less influenced by the timing of deglaciation, provided that there is no significant residual ice volume (Tzedakis 2005, 1589). Noteworthy, for the early part of interglacials coeval with boreal summer insolation maxima, palynological evidence indicates enhanced summer aridity, while isotopic data from speleothems point to increased rainfall (Tzedakis 2007). An explanation for this discrepancy argues that this excess precipitation may have come in the form of severe storm events. In turn, these events may have increased moisture, but would not add much to soil moisture availability for plant growth, as most of the water would have been quickly removed as fluvial runoff (ibid).

In contrast to the evidence from other records (*e.g.* the ice-core record from Antartica, EPICA 2004) in which interglacial maxima appear significantly cooler before the Mid-Bruhnes Event (MBE) than post-MBE maxima, the amplitude of interglacials in the Tenaghi Philippon (TP) sequence does not show any considerable difference in the extent of tree population expansions of the various temperate stages (Tzedakis *et al*. 2006). For instance, both Arboreal Pollen (AP) maxima and the vegetational character of MIS 13 and 15 are similar to post-MBE interglacials. The TP-record suggests that the most floristically diverse interglacials occurred before MIS 22-24 and that most of the relict taxa were extirpated during MIS 16. Moreover, there is a major shift in the vegetational profile of interglacials after MIS 16, with forests of reduced diversity and a 'modern' appearance (ibid). Consequently, a major vegetational change is associated with the MIS 16-15 transition, rather than that of the Mid-Bruhnes Event (MIS 12-11 transition).

When comparing the TP and Ioannina records, a first conclusion to be drawn is that they display a marked degree of correspondence in the relative expansion and contraction of forest and open vegetation communities and a tripartite division into temperate substages for the interglacial complexes of MIS 5, 7 and 9 (Tzedakis *et al*. 1997, 2003b). From the TP record, we see that during MIS 11, temperate AP values show peaks associated with substages 11c and 11a, with MIS 11b being dominated by *Pinus*. Within MIS 9, maximum forest expansion occurred during 9e, followed by 9a, with 9c displaying the lowest values (Tzedakis *et al*. 2003b). The MIS 7 interglacial complex shows broadly similar AP values for all three temperate intervals, but it is MIS 7c that had the longest duration and the most floristically diverse forest expansion, displaying also the highest insolation values within this complex and of the last 450 kyr (ibid). Instead, during MIS 7e (*ca*. 239-237 ka) there was a shift to drier and cooler conditions which resulted in forest decline and a premature ending of the terrestrial interglacial across southern Europe (this is also evident in other records from France and Italy; Tzedakis *et al*. 2004b).

With regard to MIS 7, the results from TP have recently been supported by new pollen and sedimentological data with a centennial-scale resolution from Ioannina Lake (core I-284). Four forested intervals have been identified here, with similar percentages for temperate trees suggesting similar extents of tree populations (Roucoux *et al*. 2008). In the intervening periods, drier conditions are indicated by the predominance of open vegetation, where Graminae and semi-desert taxa (*Artemisia*, Chenopodiaceae) are abundant and specific trees and shrubs disappear, reflecting decreased temperatures; yet, throughout these periods, coniferous and temperate trees persisted, signifying survival of small populations. Whereas at Tenaghi Philippon trees almost completely disappeared during MIS 7d, at Ioannina they survived in abundance. The discrepancy between the two records is thought to reflect local conditions and climatic contrasts between north-western (Ioannina) and north-eastern (TP) Greece, with the latter experiencing drier and more continental conditions. Overall, during the temperate stages, summers were probably wetter and winter temperatures were most likely lower than those of the Last Interglacial and Holocene in NW Greece. At the beginning of MIS 7e and 7c, expansion of pioneer taxa indicates colonization of open habitats with immature soils, followed by deciduous populations as the climate warmed. This pattern indicates that parts of the catchment were previously de-vegetated, experiencing severe soil erosion during the cold intervals of MIS 8 and 7d, respectively (Roucoux *et al*. 2008, 1391). In contrast, there is no clear pioneer succession at the other two forested intervals (MIS 7a and post-7.1), indicating that soils which had developed during preceding warm periods were essentially maintained through the next stadials, remaining vegetated and resisting erosion. Sediment organic content increases two- to four-times more during the forested intervals and organic content peaks coincide with times of highest pollen concentration, which in turn indicates highest percentages of vegetation biomass. Peaks in magnetic susceptibility values match closely with reductions in vegetation cover and are thought to reflect soil inwash to the lake and increased erosion rates, whereas low values would correspond to low catchment erosion under continuous vegetation cover (Roucoux *et al*. 2008, 1392).

Episodic contractions of temperate tree populations in the Ioannina record, such as that of MIS 7d, indicate oscillations on suborbital time-scales. At the onset of substage 7e, climate warming was briefly interrupted, as it is suggested by a short tree-population contraction (ibid). Such a reversal in forest expansion precedes also the onset of the last interglacial in the Ioannina record (Tzedakis *et al*. 2003a) and is manifested in the Portuguese marine core as well (Roucoux *et al*. 2006). Altogether, this data support the view that abrupt and short-lived climatic fluctuations originating in the North Atlantic were influencing Greece and have been a consistent attribute of transitions from cold to warm stages (Roucoux *et al*. 2008).

There is now growing evidence from Southern Europe pointing to abrupt events within interglacial complexes that are not accompanied by changes in ice volume (*e.g*. Brauer *et al.* 2007; Desprat *et al.* 2009). This in turn is reflected in a diachrony between terrestrial and marine stage boundaries, with temperate vegetation lagging changes in ice volume and sea level (Tzedakis *et al.* 2002a; Desprat *et al.* 2009). At Ioannina, the onset of the Last Interglacial is placed at *ca*. 127.3 ka, hence well within MIS 5e and after deglaciation was complete, whilst its end is at *ca*. 111.8 ka, indicating that terrestrial interglacial conditions persisted into the marine interval of MIS 5d (Tzedakis *et al*. 2003a), lagging *ca*. 5000 years after the building-up of ice volume. Whether such a diachrony between glacial inception and vegetation changes in southern Europe during the last interglacial was an attribute of earlier interglacials remains an open issue. Notably, as manifested in the Greek and other southern European records, a duration of *ca*. 15.5 kyr for the Last Interglacial is in contrast to estimates of *ca*. 10 kyr for the Eemian in Northern Europe (Turner 2002; Shackleton *et al*. 2003; but see also Kukla *et al*. 2002;), supporting the view of a prolonged interglacial duration in southern Europe (Tzedakis *et al*. 2004b; but see also Tzedakis 2007, 2060). For the early part of the Last Interglacial, pollen data show a peak in Mediterranean sclerophyllous taxa and oxygen isotope analyses from calcites indicate a decrease in the precipitation/evaporation ratio during the same pollen zone (Frogley *et al*. 1999, 1887); together, these proxies suggest a warmer climate with mild winters and drier summers (drought conditions). Mediterranean fluvial records indicate that this was a period of valley floor incision and soil development on stable terrace surfaces (Macklin *et al*. 2002). Comparison of last interglacial pollen evidence from Ioannina, TP and Kopais shows a general trend from mixed interglacial forests with high biomass to more open and less diverse forests, going from Ioannina to TP and Kopais (Tzedakis 2000). This variability reflects differences in climatic regimes that are also obvious in Greece today, hence suggesting similar spatial climatic patterns during the last interglacial (ibid).

Before the onset of last interglacial conditions at Ioannina, the penultimate glacial maximum (*ca*. 133-129.3 ka) was characterized by relatively dry conditions with reduced precipitation and it was followed by an interstadial (*ca*. 129.3-128.0 ka) and a stadial (*ca*. 128.0-127.3 ka; Tzedakis *et al*. 2003a). Similarly, as it is attested also in TP and Kopais, the ending of interglacial conditions took place in a stepwise fashion, with a pulse in the establishment of open vegetation followed by a short-lived reappearance of tree populations before the arrival of stadial conditions (ibid). Fluctuations occurred also within the interglacial proper in a series of subdued steps (Fig. 6.4), which were markedly smaller in amplitude and with a longer duration than the oscillations before and after the onset of full interglacial conditions (Frogley *et al*. 1999). The high-frequency and large-amplitude climatic shifts during the early and late part of the interglacial could be a reflection of rapid and severe events associated with ice-sheet decay/growth in the North Atlantic. On the other hand, for the interglacial proper, the subdued character of the oscillations may suggest that, during periods with minimum ice volumes, North Atlantic variability had a reduced influence on climatic conditions in Greece. In this latter case, the oscillations arise from responses to gradual changes in insolation, representing the crossings of environmental thresholds and "jumps" between preferred climate states: "climatic conditions would then remain quasi-stable with little vegetation overturn until the next threshold was crossed" (Tzedakis *et al*. 2003a, 165).

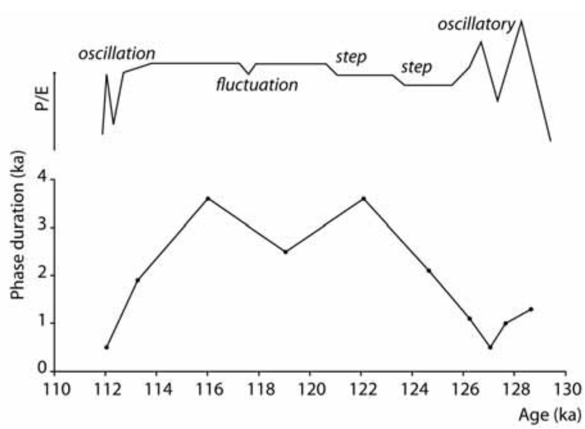

Fig. 6.4 Diagram showing climatic variability during 130-110 ka at Ioannina. Variations in the precipitation/evaporation ratio (P/E) are drawn on an arbitrary scale. After Tzedakis *et al*. 2003a: fig. 5

Suborbital climate fluctuations also characterize the Holocene (Mayewski *et al*. 2004) and, although generally weaker in amplitude than those of the last glacial cycle, many of these shifts occur rapidly (*i.e.* in a few hundred years or shorter; ibid), perhaps legitimizing the view of the present interglacial as "a period of climatic instability" (Jalut *et al*. 2009, 13). Holocene climate variability indicates that quasi-periodic changes could be abrupt and profound even in the absence of the voluminous and unstable ice masses of the Pleistocene (Mayewski *et al.* 2004), manifesting a pervasive millennial-scale climate cycle that operates independently of the glacial-interglacial climate state (Bond *et al*. 1997). As elsewhere in the Mediterranean (Jalut *et al*. 2009), the early to middle Holocene in Greece is marked by increases in non-steppe herbaceous taxa and expansion of mixed deciduous woodland (Willis 1994), which were favored by wetter conditions, as it is indicated also by lake-level data (Digerfeldt *et al*. 2007). However, since moisture availability was the main controlling factor for reforestation, the latter was completed sooner at sites with abundant precipitation, such as Ioannina in western Greece, than in the northern borderlands of the Aegean and probably also at the surroundings of Lake Xinias and Lake Kopais in central-eastern Greece (Kotthoff *et al*. 2008). The general trend towards a warmer and wetter climate in the early Holocene was interrupted by short-term climatic deteriorations, and vegetation communities were subjected to repeated, centennial-scale setbacks, mirrored by decreases in arboreal pollen and occasional increases in steppic taxa, which probably reflect reduced moisture availability (ibid). One such abrupt deterioration at around 8.1 ka is thought to be correlative with the well known 8.2 ka cold event of the Northern Hemisphere (Alley and Ágústsdóttir 2005). The colder and drier conditions of this short interval are also recorded in the isotopic record of the Soreq Cave (Bar-Matthews *et al*. 1999) and it is possible that they correspond to a major erosional event in Theopetra Cave (Thessaly) and a stratigraphic gap in Franchti Cave (Argolid) as well, suggesting a broader impact on the caves of the area (Karkanas 2001). Aridification was gradually intensified during the mid- and late Holocene, culminating at around and after *ca*. 5.6 ka (Jalut *et al*. 2009) and short-term AP minima (*e.g.* at *ca*. 5.6, 4.7, 4.1 and 2.2 ka) are thought to represent drought events in the Aegean region (Kotthoff *et al.* 2008). Nevertheless, for this later part of the Holocene and due to the advent of the Neolithic period, it is difficult to distinguish climate-induced terrestrial responses from those that should be attributed to the human impact. Since the publication of Vita-Finzi's classic work 'The Mediterranean Valleys', in which he argued for climate forcing behind major last glacial and Holocene alluviation and erosion events (the so-called 'Older' and 'Younger Fill' respectively), a fierce debate has been generated[46]. Geoarchaeological research shifted the 'Vita-Finzi paradigm' towards human agency, manifested in two main ways: settlement expansion accompanied with forest clearance and soil disturbance, and abandonment of land use practices such as terrace maintenance (*e.g*. van Andel *et al*. 1986; 1990a). Although anthropogenic causation is still retained in many interpretations of alluvial aggradation (*e.g.* Lezpez 2003) and not less in pollen-based investigations (*e.g.* Jahns 2005), a better understanding of Holocene climatic variability and the synchronicity of some fluctuations in the Mediterranean (Jalut *et al*. 2009) or globally (Mayewski *et al*. 2004), re-entered climate as a key-player and forced researchers to accept a mutual feedback between climatic triggering and human-induced disturbance (*cf.* Bintliff 2002). Importantly, what seems to progressively gain attention in explaining past landscape changes, is the significance of short-lived, natural extreme events that can be, for instance, related to tectonism (*e.g*. Zangger 1994, for a flash flood at Bronze Age Tiryns possibly associated with an earthquake); or to recurrent but non-linear climatic episodes, bringing torrential rains and/or dramatic reductions in vegetation cover (Thornes cited in Bintliff 2002).

Glacials and stadials

Absolute minima in AP values from the Tenaghi Philippon record show that the most extreme glacial interval of the last 450 kyr was MIS 12 (AP = 0%), followed by MIS 6, MIS 2 and MIS 10, with MIS 8 having the least extreme values, although AP completely disappears during its early part (*ca*. 275 ka; Tzedakis *et al*. 2003b). This pattern generally agrees well with other palaeoclimatic reconstructions (*e.g*. McManus *et al*. 1999) as well as field evidence from Northern Europe, indicating a close link between size of tree populations and ice volume not only during glacial extremes but also during periods of intermediate ice extent (Tzedakis 2005). However, while during MIS 8d the AP minimum was as extensive as that of MIS 6, this was probably due to suborbital variability (Heinrich-type event) rather than increased ice volume. Extreme phases of open vegetation are not restricted to full glacials, but occur also within interglacial complexes, for instance during MIS 5b and MIS 7d (ibid). For the interval between 920 and 450 ka the TP record shows again a close correspondence with ice volume data, with MIS 22 and MIS 16 displaying the most extreme and sustained AP minima. In fact, MIS 16 emerges as the most extensive glacial of the last 1.35 myr (the base of the TP record), not with regard to AP minima but rather because of the prolonged suppression of tree populations (Tzedakis *et al*. 2006). Interestingly, a marked transition in the vegetational composition of interglacials (thereafter being dominated by *Quercus* and *Carpinus*) occurs after MIS 16, hence at the end of the Mid-Pleistocene Transition and at the onset of the 100-ka periodicity, but it is not clear whether it relates to the effects of the MIS 16 extreme glaciation or to a change in interglacial conditions. Prior to 920 ka, AP minima are comparable to those of the Middle and Late Pleistocene, but their duration is shorter (<10 kyr) than that of MIS 16, 22, 12, 6 and 2 (ibid). Nevertheless, the Early Pleistocene AP minima at TP indicate that even short glacial intervals could occasionally have been severe enough to impose major contractions of tree populations, probably as extreme as in post-920 ka glacials (ibid).

Marine sequences from the Portuguese margin (Roucoux *et al*. 2001), the Alboran Sea (Fletcher and Sánchez-Goñi 2008) and the Ionian Sea (Allen *et al*. 1999) document the respond of vegetation to suborbital-scale North Atlantic climatic variability during the last glacial, with the largest tree population contractions associated with Heinrich Events (HEs) and less extreme climate changes corresponding to Dansgaard-Oeschger (D-O) stadials. During the low-temperature intervals of the HEs and D-O stadials, this variability would extend eastwards, intensifying cooling and aridity and triggering in-phase responses of terrestrial ecosystems across southern Europe

46. The literature around this discussion is vast and out of the scope of this chapter. For a recent review and references see Bintliff 2002.

(Tzedakis 2004a)[47]. The effects on tree populations would be most dramatic in areas with (a) moderate to low precipitation levels, as southern European forests are largely limited by moisture availability (b) low topographic variability, which reduces protection from polar air incursions, and provides only limited opportunities for altitudinal migrations (Tzedakis *et al.* 2003b). The Ioannina (I-284) pollen core reveals three general vegetation types for the last *ca*. 130 kyr: (1) forest communities during the Last Interglacial, Interstadial I (104.5-88.0 ka), Interstadial II (83-68 ka) and the early Holocene (11.5-5.0 ka); (2) communities of intermediate forest cover during the Middle Pleniglacial (59-26 ka), and Stadials 1 (111.8-104.5 ka) and 2 (88-83 ka) of the last interglacial complex; (3) open vegetation communities with woodland of scattered trees during the Early Pleniglacial (68-59 ka), the Late Pleniglacial (26.0-11.5 ka), and short intervals of the Middle Pleniglacial and during the late Holocene (Tzedakis *et al.* 2002b). The equivalents of the "Oldest Dryas" stadial and the Meiendorf/Bølling/Allerød interstadial complex have been identified in the pollen assemblages of a marine core from the northern borderlands of the Aegean Sea (Kotthoff *et al.* 2008). The Younger Dryas (*ca*. 12.7-11.7 ka) is also identified in the latter record (ibid), as well as in terrestrial pollen cores (*e.g*. Tzedakis *et al*. 2002b) and probably in the sedimentary sequence of Theopetra (Karkanas 2001), with all proxies pointing to strongly cold and arid conditions, and contraction/opening of forest cover (see Kotthoff *et al.* 2008 and Karkanas 2001 for references and discussion about the Younger Dryas, which was not, until recently, unequivocally identified in the Eastern Mediterranean).

In Greece, tree growth during the last glacial would have been constrained by increased aridity, lower atmospheric CO_2-content which intensifies water stress, and minimum winter temperatures (Tzedakis 2004a). Comparison of the three main pollen records from Greece shows that those factors had a different impact on tree populations, according to local properties and ecological threshold limits. The Ioannina basin (470 m asl) lies on the west side of the Pindus mountain range and has a sub-Mediterranean climate with high annual precipitation values (~1200 mm). The TP basin is surrounded by mountains and experiences a more continental climate with a mean annual precipitation of *ca*. 600 mm. Finally, Kopais falls within the eu-Mediterranean climatic regime with annual precipitation of 470 mm. In other words, the general trend is one of reduced precipitation from Ioannina to the sites towards the east and south, and of increased temperature from TP towards the south. For the period between 52 to 11 ka BP, pollen diagrams from TP show an almost complete disappearance of trees during HE 4 and 3 and intervening stadials, with low interstadial increases in-between, whilst similar population crashes occur also at Kopais (Tzedakis 2004a). In contrast, the Ioannina record tells a different story: although large reductions do occur, the curves remain continuous for several taxa and the minimum AP values do not fall below 21%, indicating that even during the most severe contractions there was never a complete elimination of tree populations. Palaeoclimatic simulations for the LGM (Pollard and Barron 2003) suggest that the factors controlling precipitation in western Greece today (*i.e*. basically, orographic uplift of air masses bringing moisture from the Ionian Sea) were also at work during the LGM, buffering regional aridity at Ioannina (Tzedakis *et al*. 2004a). According to the simulated values, mean January temperature and annual precipitation were -5°C and ~655 mm at Ioannina, 1°C and 180 mm at Kopais, and -5.3°C and 260 mm at Tenaghi Philippon (Tzedakis *et al*. 2004a). Hence, at Ioannina, precipitation values remained higher than the ~300 mm threshold for tree survival, whereas moisture deficiency at Kopais and TP resulted in extreme aridity and tree population crashes (Tzedakis *et al*. 2002b). Moreover, high topographic variability at Ioannina provided trees with shelter from cold air masses, the opportunity to migrate vertically, as well as a range of microhabitats suitable for survival (Tzedakis 2005). Overall, a rather distinct biogeographical pattern emerges east and west of the Pindus Mountains: the eastern arid and exposed lowlands would experience significant tree po-

47. Related to the HEs of the North Atlantic, polar waters entered the Mediterranean through the Straits of Gibraltar. Cooler sea surface temperatures during these incursions would have inhibited moisture supply to the atmosphere, in turn reducing precipitation on land and enhancing moisture stress in vegetation communities. Rapid climatic oscillations associated with North Atlantic events are also indicated by evidence from the southern Aegean Sea (Geraga *et al*. 2005).

pulation crashes during glacials and stadials, whilst the mid-altitude sites of western Greece acquired a refugial character, providing the sources for survival of residual populations and recolonization during interstadials and interglacials (Tzedakis *et al*. 2002b).

Computational experiments that compare modern conditions with possible climatic scenarios for the LGM at Ioannina reveal increased winter runoff as well as total runoff during the LGM (Fig. 6.5; Leeder *et al*. 1998). As depicted by the monthly soil erosion potential, the distribution of erosion levels is also significantly different between the two periods, reflecting an increase in the seasonality of runoff. Since the modeled values for the LGM indicate annual precipitation similar to that of the present at Ioannina, it is reasonable to assume that the effects of changing water balance upon erosion rates and sediment supply would have been more profound in less humid areas, for example Tenaghi Philippon and Kopais.

Glacial sequences in Greece (and across much of the Mediterranean) were until lately thought to have been generally restricted to the last glacial stage (Woodward *et al.* 2008), and a tentative assignment of glacial units on Olympus Mountain to MIS 8, 6 and 4-2 (Smith *et al.* 1997) could not be confirmed due to the lack of radiometric dates. Recent research on the glacial succession in Greece established a geochronological framework based on a combination of radiometric dating with morpho-lithostratigraphic analyses and pedogenic data. Glacial and periglacial units have been correlated with cold intervals in the pollen stratigraphy of the Ioannina record, which is used as a parastratotype for indirect comparisons with the marine isotope record (Hughes *et al.* 2006c). Altogether, various lines of evidence along with multiple dating techniques allowed the development of a regional chronostratigraphy, which makes the glacial sequence in Greece the best-dated in the Mediterranean and one of the best-dated in Europe[48]. Palaeo-glacial features have been reported earlier for

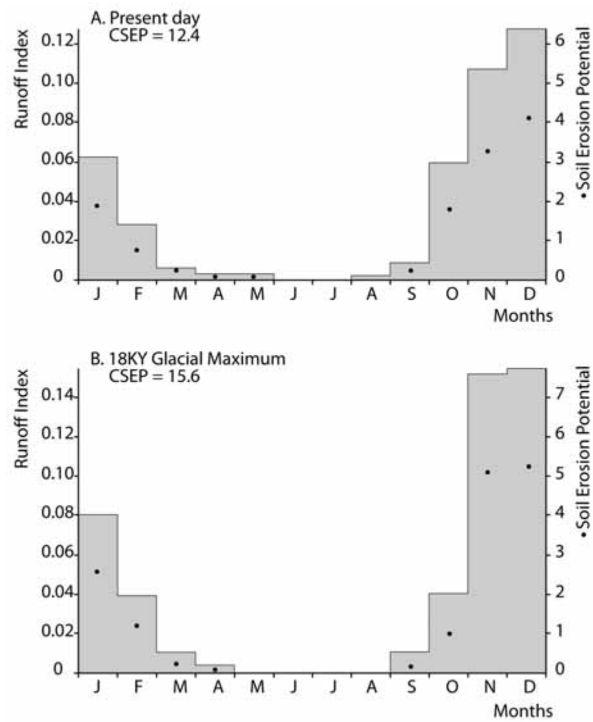

Fig. 6.5 Results of Cumulative Seasonal Erosion Potential (CSEP) experiment, comparing runoff-erosion relationships between present-day (A) and the LGM (B) at Ioannina, north-west Greece (after Leeder *et al*. 1998: fig. 6). The graphs show the values of output once equilibrium has been reached. The CSEP index is essentially a climatic index for soil erosion potential including seasonal and vegetation factors

the mountains of Epirus, Mt. Oeta, Mt. Oxia, the Agrafa area and Mt. Parnassus in central Greece, and as far south as Peloponnesus (Mt. Taygetos) and Crete (Woodward *et al*. 1995; Hughes *et al*. 2006b), but the most recent research advances mentioned above refer to Mt. Smolikas and, primarily, Mt. Tymphi; hence the following discussion is restricted to the results from research in the latter two neighboring mountains of the Pindus mountain chain.

The most extensive valley glaciers and ice fields, extending down to altitudes as low as 850 m asl, were formed during the Skamnellian Stage, which is correlated to MIS 12. Vlasian Stage glaciers (MIS 6) occupied mid-valley positions and were not as extensive as the previous ones, but did reach lower elevations than the glaciers of the Tymphian Stage

48. This is stressed here, because, as Hughes and colleagues put it (2006c, 431), "the establishment of a formal stratigraphical framework in conjunction with a nearby reference pollen parasequence has enabled, for the first time, the development of a Middle and Late Pleistocene chronostratigraphy for Greece".

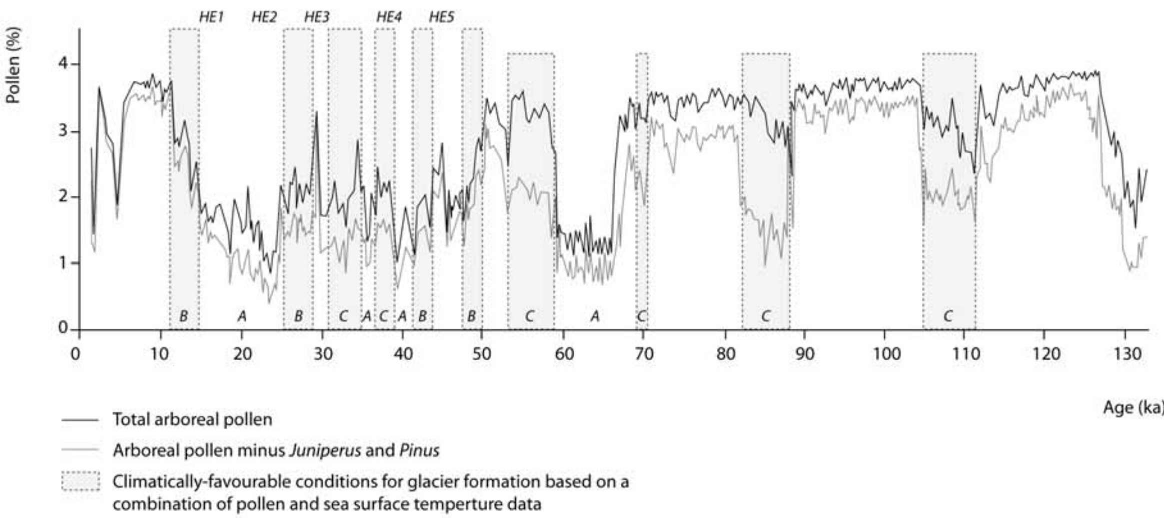

Fig. 6.6 Last glacial pollen curves from Ioannina (I-284) and potential intervals favoring glacial formation. A: major stadials characterized by low arboreal pollen; B: intermediate periods between the peaks of stadials/interstadials; C: intervals characterized by large differences between total arboreal pollen frequencies and arboreal pollen frequencies excluding *Pinus* and *Juniperus*. Heinrich events (HE) are also shown. Modified after Hughes *et al*. 2006a: fig. 5

(MIS 5d-2) (Hughes *et al*. 2007b). A similar pattern in the amplitude of glaciations, going in decreasing order from MIS 12 to MIS 6 and then 2, is reflected in the Ioannina pollen record; additionally, Late Pleistocene glaciers appear to have been significantly smaller than those of the Middle Pleistocene also in NW Spain, the Pyrenees and the Apennines (Hughes and Woodward 2008).

Hughes and colleagues (2006a) argue that the forming and decaying of Mediterranean mountain glaciers would have been fluctuating in response to the millennial-scale climate oscillations of the last glacial, and, by using the Ioannina record as a reference, they have identified at least ten time-windows that would have favored glacial formation (intervals labeled 'B' and 'C' in Fig. 6.6). Suitable conditions for glacier formation would not have been met during the climate extremes of stadials (including HEs) and interstadials, but rather during intermediate phases, when the climate was sufficiently wet -but not too warm, as during interstadials, and sufficiently cold -but not too dry, as during stadials. The last such glacier-favorable phase before the most severe and driest peak of the last glacial, occurred between 30,000-25,000 cal years BP, which is also the time of deposition of a major alluvial unit in the Voidomatis basin (Woodward *et al.* 2008). In short, glaciers on Pindus are likely to have decayed during stadials and interstadials and advance during intermediate conditions of 'cold-yet-moist' climates, rather than during the regional peaks of extreme climate (*i.e*. at *ca*. 24,000 cal. years BP in Ioannina record) or the global LGM (21,000 cal. years BP; ibid). Instead, during the heights of climatic extremity and a shifting to drier regimes, periglacial phenomena, such as rock glaciers and debris accumulation would have been prevalent (Hughes *et al*. 2003). Glacier behavior would have been unstable and glaciers may have been responsive to centennial- or even decadal-scale changes, given their small size (Hughes *et al*. 2006a).

Insights into the climatic conditions of the Middle Pleistocene glacier maxima on Mt. Tymphi can be extrapolated relative to the above-mentioned glacier-climate reconstructions. Thus, the lower equilibrium line altitudes of the Vlasian Stage (MIS 6), compared to those of the Tymphian, can be attributed to lower summer temperatures and/or higher precipitation (Hughes *et al*. 2007). Indeed, other records indicate that MIS 6 was as cold as the last glacial but with higher precipitation, as it is documented for instance between 180-170 ka (ibid, 55). By extension, Skamnellian (MIS 12) glaciers would have formed under even lower temperatures and/or higher precipitation than their successor, since they were the most extensive ones. In fact, during the Skamnellian Stage, climate can be envisaged as even colder and/or wetter than that of both the Vlasian and the Tymphian stages, with summer temperatures *ca*. 11.1°C lower than present, representing the coldest mean summer temperatures recorded in Greece for at least the last 430,000 years (Hughes *et al.* 2007). At lower altitudes such as that of nearby Ioannina (484 m asl) mean summer temperatures would have been ≤12.4°C, with winter temperatures at least -0.8°C and perhaps significantly less than that. Overall, continental conditions would have accentuated periglacial processes and physical weathering, promoting frost shattering of bedrock over large areas of the Pindus Mountains and hence also increasing the sediment supply in river systems. Importantly, between these extremely cold highlands and the very dry lowlands, the intermediate climatic zones would have been significantly narrowed compared with later glacial phases (ibid).

The Voidomatis river basin, with its highest reaches and headwaters lying within the glaciated areas of Mt. Tymphi, offers now a well-dated record of glacio-fluvial activity, representing the long-term response of the fluvial system to changes in sediment supply and valley-floor geomorphology driven by changes in the location and volume of glaciers (Woodward *et al.* 1995; 2008). The influence of glaciations to alluvial channels, by, for instance, enhanced flood magnitude and sediment supply, extended to low elevations below 500 m asl. and even to the coastal zone. Meltwater and sediment fluxes were probably greater in both magnitude and amplitude during the Skamnellian Stage, indicating a *strong coupling* between the upland glaciated plateaus and the middle and lower reaches of the Voidomatis. However, fluvial sediments of pre-MIS 6 glaciations have been either not preserved or buried below 'Vlasian' deposits. The latter are represented in one alluvial unit that indicates a major increase in sediment supply from the upstream catchment; remarkably, such major aggradation episodes, correlated to MIS 6, have been identified elsewhere in the Mediterranean as well (Macklin *et al*. 2002). Nonetheless, as in the case of MIS 12, Vlasian deposits are not well-preserved, and this is explained by the geomorphological setting: in high energy, narrow and incised gorges, long-term storage of sediments is not favored, because the formation of new valley-floor units proceeds at the expense of reworking earlier ones. This is evident in the Voidomatis record, where Late Pleistocene units are seen as the result of large floods that reworked glacial material, which had been deposited during the Middle Pleistocene glaciations. In turn, this partly explains that, although the Tymphian glaciers were the smallest ones, large-scale aggradations took place downstream by reworking limestone-dominated and till-derived coarse sediments that were inherited from previous glaciations. Woodward and colleagues (2008, 55) note that "the pattern of coarse sediment reworking and downstream transfer observed in the Late Pleistocene alluvial record […] may be a good model for the earlier glacial-fluvial interactions of Stage 12 […] In other words, the Middle Pleistocene glaciations may have generated extended periods of paraglacial sedimentation". Strikingly, large amounts of limestone in the Holocene and modern river gravels may be reflecting the *continued reworking* of coarse-grained material that belongs to the legacy of former, Middle Pleistocene glaciations (ibid).

6.2.4 Geomorphic responses to Quaternary climate changes, fluvial erosion and slope processes

Apart from the example of the Voidomatis river responding to glaciations on the Pindus Mountains, the sensitivity of Greek and other Mediterranean rivers to Quaternary climate changes is now sufficiently understood for at least the last 200 kyr, for which a relatively reliable dating control has been acquired (Macklin *et al*. 2002). Either referring to terraced fluvial sequences (*e.g.* Lewin *et al*. 1991; Woodward *et al*. 1995) or to alluvial fans (*e.g.* Demitrack 1986; Wilkinson and Pope 2003), there is a widely held consensus in correlating episodes of alluvial sedimentation with glacial/stadial periods and intervals of non-deposition/stability and/or incision with interglacial/interstadial periods (Macklin *et al*. 1995). Soil formation is also usually attributed to milder climatic conditions and soils mark the position of buried or relict palaeo-surfaces, overall representing periods of relative geomorphological stability (Woodward *et al*. 1994).

Although such one-to-one correlations (*i.e.* glacials-aggradation, interglacials-incision) have been recently challenged (for a review see Vandenberghe 2003) and great caution is needed before a direct link is demonstrated (*e.g*. Pope and van Andel 1984), results from latest fluvial research in Greece and the Mediterranean seems to corroborate this notion (Macklin *et al.* 1995; 2002; Woodward *et al*. 2008), notwithstanding the decisive role of local intra-basin attributes (*e.g*. topography, lithology; Wilkinson and Pope 2003), tectonic controls (*e.g*. Starkel 2003) or preservation biases (Bridgland and Westaway 2008a). At a 100 kyr-scale (*i.e*. a glacial-interglacial cycle), fluvial responses are broadly climate-dependent within the regional tectonic frameworks; at the 104-timescale, responses are conditioned mostly by indirect climatic impacts (*e.g.* vegetation-soil-runoff relationships), whilst at the 1000-yr and 100-yr timescales, intrinsic properties of the hydrological system and threshold conditions (*sensu* Schumm 1979), either climatic or terrestrial, become most striking (Vandenberghe 1995)[49]. At the largest scales (10^5 and 10^4) unstable phases coincide with the major climatic transitions (glacial-to-interglacial and *vice versa*) and this would be also true for 10^2- and 10^3-scales and the sub-MIS transitions (*e.g*. at the MIS 5b-5a boundary; Macklin *et al*. 2002). The problem with the latter transitions (of higher-frequency, lower-amplitude events within a glacial or interglacial stage) is that they are reflected by short-lived and sporadic episodes in the fluvial archive, which are usually either not preserved or impossible to grasp by the available geochronological techniques, as the confidence intervals on the dates typically overlap. Therefore, considering that short unstable phases alternate with longer periods of inactivity (and/or stability), the critical question remains: how short are these unstable phases and their recurrence intervals? Hosfield and Chambers (2005, 291) argue that, for north-west European fluvial systems, the time-spans of fluvial incision/erosion and sedimentation stretch over a few hundreds or at most a few thousands of years; and that archaeological assemblages in fluvial sedimentary contexts are "*unlikely* to have lain undisturbed on river floodplains and/or channel margins for more than 2-3 kyr between major/minor episodes of fluvial activity". The latter authors suggest that artefact accumulations which are now associated with secondary fluvial contexts, would be reworked into fluvial sediments *every few thousand years*; hence they represent temporal palimpsests with a time-depth that varies according to site-specific factors but is generally in the order of a few thousand rather than tens of thousands of years (ibid, 294). Although this conclusion was drawn based on north-west European fluvial research, it can be viewed as broadly applicable to Mediterranean landscapes, for reasons that I explain below.

By both global and Mediterranean standards, most of the river basins in Greece are small, and, taking the 500 m contour as the mountain-lowland boundary, are drained by steep-land river systems[50]. This configuration stems from an intense (and still active) tectonic history that sets the background for a basin-and-range topography, which has restricted the development of extended alluvial channel reaches (Macklin *et al*. 1995, 19). Coupled with a strong climatic seasonality, this results in hydrological regimes that are marked by steep hydrographs[51] (Paspallis 2003). Moreover, sediment yield data from Mediterranean landforms lie high above the world averages, and rates of erosion are one or two orders of magnitude higher in basins with steep relief (Inbar 1992).

Besides rivers that are fed by groundwater in limestone terrains, Mediterranean river regimes today reflect the seasonality in the distribution of precipitation, and runoff patterns essentially result from rainfall alone (Macklin *et al.* 1995, 12). In most of Greece, rainfalls occur during winter or winter and autumn, with small parts of the NE and the NW ex-

49. Although these scale-related relationships stem from research in north-west Europe, they have been noted by researchers working in Greece as early as in the 1980's (*e.g.* Pope and van Andel 1984), and are reflected also in more recent publications (*e.g*. Wilkinson and Pope 2003), even if the authors do not touch these issues explicitly or as a primary focus.

50. Overall, there exist an inverse relationship between slope steepness and the extent of drainage area; see for example the plot of valley slope/drainage area in Schumm 1979.

51. A hydrograph plots the discharge of a river as a function of time.

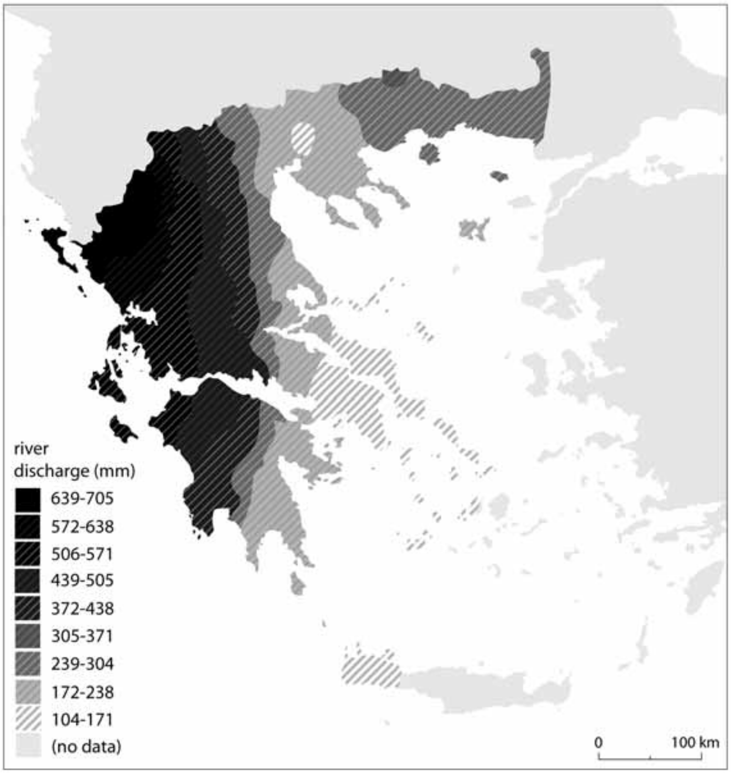

Fig. 6.7 Total annual river discharge in Greece. A comparison with Fig. 6.1 shows clearly how runoff patterns follow closely those of rainfall. Source: (Greek) National Data Bank of Hydrological and Meteorological Information

periencing rainfall peaks during autumn, whilst only the north-central part has rain throughout the year (Bartzokas *et al.* 2003). In other words, because of the seasonality in the distribution of precipitation, fluvial runoff in most of Greece is also seasonal (Mimikou 2005), chiefly concentrated in a few months of the year, with a winter or early spring peak and minimum flow in the summer (Fig. 6.7).

As noted earlier, similar spatial climatic patterns characterized the last interglacial, and, indeed, we can envisage a comparable situation for other past interglacials (Tzedakis 2000; Cheddadi *et al*. 2005), expecting perhaps a further increase in seasonality for those interglacials that were warmer than today, resulting in an accentuation of the summer (drought) season. On the other hand, during cold stages, river regimes were most probably even more seasonal and displayed greater spatial and temporal variability (Macklin *et al*. 1995). Notwithstanding this variability, seasonal flow fluctuations in rain-fed catchments would have been more pronounced, especially in southern and eastern areas, but also in northern sites with a more continental climatic regime (*e.g*. TP). Increased seasonality of precipitation has already been suggested for the LGM (Prentice *et al*. 1992), whilst excess precipitation may have taken the form of extreme storm events during the early parts of interglacials, and this latter situation could be reflected in the absence of alluvial units dated to the early part of the last interglacial, which emerges as a period of valley floor incision with soil development on stable terrace surfaces (Macklin *et al.* 2002, 1638). Given that, as stressed with regard to documented events (*e.g*. Kosmas *et al*. 1997; 2000), the dominant driver of erosion is the precipitation input, any increase in the ephemerality of flow regimes and/or a decrease in the recurrence interval of high-discharge peaks that interrupt periods of quiescence, would result in river regimes with even steeper saw-tooth hydrographs, perhaps along with concomitant increases in the frequency of high-magnitude, short-lived extreme flood events; and these are, ultimately, the major causes of landscape disturbances and soil erosion. Results from recent physically-based modeling of the effects of seasonality on annual and intra-annual water balance support the arguments above (Yokoo *et al*. 2008): the presence of climatic seasonality tends to decrease evapotranspiration and increase runoff compared to when there is no seasonality; and the effects of seasonality are stronger when seasonal variability of precipitation and potential evapotranspiration are out of phase; moreover, this tendency increases as the dryness index increases (*e.g*. in semi-arid areas), and seasonality effects are high in basins with steep topographies.

Due to the small size of the Mediterranean drainage basins and the fact that most of them include high-relief, steepland catchments, there is an overall strong slope-channel coupling, in contrast to lowland rivers of north-west Europe, which are thought to have a less effective coupling between hillslopes and stream channels (Macklin *et al*. 2002). In effect, processes operating in upland reaches are quickly affecting lower areas downstream, and we have already seen that changes in sediment supply in the uplands (*i.e.* >2000 m asl) of the Voidomatis river extended their influence below 500 m asl and down to the coastal zone. As Grove and Rackham vividly put it when describing river erosion after a deluge in Crete (2001, 248), "although rainfall was presumably more than 300 mm in the mountains, the greatest effects were at low altitudes...Most of the mayhem in the valleys was the recycling of deposits already in them". Reworking of older sediments by renewed aggradation and erosion has already been noted for the Voidomatis river, where it was stressed that there is evidence for the post-glacial, Holocene river still reworking glacial (till-derived) material. Furthermore, the slope-channel coupling would have been even more accentuated during, for instance, extreme glacial stages (*e.g*. MIS 12), when transitional climatic zones between uplands and lowlands were narrowed.

According to the above, if episodes of incision/erosion and alluviation in NW Europe "occur across timespans stretching over a few hundred, or at most a few thousand years" (Hosfield and Chambers 2005, 291), it could be argued that this may be very well true also for Greek catchments during both interglacial and glacial stages. In fact, it could be equally possible that river systems in Greece experienced even more "rapid chronologies" (ibid) and fluvial events had an even shorter duration and/or recurrence intervals, with process-response patterns orchestrated by climatic oscillations and accommodated in the highly dynamic background of a steep relief and omnipresent tectonism. Even if we neglect the undoubtedly critical role of topography and tectonics, the assertion above would be perhaps falsified only if suborbital, rapid climatic oscillations either did not occur or did not affect Greece, and/or if the coupled slope-channel system in watersheds did not respond equally fast. But that is most likely not the case. As attested by the Greek pollen records, centennial-millennial climatic fluctuations did occur and influence Greek landscapes, triggering in-phase responses of terrestrial ecosystems (Tzedakis *et al*. 2004a; see above). In Italy, the Monticchio pollen sequence furnished varved-counted intervals of decades-centuries for the end of the last interglacial (Brauer *et al.* 2007), whereas, for Greece, the last glacial glaciers of Tymphi are assumed to have been responsive to centennial- *or even decadal-scale* changes, given their small size (Hughes *et al*. 2006a, 95). Thus far, the available chronological control for Greek and other Mediterranean fluvial archives does not permit a precise estimation of the length of time over which depositional and/or erosional events occurred, but there have already been suggestions for catchment-wide sedimentation "over time periods of between 10^3 and 10^4 years" (Macklin *et al*. 2002, 1640), hence supporting the arguments presented here. Examples from the Holocene cannot be extrapolated to past climatic cycles because of anthropogenic interference, but are still elucidating how brief depositional events can be (at least from the Neolithic onwards): a late Holocene alluvial episode in Southern Argolid lasted less than 2-3 centuries, whereas 5 m of Holocene alluvial sediments in the Argive plain accumulated in a maximum of 50 years (van Andel and Zangger 1990).

However, is it then possible to see these 'rapid chronologies' as a general trend for all Greek basins? Harvey (2002, 198) notes that "in well-coupled systems, assuming effective thresholds are exceeded, and especially where there is a rapid response to environmental change, there is likely to be near synchroneity and spatial uniformity in the geomorphic response to environmental change. In well-coupled systems, the results of such changes will be basin-wide aggradation or dissection sequences." On the other hand, Pope and van Andel (1984) and Wilkinson and Pope (2003) showed that aggradation and dissection were much more localized and varied throughout the basins of Argolid and Evrotas River, respectively, and this would indicate a rather poor coupling with depositional/erosional records representing only local sequences. Still, "in poorly coupled systems, there is likely to be spatial non-uniformity, and possibly major contrasts in process regimes or stability between different parts of the system, but nevertheless, those parts of the system affected by environmental change would respond fairly rapidly" (Harvey 2002, 198). Consequently, the argumentation on the relative duration of events and/or the frequency of recurrence is not critically affected whether the basin systems are well-coupled or not, and notwithstanding the wide range of variability in inter- and intra-basin processes.

6.2.5 Discussion

Dating and preservation constraints (*e.g*. hiatuses) restrict the actual basis of the above-listed documented patterns and extrapolated assumptions to events of the Late Glacial and early Holocene. Therefore, the question is whether they are applicable to earlier periods. Early and Middle Pleistocene suborbital climatic oscillations are not as well-known as Late Pleistocene ones, but they do emerge in high-resolution records, as it is the case for MIS 7 in the latest core from Ioannina (Roucoux *et al*. 2008) or for MIS 8, in which the vegetation contraction during substage 8d is thought to be related to a Heinrich-type event (Tzedakis 2005). Furthermore, such 'sub-Milankovitch' fluctuations have been recognized for the interval of *ca*. 0.5-0.34 Ma (Oppo *et al*. 1998) and even for 1.0-0.7 Ma, displaying highest amplitudes during interglacials, particularly after ca. 900 ka: short oscillations occur during the transitions of MIS 25-24, 23-22 and 19-18 (Head and Gibbard 2005, 5). Already in 1993, the Greenland Ice-Core Project members reported on the intra-MIS 5e oscillations, noting that "the mode switches may be completed in as little as 1-2 decades and can become latched for anything between 70 yr and 5 kyr" (GRIP members 1993, 207). On the basis of ice-core data, there are now suggestions that glacial-to-interglacial reversals in the MIS-scale "may have been equally as rapid" (McNabb 2005, 290). In short, if we accept that climatic oscillations in both the MIS and sub-MIS scales are linked with fluvial activity in the order of 10^2- and 10^3-phases, then similar associations are likely to be valid for the Middle and even the Early Pleistocene. The fact that they are hardly detectable is most probably an artefact of the low-resolution in dating methods, the incomplete representation of climatic events in terrestrial archives and the fragmented preservation of these archives (*cf*. Hosfield and Chambers 2005, 293).

Different lines of evidence, from pollen diagrams (*e.g*. Tzedakis 2005), glacial records (*e.g*. Hughes *et al.* 2006c), glacio-fluvial sequences (Woodward *et al*. 2008), lacustrine stratigraphies (*e.g*. Okuda *et al*. 2001), pedogenic data (*e.g*. Woodward *et al*. 1994), alluvial sequences (*e.g.* Wilkinson and Pope 2003) and cave records (Karkanas 2001), all point to the conclusion that landscape instability was prevailing during climatic transitions. In Theopetra Cave, cryogenic events that record climatic deteriorations have been correlated with Heinrich Events of the Late Glacial and have been shown to be followed by major erosional episodes, alternating with periods of milder conditions and stability (Karkanas 2001). The evidence from Theopetra suggests that erosional events may be more prominent during changes from cold to warmer conditions (ibid, 392), which has also been assumed for the fluvial sequence of Pineios in the same region (Demitrack 1986). This may be related to the picture from the pollen record, which documents abrupt oscillations during the climatic shifts from MIS 8 to 7e and MIS 6 to 5e, indicating that such short-lived spells may have been a consistent attribute of cold-to-warm transitions (Roucoux *et al*. 2008, 1392).

A few years ago, the recognition of sub-orbital climatic variability within the last glacial revolutionized our view of Quaternary climates. At present, there is evidence that such rapid switches (Fig. 6.8) were not confined to the last glacial but occurred also during earlier glacials and interglacials, as well as during transitional phases. Both the rate and the amplitude of those sharp reversals would have had major implications with regard to terrestrial responses. Climatic seasonality, precipitation, and the seasonality of precipitation, have been stressed here as crucial factors in this context. Fluctuations in any of these variables would have imposed major stresses in landscape stability, with significant consequences for the preservation, visibility and reworking of archaeological assemblages.

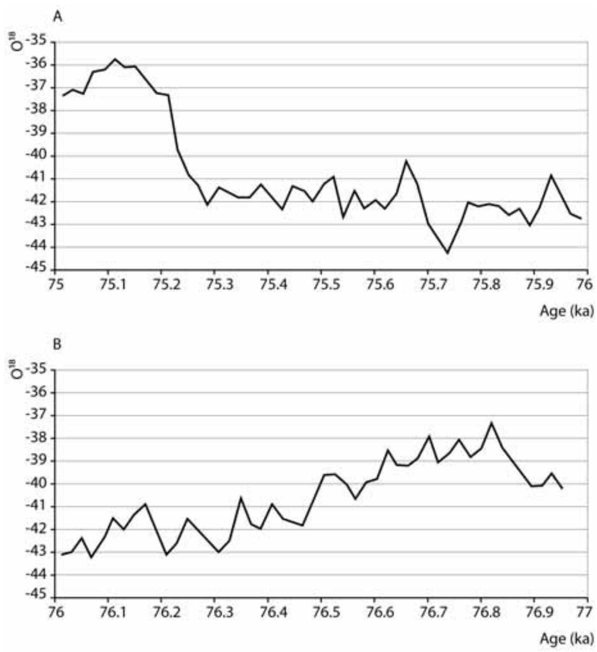

Fig. 6.8 Centennial and sub-centennial climatic changes during 76-75 ka (A) and 77-76 ka (B), as revealed in the GRIP ice-core. After McNabb 2005: fig. 1

We can therefore assume that the most stable geomorphological conditions would be confined to full glacial (stadial) and full interglacial (interstadial) periods. Yet, this proposition is valid in the most relative terms and runs the risk of lumping together periods of geomorphic activity with phases of inactivity, due to the complexity of process-response patterns and our inability to accurately pinpoint the respected intervals. For instance, paraglacial processes –*e.g.* in the form of slope failure, debris flow or fluvial reworking of sediments- are at work when deglaciated landscapes readjust to nonglacial conditions (Ballantyne 2001). On the Pindus mountains, when recently deglaciated terrains would have been in an unstable or metastable state with accelerated erosion prevailing in the context of deglacierization, paraglacial environments would be promoted under full (rather than intermediate) stadial and interstadial conditions (Hughes *et al.* 2003). Conversely, instability associated with ice progression phases, when erosion rates are increased along with glacial advances, would correspond to 'intermediate periods', *i.e.* those in-between the regional peaks of extreme climate. This example is used here to stress that, according to regional spatio-temporal circumstances, climatically-induced instability would not have been restricted only to transitional phases: rather, it would occur also within the relatively 'stable' climatic conditions of a full glacial and full interglacial period. In this light, suborbital events within interglacials (*e.g.* MIS 9e, 7e and 5e) were also highlighted here, as they mark the end -and possibly also determine the duration- of interglacial climatic optima (Desprat *et al.* 2009).

6.3 TECTONIC CONTROLS

6.3.1 Introduction

Whereas patterns of erosion and deposition (*e.g.* fluvial incision and aggradation) are largely climatically induced, these patterns are superimposed on tectonic movements, which in turn provide the underlying mechanisms for the development of topography (*e.g.* Cloetingh *et al*. 2007). In other words, if climate can be considered as the main agent of external forcing upon landscape configuration, tectonism constitutes the internal forcing of geomorphological evolution (Bloom 2002). Tracing this kind of tectonic controls over the landscape and the associated Lower Palaeolithic record that we find today, this chapter addresses two main issues:

1. The kind of tectonic regime that was prevalent during the Pleistocene, its origins and the heritance that it left upon the present landscape.
2. How did this tectonic regime influence and control large-scale geomorphological processes, particularly with regard to the preservation and visibility of the (Pleistocene) geoarchaeological record.

Despite Greece being in a convergent plate-tectonic setting, with the African plate subducting below Europe, during the Alpine orogeny (*i.e.* from the Cretaceous up to Miocene), the Internal Hellenides and the Rhodope experienced widespread extensional tectonics, notably since the Oligo-Miocene (Gautier *et al*. 1999). Here we focus in the extension that affected Greece from the Miocene onwards. The present configuration of the Greek landscape is es-

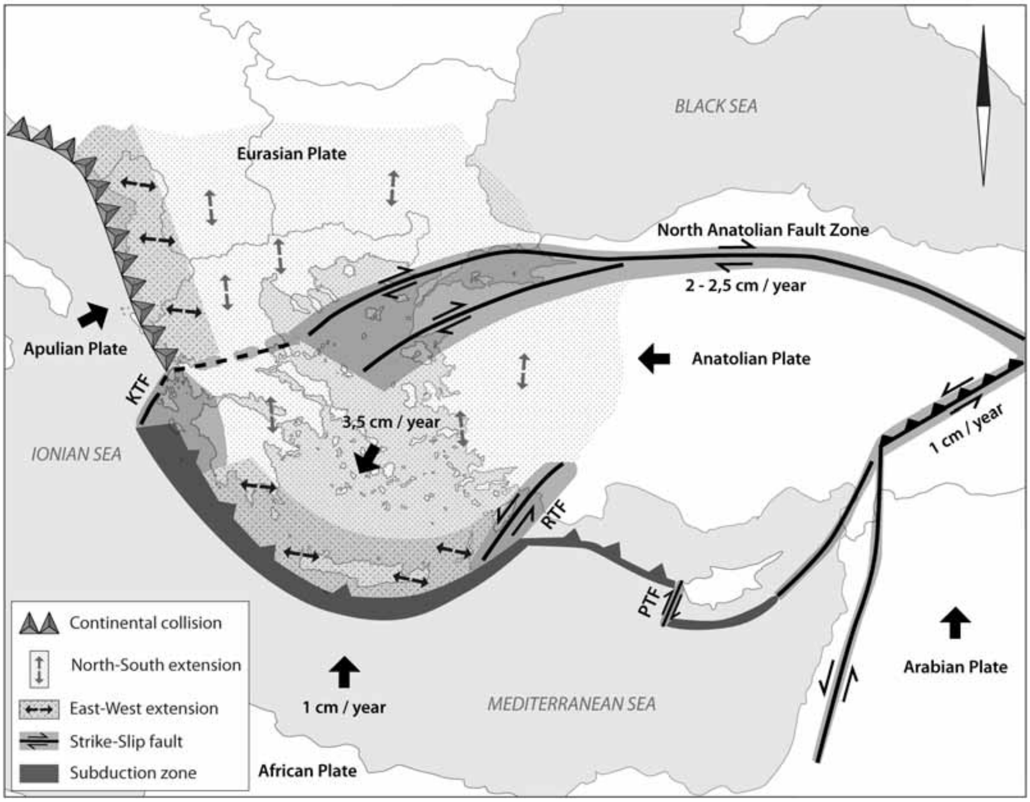

Fig. 6.9 Map of the eastern Mediterranean showing the presently active geodynamic domain in the broader Aegean area. KTF = Kephallonia Transform Fault, which marks the area where the zone of subduction gives way to a zone of continental collision between the Apulian and the Eurasian plate. Modified after Mountrakis, university notes (http://www.geo.auth.gr/courses/ggg/ggg871y/)

sentially the result of late-orogenic tectonic movements that affected the Hellenic Peninsula since the Miocene and largely continue until today (Meulenkamp 1985; van Hinsbergen *et al* 2006). These movements were polyphase and complicated, including both extensional and compressional events, within tectonic regimes that were often fundamentally opposite in adjoining areas and with multidirectional and highly variable axes of stress and deformation (Angelier 1978). Therefore, important aspects of the kinematic evolution of the region, such as the timing and style of deformation or the temporal and spatial distribution of strain, still remain controversial (Aksu *et al.* 2005).

6.3.2 Overview of the main tectonic phases

The Aegean region is one of the most rapidly deforming regions in the world (Jackson 1994; Cloetingh *et al.* 2007) and seismically the most active region in the Mediterranean and West Eurasia (*e.g.* Tsapanos *et al*. 2004). The Aegean Sea is an extensional back-arc basin that formed as a result of differential convergence rates associated with the northeastward subduction of the African plate beneath the Eurasian plate (Mather 2009). Intense and widely distributed extension commenced in the Miocene (Le Pichon and Angelier 1979), with time it migrated south-westward and was accommodated along a series of large-scale normal faults on the Hellenic Peninsula (Angelier 1978), and at present it remains ongoing (*e.g*. Billiris *et al*. 1991; Reilinger *et al*. 2010). This extensional regime is genetically related to two first-order structures (Doutsos and Kokkalas 2001; Fig. 6.9): the North Anatolian Fault Zone (NAFZ, including the North Aegean Trough) and the Hellenic subduction zone (or, 'Hellenic Trench'). The NAFZ was activated during middle to late Miocene times

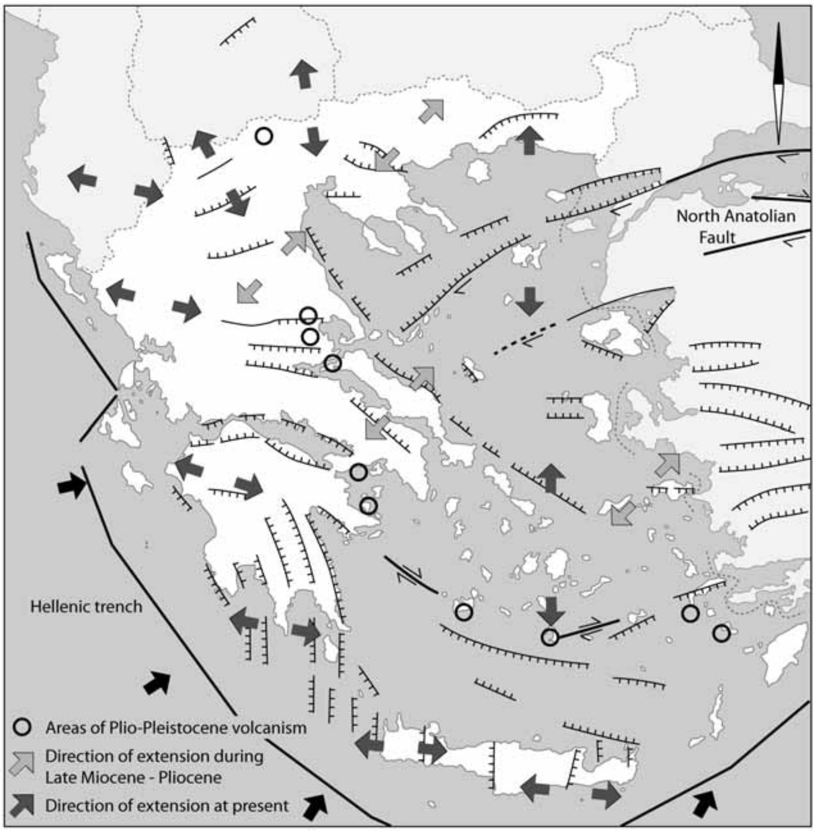

Fig. 6.10 Extensional tectonism and major faults in the Aegean region. Black arrows show the movement of the African plate and the direction of compression in the Ionian Sea. Modified after Mountrakis, university notes

and facilitates westward escape of the Anatolian block, which is driven in response to the northward collision of the Arabian plate into the Eurasian plate (Sengör *et al.* 2005). The Hellenic arc defines the southern boundary of the Aegean extensional domain, along which the African plate is consumed northwards; in most geodynamic models this subduction zone represents the 'free edge' allowing the Aegean lithosphere to either spread or translate the push of the Anatolian microplate (Gautier *et al.* 1999). The onset, spatio-temporal evolution and driving mechanisms of the extensional regime are related to these two structures (NAFZ and Hellenic Trench), with some authors emphasizing the role of the westward push of the Anatolian microplate (*e.g.* Jackson 1994), whereas others underline the subduction roll-back and trench suction along the Hellenic arc (*e.g*. Le Pichon and Angelier 1979), which constituted the outer boundary of the Aegean spreading sheet probably already from the early (Gaultier *et al.* 1999) or middle-late Miocene (Le Pichon and Angelier 1979).

Tensional subsidence of the Thrace basin started during the late Eocene and the oldest extensional basins of the Hellenides were formed in the Rhodope (Burchfiel *et al*. 2003). Early Miocene N-S to NNE-SSW extension shaped the Katerini Basin north of Mt. Olympus and the Klematia-Paramithia half-graben in Epirus (van Hinsbergen 2004). Subsequently, the extensional regime invaded the central and southern Aegean, reaching the latter by the late Miocene (around 11 Ma); by that time, E-W extension and increased subduction affected the south Aegean area, whereas between 15 and 8 Ma compressional deformation during 40º clockwise rotation shaped the western Aegean domain (van Hinsbergen *et al*. 2005; Cloetingh *et al*. 2007). Late Miocene extensional stresses resulted also in the formation of the basins of the North Aegean region, namely those in the areas of Kavala, Xanthi, Komotini and Alexandroupolis (Rondoyanni *et al*. 2004), as well as those associated with the Strymon River (Snel *et al*. 2006; see also below, Fig. 6.11). During late Miocene to Pliocene (*ca*. 8-3.5 Ma), E-W extension accelerated in southern Greece and the entire Aegean-West Anatolian domain experienced a new extensional phase that overprinted the earlier-formed grabens (ibid; van Hinsbergen 2004). Northeast-southwest extension (Fig. 6.10) created the lacustrine Florina-Vegoritis-Ptolemais basin system in Macedonia (Pavlides and Mountrakis 1987) and the intramontane Larissa and Karditsa basins in Thessaly (Caputo *et al*. 1994), whereas a stress field of the same direction impacted also on the basins of eastern Macedonia and Thrace (Rondoyanni *et al*. 2004). Compressional movements affected the Ionian Islands, and the entire western coast of Greece was subjected to strong uplift (Doutsos *et al*. 1987; van Hinsbergen *et al*. 2006). This compressional regime, with folding and thrusting on the Ionian Islands and the western parts of the mainland, results from the convergence between the underthrusting Apulian plate and the Aegean plate, and it led to the formation of taphrogenic basins trending parallel and transverse to the already existing Alpine folds (ibid).

The last phases of landscape development took place from the late Pliocene up to the present, when southwestward motion of Greece continued and most of the seismically active structures that we see today were formed (Jackson *et al.* 1982; van Hinsbergen 2004). Specifically, most of the systems of horsts and grabens that at present divide the modern topography into rising and subsiding blocks were generated during this period, and are best exemplified in central Greece, which displays the classic basin-and-range type of extensional area (Angelier 1978). With internal deformation increasing, the basins were subjected to the strong influence of a N-S extension, whilst the impact of an E-W extensional trend was being superimposed, escalating gradually from north (west) to south(east) (van Hinsbergen *et al*. 2006). This is indicated by the curved form of the main basin systems in central Greece and Peloponnesus, namely that of the Gulf of Patras-Gulf of Corinth-Saronic Gulf, the Ambracian Gulf -Spercheios Basin-Gulf of Euboea, as well as the Pyrgos Basin-Megalopolis Basin-Evrotas Basin system, all of which are marked by N-S extension in the northwest, grading into E-W extension in the southeast (ibid). Recent studies showed that around 2.2 Ma a major change in internal plate dynamics resulted in the majority of active strain within the Aegean-Anatolian plate becoming focused along the Corinth margin (Leeder *et al.* 2008).

During the early Pleistocene, compression was affecting the western marginal parts of the Aegean micro-plate, with the most profound compressive structures occuring in the Ionian Islands (Angelier 1978). This outer compressive regime invaded the Aegean during the early-middle Pleistocene, but the prevailing tectonics were still mostly extensional, although reverse faulting persisted and still continues in the Ionian Sea (ibid). Compressional episodes of smaller amplitude than those prevailing in western Greece and the Hellenic Trench are thought to have interrupted longer periods of extensional tectonics in the central and northern Aegean. However, some authors have argued that these apparently compressional episodes are probably not regional in extent and may not be truly compressional in origin (*e.g*. Jackson *et al.* 1982).

Quaternary palaeogeographic and temporal reconstructions of vertical movements – and particularly the explanation of uplift during extensional conditions[52] – are still debatable (*e.g.* Angelier *et al*. 1982; Moretti *et al*. 2003; Westaway 2007), but during the Plio-Pleistocene/Early Pleistocene a regression probably resulted in the formation of (semi)adjoining landmasses in the Aegean (Mountrakis 1985). During the Middle Pleistocene and while compression continued in western Greece, extensional movements caused widespread depressions which in turn resulted in periodic and eustatic-related transgressions (ibid; Lykousis 2009; see also section 6.4). Although the prevailing tectonic regime remained extensional in the entire course of the Quaternary and up to the present, the early-middle Pleistocene seems to mark local and/or regional changes with regard to tectonic movements, according to various lines of evidence from different parts of Greece. A change in the direction of stretching affects Thessaly, with important implications for the hydrographic balance and the (re) location of the main depocenters (Caputo and Pav-

52. For example, the explanation of differential uplift along the southern coast of the Gulf of Corinth (see below) is still controversial and many authors believe that the cause may be external to the extensional system itself (Rohais *et al.* 2007).

Fig. 6.11 The main Neogene-Quaternary basins of Greece; modified after Mountrakis 1985: figure 81. Note that most of the basins essentially follow the main structural trend (NW-SE) as inherited from the orogenetic axis of the Hellenides. 1) the basins of the Strymonas river, *i.e.* the Serres, Angitis, Strymonikos, Pieria and Akropotamos basins 2) the Axios river-Thessaloniki basin 3) the Florina-Vegoritis-Ptolemais basins 4) the Larissa-Karditsa basins 5) the Pyrgos-Kyllini basins 6) the Argos-Korinthos-Xylokastro basins

lides 1993; Caputo *et al*. 1994; see above 4.6.2). A reorganization of the stress trajectories occurs in the southern Aegean (Angelier *et al*. 1982; Piper and Perissoratis 2003), whereas the north Aegean region experiences also a change in the direction of the stress field from NE-SW during Pliocene-Early Pleistocene to N-S from Middle Pleistocene onwards (Rondoyanni *et al.* 2004). A regional clockwise rotation of 10º that affected western Greece after 4 Ma (van Hinsbergen *et al*. 2005) appears to have taken place in Zakynthos during the last 0.77 kyr and is thought to be associated with concurrent uplift in mainland Greece, caused by rebound processes (Duermeijer *et al.* 1999; but see van Hinsbergen *et al.* 2006). Uplift is, for example, recorded in southwestern Peloponnesus, where from Middle Pleistocene onwards the regime of subsidence gives way to a regime of uplifting blocks (Mariolakos *et al*. 1994; Papanikolaou *et al*. 2007). In the northern coastal areas of the Peloponnesus, uplift became more intense from the mid-Middle Pleistocene onwards, when a third phase of the opening of the Corinth Gulf is thought to be initiated (Moretti *et al*. 2003; Sakellariou *et al.* 2007; see also below). In sum, during the early-middle Pleistocene (after *ca*. 1.0 Ma) there is a change in the direction of extension: in the western part of the Hellenic arc and in central and northern Greece, the tensional stresses shifted from NE-SW to NNW-SSE, whilst the Gulf of Corinth was subjected to a rapid phase of rifting (Schattner 2010, 545). Convergence rates increased at the outer perimeter of the Hellenic arc, probably as a result of a short but intense compressional phase during 1.0-0.7 Ma that prevailed between the extensional periods (ibid). Overall, this change is thought to be related with a major kinematic transition that affected the tectonic regime of the entire eastern Mediterranean region (Schattner 2010).

6.3.3 Geodynamic interpretation and geomorphological consequences

In the Late Pliocene-Pleistocene and during the course of the Quaternary, tensional stresses (re)appeared in many areas, causing further subsidence along new normal faults and/or pre-existing structures that were re-activated, as it is, for instance, the case for the Larissa basin in Thessaly (Caputo and Pavlides 1993) or the Florina-Ptolemais basin complex (Pavlides and Mountrakis 1987). The latter is also a good example of the way in which Quaternary tectonics formed hills and ridges that subdivided the wider basin into several sub-basins, most of which occur now as fault-bounded lakes (ibid). In essence, most of these tectonically controlled basins (grabens) are bounded by still active faults, and they can be viewed -in geomorphological terms- as the key sedimentary basins, which served as depocenters for sediment accumulation. They are morphologically defined by major drainage courses (*e.g.* the Alpheios and Peneios rivers), lakes (*e.g.* the lake Trichonis) and shallow gulfs (*e.g.* Patras and Ambracian gulfs) (Doutsos and Kokkalas 2001; Caputo *et al*. 1994). Consequently, it is not surprising that the vast majority of Pleistocene sediments outcrop and/or are buried inside those basins (Mountrakis 1985). The most important Neogene-Quaternary composite basins of Greece are shown in Fig. 6.11.

Evidently, neotectonic[53] movements have been major controlling agents in the evolution of the present landscape and the distribution, preservation and visibility of the Pleistocene deposits. First and foremost, extensional tectonism dictated the number, structure, spatial extent and distribution of accommodation spaces for Pleistocene sediment accumulation. In contrast to compressional movements and the resulting 'shortening' of the landscape, the stretching of the land during extension favors the formation of such accommodation spaces. Compared to the situation in northern latitudes, where glacial and periglacial processes were both more widespread and more severe, the extensional tectonic regime in Greece facilitated the generation of landforms that were potentially suitable for protecting thick sedimentary sequences from climatically-induced erosional processes, as it is, for instance, manifested by the unique record of Tenaghi Philippon. Yet, records of this kind, with long and relatively undisturbed infills of terrigenous material, are mostly to be found in basins that are now submerged by the sea (*e.g.* on the northern borderlands of the Aegean and offshore Thrace) or occur at present as lakes (*e.g.* Ioannina lake).

53. The term "neotectonics" has not been rigorously defined; following Robertson and Mountrakis (2006), it is used here to refer to stress regimes that essentially remain active at present (broadly from Miocene to Recent in the Eastern Mediterranean region).

Sedimentary basins are the chief means of the medium- to long-term preservation of the changing geological record, and tectonism exerts major controls on rates of sediment discharge through the effects of progressive uplift and subsidence (Leeder 1997), which in turn influence also the visibility of Pleistocene deposits within a sedimentary sequence (Leeder and Jackson 1993). Subsiding areas have been acting as sedimentary receivers, whereas uplifting blocks commonly served as source areas for sediment weathering and transport (ibid). In the case of subsided blocks, sediments may have been locally protected from erosion as long as they remained buried, and especially in stacked sequences; in this case, preservation potential is high, but archaeological visibility is low when sediments are deeply buried. Sediments that were originally deposited in subsiding/subsided blocks, were subsequently buried by younger deposits but at a later stage were subjected to uplift, may offer better visibility/accessibility if uplift-triggered erosion has removed their cover and exposed them; but at the same time they may offer little potential for preservation, depending on the time that has elapsed since their subaerial exposure. In the case of uplifting/uplifted blocks, sediments are undoubtedly more prone to erosional processes; hence they may provide low preservation potential but better visibility, again depending on the *timing*, *duration* and *intensity* of erosion: in case they were once buried, the ideal situation would be when erosion affects only the overlying cover, and archaeologists would like that to occur in relatively recent times, so that there are less possibilities for the target-deposits themselves to be subjected to erosion. Obviously, it all depends on sedimentation rates, their temporal relationship with tectonic movements, and the effects of the prevailing climatic conditions.

As in the rest of the Mediterranean (Macklin *et al.* 1995), the drainage basins in Greece can be divided in two main categories according to topographic settings: (1) steepland fluvial systems above 500 m elevation, mainly involving rivers with steep, cobble-bed channels and high sediment loads and (2) basin and range drainages. Arguably, it is mainly the latter type that offers the potential for the development of depositional landforms (*i.e.* where archaeological finds may have been buried and preserved), in the form of alluvial fans, stacked sequences or extensive

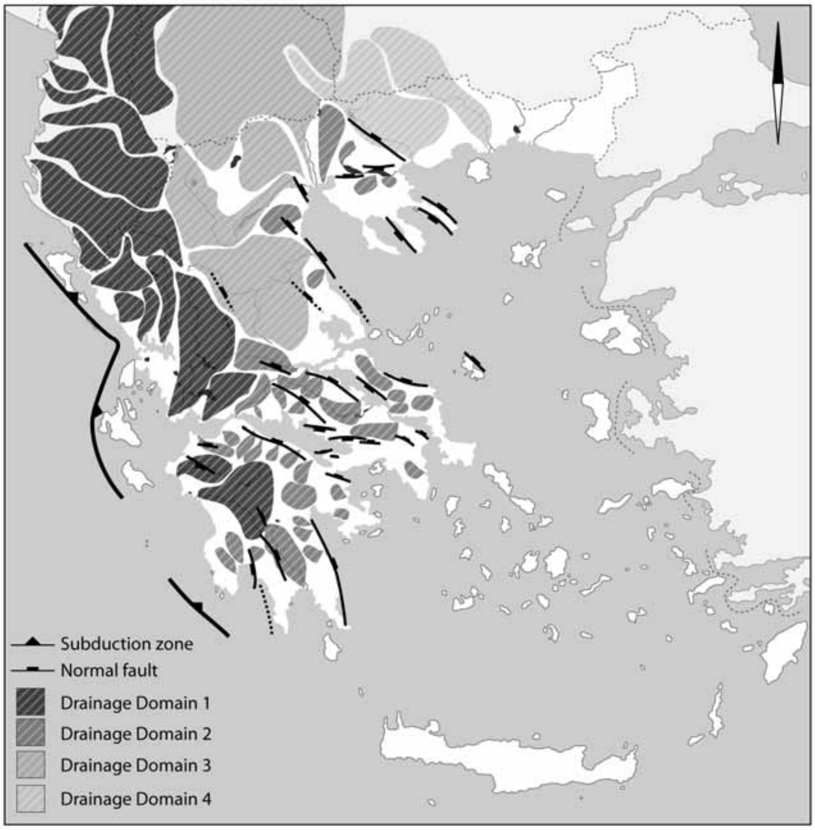

Fig. 6.12 Map of drainage domains in Greece, showing also some of the most important Neogene-Quaternary normal fault systems. The grouping was made according to regions in which drainage basins are broadly consistent in character, in terms of scale and predominant direction of flow. Modified after Collier *et al*. 1995: figure 1.

flights of river terraces. On the other hand, faulting and folding has generally constrained the development of extensive alluvial channel systems (Macklin *et al*. 1995) and Greece is lacking large basins comparable to other Mediterranean countries, such as those of the rivers Ebro (Spain), Rhône (France) or Po (Italy). In Greece, the size of the areas available for the development of drainage basins is controlled by fault spacing, fault overlap and by the tilting that results from extension (Collier *et al*. 1995). Moreover, the growth and form of drainage basins is largely influenced by the slopes, which, in the case of Greece, are often produced by normal faulting (in areas of extensional tectonics), subsequently modified by erosional and depositional processes that in turn relate to rock properties and the emergent relief (Leeder and Jackson 1993 with references therein; see also next section). For example, "the longer the initial tectonic slope, the greater the drainage basin area and hence the larger the alluvial fan area" (Leeder 1991, 456). Overall, footwall uplands exhibit short slopes that are drained by numerous small drainage basins, whilst hanging-wall depressions contain larger slopes drained by larger, usually dendritic basins (Leeder and Jackson 1993).

On a large-scale perspective, the catchments of Greece can be grouped into four drainage domains, approximately corresponding to regions with a particular geological and tectonic history (Fig. 6.12; Collier *et al*. 1995). Drainage domain 1 covers western Greece, where, especially in its westernmost parts, thrusting, reverse fault reactivation of normal fault planes and strike-slip faulting control slope lengths and stream patterns (ibid). Here, Pliocene-Plesitocene compressional movements and the resultant crustal shortening (King and Bailey 1985) restrained the development of large basin systems. Drainage domain 2 consists also of small catchments, but this domain is the most tectonically active in Greece, and here the magnitude of the catchments reflect normal fault mechanisms (*e.g.* fault segmentation and fault spacing) produced by extensional tectonism (Collier *et al.* 1995; see below for particular examples). The third, north-central domain includes some of the largest basins of Greece, like the Mesohellenic molasse

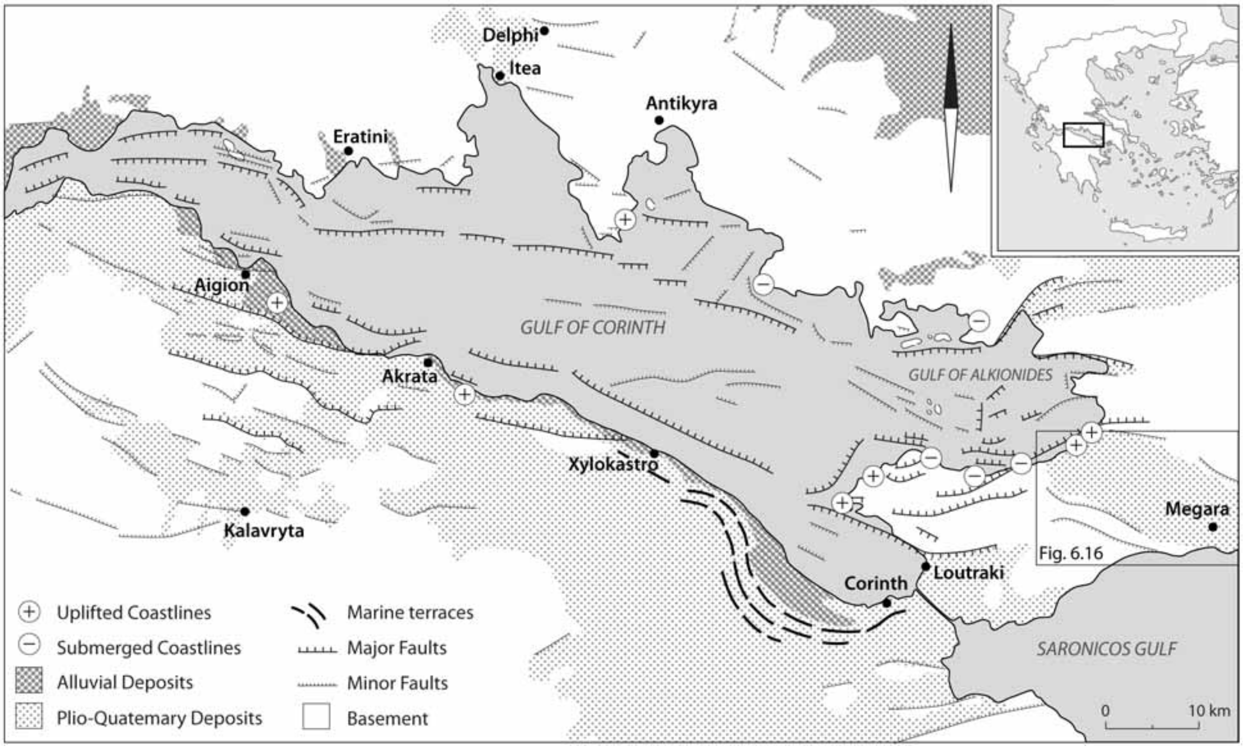

Fig. 6.13 Structural map of the Gulf of Corinth; modified after Moretti *et al*. 2003: fig. 1. Note the plus/ minus symbols indicating uplift/subsidence, and the uplifted marine terraces mentioned in text

trough and the basins of Karditsa and Larissa, which are, again, systems formed and controlled by tensional movements and normal fault geometries (Caputo and Pavlides 1993). The fourth domain occupies the north-eastern part of the country and involves large catchments that follow the axis of extensional basins, mainly parallel to the structural trend of the Hellenides (Collier *et al*. 1995). It has to be noted that the latter two domains, which overall host the largest catchments, experience significantly less tectonic movements than the former two domains, where intense tectonism dictated the small-scale development of basin systems (ibid). As a result, the landscape of Epirus (domain 1) differs markedly from that of Thessaly (domain 3).

It is primarily these differences that are reflected in the divergences between Epirus and Thessaly, with regard to the degrees of preservation and visibility of the Early-Middle Pleistocene deposits and their associated Palaeolithic records. Due to the extensional tectonics that affected Thessaly in two main phases (Pliocene to Early Pleistocene, and Middle Pleistocene to present), Early and Middle Pleistocene sediments are essentially restricted to areas that were uplifted and/or did not experience any subsidence (*i.e.* mainly the so-called Middle Thessalian Hills; Schneider 1968). In contrast, areas like the Larissa and Tyrnavos basins that were subsiding during the two phases, respectively, served as depocenters, in which sediments now are deeply buried. Consequently, uplift of the Middle Thessalian Hills made Early-Middle Pleistocene sediments archaeologically accessible at present (*i.e.* good visibility), but also exposed them to erosional processes and we therefore find them only as isolated patches that are discontinuous and fragmented in both their horizontal and vertical arrangement (*i.e*. low preservation). On the other hand, the story of the Palaeolithic sites in the rugged setting of Epirus is somewhat different. Here, tectonically-formed karst basins called poljes have acted as sediment traps that collected sediments from the surroundings, thereby concealing and preserving artefact scatters (*cf*. section 4.5). In this 100 km-wide, seismically active thrust belt zone, compressional movements were and still are prevailing,

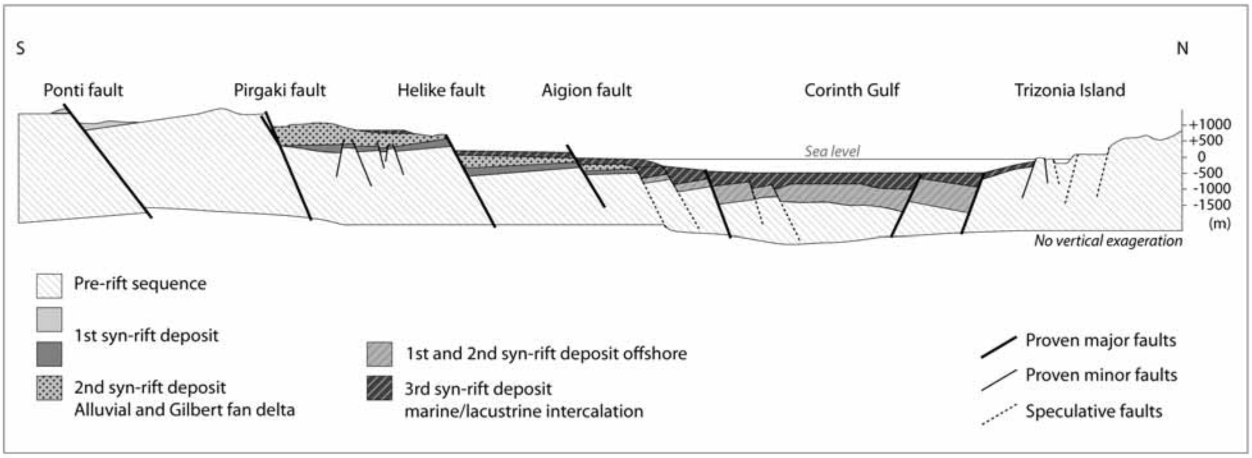

Fig. 6.14 Schematic cross-section of the Gulf of Corinth, The 1st, 2nd and 3rd syn-rift deposits largely correspond to the Lower, Middle and Upper Group of Rohais *et al*. 2007. Modified after Moretti *et al*. 2003: fig. 5

inducing considerably high uplift rates (King and Bailey 1985; King *et al*. 1993). Continuous and/or accumulated uplift has raised many of those enclosed tectonic depressions, forcing rivers to cut back upstream and capture the basins, draining former lakes and exposing the stratigraphy. Then again, whereas uplift, incision and basin dissection altogether assist the ease of site discovery, it also intensifies erosion, deformation and reworking of sediments, thereby enhancing the possibilities for the formation of archaeological palimpsests and/or secondary contexts.

We will now focus on the main tecto-sedimentary mechanisms which condition the above-mentioned interplay between low/high preservation potential and archaeological visibility of basin sediments. These mechanisms are essentially related to basin inversion, when parts of former sediment-receiving basins become uplifted and are turned into exhumed source areas (Mather 2009). In mainland Greece, the general trend of N-S extension is accommodated along E-W trending, sub-parallel normal faults, and within these systems, tectonic activity in each paroxysmic phase usually migrates from one system to another (Goldsworthy and Jackson 2001). When fault activity migrates, a basin that has been formed as the hanging-wall (downthrown block) of an old fault, may now become part of the footwall (upthrown block) of a new fault and hence be subjected to uplift, dissection and erosion (ibid). As I argue later in chapter 7, the timing of uplift and basin inversion is the most crucial factor in assessing the preservation/visibility of the Lower Palaeolithic archaeological record of Greece and it therefore deserves the thorough evaluation that is presented here with regard to specific cases.

Classic examples of fault migration and direct tectonic impact on Quaternary landscapes can be found in the area of the Gulf of Corinth (Mather 2009; Fig. 6.13). This is the most rapidly extending graben system in Greece and the most seismically active zone in Europe, with up to 1.5 mm/year of uplift rate at its southern margin since the late Middle Pleistocene, and a total minimum extension of 5.8 km over the last 1 Ma (Billiris *et al.* 1991; Lykousis *et al.* 2007; Leeder et al. 2008). The syn-rift sedimentary sequence is up to 2.4 Km thick (in the eastern and central part) and the greatest bulk of these sediments dates to the Pleistocene, with much of the sequence being younger than 1.0 Ma[54] (Moretti et al. 2003; Lykousis et al. 2007).

The timing of the initial phase of rifting ('Proto Gulf of Corinth') remains controversial, with proposed ages spanning the Pliocene to early Pleistocene (Ly-

54. The maximum mean sedimentation rate calculated for the depocenter of the central part is 2.7 m kyr^{-1} and the 2.2 km-thick sequence below this part of the basin could have been accumulated within the last 0.8 Ma (Lykousis *et al*. 2007, 48).

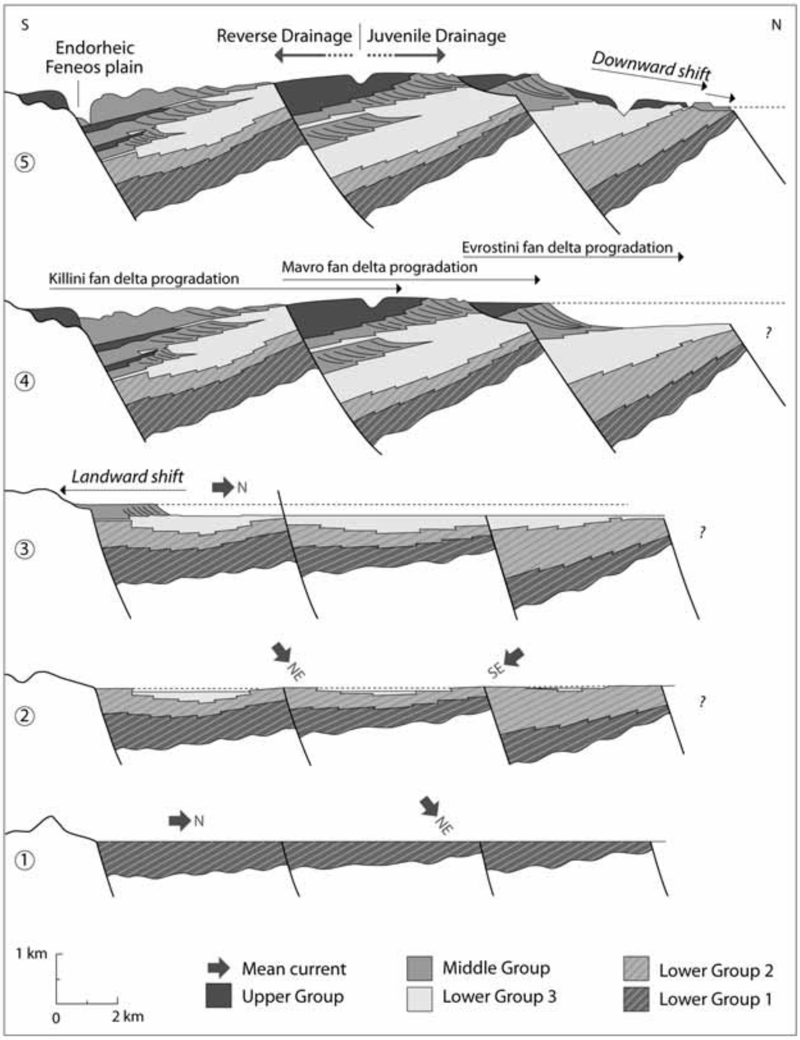

Fig. 6.15 Palaeogeographic and tecto-sedimentary evolution of the Akrata-Derveni area. Vertical exaggeration x4. Modified after Rohais *et al.* 2007: fig. 10

kousis *et al*. 2007; Rohais *et al.* 2007; Leeder *et al*. 2008); during this phase, deposition involved continental and shallow-water deposits (lacustrine-lagoonal, possibly alternating with marine sediments). The second phase of the opening started around 2.2 Ma (Leeder *et al*. 2008) or at *ca*. 1.5-1.0 Ma (Moretti *et al.* 2003; Rohais *et al*. 2007), it was marked by a dramatic increase of basin subsidence versus margin uplift and involved the development of giant Gilbert-type fan deltas along the southern margin (Lykousis *et al*. 2007). A third, most recent phase has been recently suggested, beginning around 350 and 120 ka in the eastern and western sectors respectively, according to the ages of the oldest-dated uplifted marine terraces (Moretti *et al*. 2003; Lykousis *et al*. 2007). Overall, current tectonism mainly involves the rapid uplift of northern Peloponnesus, namely the southern part of the Gulf (ibid). Due to the shallow depth of the Rio-Antirion sill in the western part of the gulf, and probably due to the Corinth Isthmus in the east (Lykousis *et al*. 2007), the gulf has been repeatedly transformed to a lake when the sea level fell below the level of the Rion sill, but these eustatic alterations are better-constrained only for the Late Pleistocene (Perissoratis *et al*. 2000).

The southern side of the gulf lies at the foot of a staircase of parallel, E-W trending normal faults that have been active mostly from the Quaternary to the present (Fig. 6.14; Goldsworthy and Jackson 2001). The marine terraces (see Fig. 6.13), which were cut during late Pleistocene sea-level highstands, are in the hanging-walls of the southern faults but have been uplifted (and thus preserved) for at least the last 300 kyr, as part of the footwalls of the coastal (offshore) faults, due to a basinward (*i.e.* northward) shift of faulting and sedimentation in the course of the Quaternary (ibid). The same phenomenon applies to the Plio-Pleistocene delta deposits, which are also part of these footwalls, reaching elevations of up to 1.75 km asl and thereby indicating considerable uplift (Fig. 6.14; Goldsworthy and Jackson 2001). The generic mechanisms of such a substantial uplift are still debated (*e.g.* Westaway 2007), and there might be also other causes besides footwall uplift (Goldsworthy and Jackson 2001). In any case, what needs to be stressed is the above-mentioned role of tectonism with regard to the exhumation, preservation and visibility of Pleistocene syn-rift deposits. The latter have recently been thoroughly researched in the Akrata-Derveni region, where they are found in continuous outcrops that reach thicknesses of up to 2 km (Rohais *et al*. 2007). Below, the main results of these detailed studies (ibid; Rohais *et al*. 2008) are summarized.

The syn-rift deposits have been divided into three main lithostratigraphic units:

1. The Lower Group is mostly made of fluvio-lacustrine deposits, ranging in age from Late Pliocene to Early Pleistocene (*ca*. 3.6 to 1.5 Ma)
2. The Middle Group involves the above-mentioned, spectacular Gilbert-type fan deltas (see also Dart *et al*. 1994 for extensive descriptions), with subaerial lobes that range from 2 to 6 km in diameter, which are thought to have been formed during *ca*. 1.5 to 0.7 Ma. They mainly include conglomeratic facies up to 1 km-thick, much of which are coarse-grained deposits of high-energy environments, but floodplain fines and other fine-grained facies interpreted as lagoonal/lacustrine products, are also interbedded.
3. The Upper Group is composed of slope deposits, Gilbert-type deltas, stepped uplifted marine terraces along the coastline, as well as various fluvial terraces. This group is chronologically bracketed between *ca*. 700 ka to present.

Whereas the Lower Group records the initial stage of rift opening and the progressive flooding of continental to lacustrine environments (Fig. 6.15: 1-3), the transition of the latter to the Middle Group, which corresponds to the first occurrence of the giant Gilbert-type fan deltas, marks a significant tectonic event and a major transgression, with increased fault displacements and fault-related subsidence providing the necessary accommodation spaces for fan deposition (Fig. 6.15: 3-4). During the accumulation of the Middle Group, the Gilbert-type fan deltas prograded northward into an alternating marine and lacustrine water body, and in several steps, following the basinward migration of the fault activity and the position of the depocenter.

Landward shifts of the shoreline indicate interruptions of those fan-delta progradations, related to changes in the balance between accommodation space and sediment supply. In places, perched river-terrace deposits overlie the syn-rift sediments, recording a multi-phase incision history and inversion of the direction of the palaeo-channels due to the uplift and southward tilting of footwall-blocks. A marked increase in surface uplift of the gulf's southern margin, together with a northward forced regression as the hinterland tilted southward, is documented by the Upper Group sediments (Fig. 6.15: 5). During deposition of the latter, uplifted and abandoned fans of the preceding Group were re-incised, and in some places, red soils developed upon such abandoned fans, indicating periods of relative stability and reduced sedimentation.

To sum up, Early and Middle Pleistocene sediments accumulated mostly (and originally) in the hanging-walls of normal faults, with sedimentation (mainly alluvial fan and delta formation) following the basinward migration of fault activity. While tectonic activity was migrating to the north, those sediments were being uplifted as parts of the footwalls of the newly-activated faults. Uplift resulted in fan abandonment and the initiation of stream dissection either by reversed drainage (due to the tilting of up-thrown blocks) or by new, juvenile drainage systems and headward erosion. Even if the southern faults were/are still active, the coastal faults are moving fast enough to keep their downthrown blocks below sea-level, whilst the southern faults are not (Goldsworthy and Jackson 2001, 492). As sedimentation occurred in pace with active extension, the depositional systems prograded at each stage, when sediment supply overwhelmed the creation of accommodation space (Rohais *et al*. 2007). The vertical stacking of the depositional sequences indicates an overall relative

Fig. 6.16 Oblique view of the Megara basin showing the inactive Megara fault and the active Alepochori fault; the white dashed line marks the drainage divide mentioned in text. Vertical exaggeration x3, Landsat image from NASA World Wind. Inset: (modified after Collier *et al*. 1995: fig. 5) schematic section of the Megara basin from northwest (X) to southeast (Y), indicating the minimum amount of footwall erosion of the last 1.0 myr from the area defined by the drainage divide

base-level rise during deposition that would have been most probably dominated by hanging-wall subsidence (Dart *et al*. 1994, 558).Combined with apparently very high sedimentation rates, this kind of vertical aggradation quickly ensured protection from erosional agents, thereby providing a high degree of preservation. Subsequent (late Middle to Late Pleistocene until present) further uplift, abandonment and dissection facilitated the outcropping of the sediments in the form of large fan-sections, perched palaeo-valleys, and uplifted terrace deposits –in other words, a fairly good degree of visibility at present.

The Neogene Megara basin is today an inverted basin located at the eastern end of the Alkyonides Gulf (Fig. 6.16), and provides another example of how changes in tectonic activity influence the preservation and visibility of early Pleistocene deposits. Here, the sedimentary infill contains about 1.2 km of alluvial, fluvial and lacustrine deposits; most of the sequence dates to the Pliocene, with Pleistocene sediments being restricted to less than 80 m of the uppermost part of the infill (Leeder *et al*. 2008). When the basin was under extension during the Pliocene, it was the northeast-bounding Megara fault that controlled the contemporaneous sedimentary infill, imposing a dip of the sediments towards the northeast, at the hanging-wall of the fault (Goldsworthy and Jackson 2000). At around 1.0 Ma this fault died out and the orientation of tectonic activity changed, with activation of the Alepochori fault to the north-west (Leeder and Jackson 1993). The >300 m of footwall uplift along this -still active- fault reversed the previous, NW-draining Pliocene drainage system to the south of the footwall drainage divide, by back-tilting the former depositional surface towards the south-east (Collier *et al*. 1995).

If the Megara fault was still active after 1.0 Ma, subsidence of the basin would force the rivers to aggrade in order to reach a graded profile. Instead, due to the inactivation of the Megara fault and the activation of the Alepochori fault, uplift (of the Alepochori footwall) forced rivers to incise and erode the sedimentary infill. Stream patterns that were established in the uplifted block since 1.0 Ma have been mainly controlled by rock type: catchments draining the northern, active front of the footwall are small and steep (mean slopes 12-22º) due to the more resistant limestone and serpentinite bedrock; whereas larger, dendritic drainage systems have developed in the southern, back-tilted and poorly consolidated Pliocene and Pleistocene sediments, forming deeply incised gorges and badland-type terrains (Leeder *et al.* 2002). In the latter catchments, vertical cliff retreat predominates due to rockfalls, and undercutting ephemeral streams subsequently remove the debris (ibid). Such rockfalls and slides at the headcuts of deep gorges have promoted a high rate of divide retreat (Leeder and Jackson 1993), which in turn reflects also a high amount of catchment erosion. In fact, it has been estimated that around 9 km^3 of footwall sediments have been eroded, an amount that would be translated to a 180 m-thick layer deposited offshore over the Alkyonides Gulf (Mather 2009). It is thus not surprising that recent studies assessed a cumulative vertical throw of 1100 m between the subsiding Alkyonides basin depocenter and the uplifting Megara basin; notably, this total displacement probably took place over the period of the last 400-450 kyr (Sakellariou *et al*. 2007).

In sum, the change in tectonic activity around 1.0 Ma resulted in uplift of the basin, basin inversion, drainage diversion and high erosion rates from incising streams, as the depocenter moved out of the Megara basin and into the Alkyonides Gulf (offshore). We can therefore conclude that only sediments deposited in the basin *before* 1.0 Ma may still be locally preserved, potentially offering also a good degree of visibility where streams incise and expose them; on the other hand, after around 1.0 Ma, and especially after *ca.* 450 ka, erosion would predominate over sedimentation and hence it is very unlikely for sediments of that time-span to have been preserved. Indeed, an ash layer located at *ca*. 100 m below the surface and dated at ~2.8 Ma, and a calcrete member that caps the entire sequence and dates to around 0.77 Ma, altogether indicate that the topmost *ca*. 80 m of sediments (*i.e*. above the tuff and below the calcrete) are of Late Pliocene to Early Pleistocene age (Leeder *et al*. 2008)[55].

6.3.4 Discussion

The case-studies above illustrate how tectonism controls syn- and post-sedimentary processes that in turn influence preservation and visibility of sediments within sedimentary basins. Erosion, deposition and re-sedimentation of the sedimentary infill basically depends on the structure and evolution of drainage catchments, hence a focus was given to the manner in which tectonic forces determine the type, distribution and development of drainages. All of the examples outlined above show that basins can change from substantial subsidence to considerable uplift, erosion and re-deposition, all within one million years. This raises important implications with regard to preservation biases upon an archaeological record that has been produced within such a highly active tectonic setting. Temporal and spatial changes in tectonic activity and their derivatives, such as fault migration and basin inversion, were stressed in this regard, and yet there are numerous additional examples from Greek sites that reveal equally important tectonic signatures on landscape modification: among others, these would include changes in fault orientation, fault segmentation, block rotations, differential extension rates along strike, transfer-fault activity, relief rejuvenation and base-level control (Goldsworthy and Jackson 2000, 2001; Goldsworthy *et al.* 2002; Leeder 1991; Leeder and Jackson 1993; Mather 2009; Caputo *et al*. 1994; Macklin *et al*. 1995).

Due to the limited space available here, it was not possible to elaborate on another major tectonic agent of deformation and landscape instability, namely earthquakes, which constitute the ground surface expressions of tectonic movements that occur at depth. Earthquake-related deformation can result in surface faulting producing vertical (uplift and subsidence) and/or horizontal offsets of the surface (Stiros 2009). Seismicity is undeniably a crucial element of landscape evolution in Greece: based on an historical record of more than 2,500 years, it has been estimated that the mean occurrence frequency of strong earthquakes equals about one every 1.7 years (Papadopoulos and Kijko 1991, cited in Caputo *et al*. 2006).

Earthquakes are main controlling agents for an array of slope destabilization phenomena and gravitative mass movements, ranging from slow downslope creep and debris flows to rapid sliding of large masses (Ferentinos 1992). For instance, landslide distribution in the rift zone of the Corinth Gulf is consistent with the distribution of normal faults and the presence of tectonically highly sheared and weathered geological formations, whilst Neogene and Quaternary fine-grained and largely unconsolidated sediments are amongst the most critical landslide-prone formations (Koukis *et al.* 2009). Notably, tectonically-fractured clastic deposits, and sediments

55. In fact, the dating of this newly-discovered ash member may provide a maximum age of 2.15 Ma for the abandonment of the basin (Leeder *et al*. 2008, 136), instead of 1.0 Ma that was estimated by previous studies.

that have been subjected to mass failures in the past, are more likely to experience new gravitational collapses (ibid). Although earthquakes are obvious driving forces in such events, especially in areas with high density of epicenters, excessive rainfall is usually another interacting mechanism that can trigger large displacements (Sabatakakis *et al*. 2005). The role of precipitation (duration, intensity, recurrence intervals, deluges, etc) was examined in the previous chapter, and the role of lithology and rock properties is assessed in the following chapter. Besides the relations mentioned above with regard to tectonically-exerted controls, suffice it here to mention briefly other tectonic features that influence slope (in)stability, such as the presence of foliation, schistocity and cleavages resulting from moderate tectonic events, the presence of faults, folds, discontinuities and highly-fractured zones related to strong tectonism, and the occurrence of imbrications and overthrusts associated with even more intense tectonic movements (Rozos *et al.* 2008).

6.4 SEA-LEVEL CHANGES

6.4.1 Introduction

Two handaxes and a bifacial artefact, all attributed to the Lower Palaeolithic, were discovered in 1995 at a depth of 8 m below present sea-level offshore from the coast of Table Bay, South Africa (Werz and Flemming 2001), thereby providing palpable evidence for the importance of submerged landscapes and the potential for both preservation and recovery of maritime Lower Palaeolithic archaeological records (*cf.* Hosfield 2007; Bailey and Flemming 2008). Until recently, the prevailing notion in the archaeological discourse maintained that early hominins were unable to adequately exploit marine resources, while coastal adaptations were considered as peripheral in human evolution, essentially restricted to the times of the Upper Palaeolithic and onwards (*e.g*. Erlandson 2001; Bailey 2004; Westley and Dix 2006; Fa 2008; Bicho and Haws 2008). As new data progressively come to light, the earliest evidence for hominin use of marine resources is increasingly pushed back, to now include the Middle and even the Early Pleistocene (*e.g*. Marean *et a*l. 2007; Choi and Driwantoro 2007). In this light, the interest in coastal landscapes has been lately renewed and this is justified by studies that reveal the significance of those environments, as:

1. Areas of wider archaeological and palaeoanthropological implications, as potential sources of evidence for subsistence strategies and adaptations that might have prompted behavioral innovations (Erlandson 2001), for example with regard to sea crossings and the origins of seafaring (Anderson *et al*. 2010) or the use of marine resources mentioned above.
2. Potential refugia during periods of increased climatic stress, for plants (Médail and Diadema 2009; Rodríguez-Sánchez *et al.* 2008), animals (Hewitt 2000; Sommer and Nadachowski 2006) and hence probably also hominins (Carrión *et al*. 2008; Finlayson 2008), as areas with milder climates due to marine conditions and with valuable environmental resources, thus acquiring the status of super-ecotones (transitional ecological zones between terrestrial and marine environments), where a rich and diverse mosaic of habitats coexists over short distances (Bailey *et al*. 2008), providing water and food resources (Faure *et al*. 2002), or even shells for tool-production (Szabó *et al*. 2007).
3. Corridors of population movements (see *e.g*. Stringer 2000b; Derricourt 2005; Oppenheimer 2009).

Greece has the longest coastline in the Mediterranean, consisting of 13,780 km and including 6,000 islands and islets that make up around half of the country's coastline (http://ec.europa.eu/maritimeaffairs). It thus becomes obvious that a great component of geomorphological and geoarchaeological processes in the Greek territory has been critically influenced by sea-level fluctuations. While the palaeogeographic and palaeoenvironmental role of the coastal zones and the Aegean and Ionian Islands is also briefly discussed, the emphasis in this chapter is put on the geomorphological biases that the marine control has exerted upon the Pleistocene geoarchaeological record.

6.4.2 Sea-level changes: contributing factors and complicating perplexities

The level of the sea-surface relative to the adjacent land changes over time basically due to interactions and feedbacks between climatic fluctuations, oscillations in the gravitational potential of the earth-ice-water system, as well as tectonic movements[56] (Lambeck 1996). Therefore, in order to understand, measure and model sea-level fluctuations, three main contributions have to be taken into account (ibid; Lambeck 2004):

1. *eustatic sea-level contribution*. As the ice sheets wax and wane, the volume of the water in the oceans decreases or increases accordingly, resulting in rising or falling sea-levels
2. *isostatic contribution*. (i) glacio-isostatic; the loading of the sea floor with meltwater from the decaying ice-sheets, or the removal of water-load when water is trapped as ice, deforms the surface of the planet, stresses the mantle and forces the crust to respond to this (de)compression, as its buoyancy is decreased/increased. This process, related to crustal rebound, is known as isostatic compensation. (ii) hydro-isostatic; because of this redistribution of ice-water mass and the deformation of the earth in response to mantle flow and surface deflection, the gravity field of the earth is changing, in turn modifying sea-levels: when ice sheets grow, gravitational attraction of the ice pulls the water towards the expanding ice dome and sea-levels rise near the growing ice margin, whilst they fall far from the ice. The total effect of (i) and (ii) is called glacio-hydro-isostasy.
3. *tectonic contribution*. This refers to tectonically active areas, where vertical tectonic movements are important.

The relative importance of each of the contributing factors is space- and time-dependent: for instance, sea-level changes during the last glacial cycle are primarily the results of changes in high-latitude continent-based ice volume, whilst present-day oscillations on centennial or annual time-scales are mostly of atmospheric and oceanographic origin (Lambeck 2004). In Greece, the predicted LGM sea-level for Thrace (northernmost area) differs from the predicted value for Crete (southernmost area) by about 20 m, because the former is a continental margin site, whereas the latter lies further from the ice sheet and here the hydro-isostatic component is that of an island (Lambeck 1996). Complexities are not restricted only to the modeling of these contributions, *e.g.* when assessing crustal rheology (for the earth-model parameters), ice-sheet dimensions, advance and retreat (for the ice-model parameters), or tectonic histories; but include also the accuracy and the interpretation of the observational evidence, namely the geological indicators (*e.g.* age-height relations of marine solution notches, depths of submarine sediments and shorelines, beach deposits, marine terraces, sea caves and speleothems), biological indicators (*e.g.* coral reefs and mollusks) and archaeological indicators (*e.g.* submerged sites) (Fairbanks 1989; van Andel 1989; Flemming 1998; Bard *et al.* 1990, 2002; Lambeck 1996, 2004; Lambeck and Purcell 2005; Caputo 2007). Together with the three corrective terms (glacio- and hydro-isostasy and tectonic history), the latter indicators are used to calibrate sea-level curves when they are extrapolated from δ^{18}O-curves, because sea-level fluctuations are not a simple function of orbital forcing (Bard *et al.* 2002), the isotopic composition of ice-sheets varies with their size and latitudinal position (*e.g.* Clark and Mix 2002), and marine isotope values cannot be linearly converted into sea-level values (Bintanja *et al.* 2005). Thus, the estimation that a variation of 0.1‰ of the δ^{18}O corresponds to a variation of around 10 m of sea-level position (Shackleton 1987) is reported as commonly accepted (Caputo 2007), whereas a ± 1 ºC of uncertainty in deep-sea temperature variations implies a confidence limit of about ± 30m on inferred sea-levels (Siddall *et al.* 2003). Overall, estimates of the uncertainty in sea-level positions presented in some of the latest-published global sea-level curves are in the range of ± 11 to 13 m (Rohling *et al.* 1998; Waelbroeck *et al.* 2002; Siddall *et al.* 2003; Bintanja *et al.* 2005). In considering such variances and the precision of sea-level predictions from the perspective of an end-user, Caputo (2007) compared various published curves and showed that positions (age and height) differ up to *ca.* 18 kyr and more than 35 m between different reconstructions (note that, as in this example the emphasis was given to major interglacial peaks, even the number of these

56. Needless to say, the processes, records and models related to sea level changes are altogether an issue too complex to be explained and/or reviewed here. The interested reader is thus referred to Lambeck 1995, 1996, 2004; Lambeck and Purcell 2005; Lambeck *et al*. 2002a; as well as the special issue of *Global and Planetary Change*, vol. 66, March 2009.

peaks was shown to differ from curve to curve). The implication of this comparison, as the researcher puts it, is that "if different end-users consider different sea-level curves to analyze the very same area where absolute ages are not available for all recognized inner edges [*i.e.* edges of shore platforms created by stillstands], firstly, different correlations between marine terraces and highstand sea-levels could be obtained, secondly, different ages could be attributed to the same terrace and, thirdly, the estimated uplift rates could vary significantly" (ibid, 422).

Notwithstanding these limitations in the level of accuracy, and after corrections for the isostatic and tectonic effects have been made, regional sea-level and palaeo-shoreline positions can be usefully approximated from global sea-level data "if we make no great demands on spatial or temporal resolution", whilst "the uncertainty of palaeo-shoreline positions even on wide, almost level shelves would not exceed 10% of the width of the shelf" (van Andel 1989, 735). In view of the above and considering also the constraints imposed by the fragmented nature of the geological archive as well as the limits in the currently available geochronological techniques, it is not surprising that high-resolution observations and modeling of past sea-levels exist only for the last glacial cycle (*ca*. 130 ka to present); with the record being better constrained for the period following and including the Last Glacial Maximum (Shackleton 1987; Fairbanks 1989; Lambeck *et al*. 2002b; Siddall *et al.* 2003). It is thus now widely accepted that sea-levels dropped to about 120-135 m below present sea-level (b.p.s.l.) during the LGM (Clark and Mix 2002; Peltier 2002), with subsequent melting of the ice-sheets and rising of sea-levels starting at about 19,000 years ago and approaching their present-day levels at 7,000 years ago (Lambeck *et al.* 2002b). Earlier cycles were most probably characterized by similar fluctuations, at least back to *ca*. 900 ka and most likely back to the beginning of the Pleistocene, but with lower amplitude of variation (Bailey and Flemming 2008). Overall, during the extreme phases of glacial stages, sea-level was 125 ± 12 m b.p.s.l., with less extreme lowstands before 700 ka (Bintanja *et al*. 2005, 126). A record from the Red Sea (Rohling *et al.* 1998) yielded lowstand estimates of 120 (± 8) m b.p.s.l. for MIS 8, 122 to 134 (± 9) m b.p.s.l. for MIS 10, and 139 (± 11) m b.p.s.l. for MIS 12 (the latter confirming Shackleton's (1987) estimation that global ice volume during MIS 12 exceeded LGM values by some 15%). From the same study, the maximum MIS 5 sea-level was estimated at ~6 m above the present level, whereas during MIS 7 sea-level remained below that of the present-day; the MIS 9 highstand was from zero to 15 m above the present level, but the largest sea-level rise of the past 500 kyr followed the MIS 12 lowstand, when levels culminated in a maximum MIS 11 highstand of up to 20 m above the present sea-level (Rohling *et al.* 1998). Although the MIS 12-11 transition involves an extreme glacial (Shackleton 1987) and perhaps an equally 'extreme' -or at least 'enigmatic'- interglacial (see *e.g.* Helmke *et al.* 2008), the rise of the sea-level from the lowstand of MIS 12 to the highstand of MIS 11 implies that ice-volume changes in Quaternary glacial-interglacial cycles were over 30% greater than would be expected on the basis of the last cycle alone (Rohling *et al*. 1998, 165).

6.4.3 Quaternary sea-levels and palaeogeography of Greece

Introduction

A significant number of geological studies have been published in the past few decades concerning the Aegean and Ionian Seas, the majority of which deals with the late Pleistocene and Holocene and involves research into the stratigraphical, sedimentological and palaeogeographical processes occurring on the continental shelf or in major basins, gulfs and river deltas (*e.g*. Stanley and Perissoratis 1977; van Andel and Lianos 1984; Collins *et al*. 1981; Cramp *et al.* 1988; Perissoratis and Mitropoulos 1989; Mascle and Martin 1990; Roussakis *et al*. 2004; Kapsimalis *et al*. 2005; van Andel and Perissoratis 2006; Anastasakis *et al.* 2006; Lykousis *et al.* 1995, 2005; Papanikolaou *et al*. 2007; Poulos 2009; Lykousis 2009). For their own research objectives, all of these studies touch upon sea-level changes, even if sometimes indirectly. Compared to this substantial contribution from earth-sciences, the involvement of relevant research from an archaeological perspective is admittedly small and has been mostly associated with the later periods in prehistory, from the Mesolithic onwards (*e.g*. Kraft *et al.* 1977). Although this picture could reflect research biases, underwater methodological constraints, or the long-lasting pervasiveness of specific archaeological paradigms (*e.g.* the 'broad spectrum revolution'; or, the emphasis on excavating caves versus open-air sites), Pleistocene archaeology has been largely missing from the related investigations –of course with notable exceptions by researchers that have been repeatedly bringing the subject in the discourse (*e.g.* van Andel and Shackleton 1982; van Andel 1989; Bailey 1992; Runnels 1995; Runnels and van Andel 2003; *cf.* Lambeck 1996, 607).

Such a relative neglect stems partly from the difficulty in establishing reliable reconstructions due to the paucity of 'hard evidence': there are no ^{14}C data on former sea-level positions of the latest lowstand to the present level, let alone the direct dating of sea-level markers indicative of glacial-interglacial cycles earlier than the last one (Perissoratis and Conispoliatis 2003). Researchers are thus forced to work primarily with archaeological and geological data that require a lot of corrections and usually provide the starting points for inferences and correlations, which, in turn, are inherently based on presuppositions that cannot be directly confirmed. Archaeological evidence, including chronological data, is numerous, albeit mostly (and most reliably) for the last four to five thousand years (see for example references in Flemming 1998 for publications of site inventories; Kraft *et al*. 1977). Geological data comes in the form of submerged depositional sequences, revealed for instance by borings and gravity cores (*e.g*. van Andel *et al*. 1990b; Roussakis *et al*. 2004), submerged lagoonal or terrestrial sediments and marine-solution notches (Lambeck 1995). An important geological marker is the present position of the Last Interglacial shoreline, because it has long been well-established that sea-levels were at that time near their present level (ibid); hence, wherever this shoreline occurs within a few meters above the present level, it can be inferred that the area has experienced little vertical movements, as it is reported to be locally the case in the Gulfs of Messinia, Lakonia and Argolis (in Peloponnesus; Lambeck 1995). Overall, the identification of this shoreline makes it easier to distinguish between the tectonic and isostatic components, and to correct for the amount and continuity of tectonic rates (*e.g*. Lambeck 1996).

Middle and Late Pleistocene evidence and the first attempts at reconstructions

Van Andel and Shackleton (1982) were the first to publish a reconstruction of the late Quaternary palaeogeography of Greece and the Aegean, elaborating on the archaeological implications with regard to coastal plain resources and migration routes. Notably, they were also the first – and the last – to provide an account of the estimated loss of coastal zone due to the rising sea-level in a given area (in their example, the southeast Argolid; ibid). Later, van Andel and Lianos (1984) were able to pinpoint the sea-level position of the last glacial maximum on the shelf of the southern Argolid by using seismic profiler records: the lowstand level was recognized at a depth of -115 to -118 m as a sedimentary surface of a former subaerial coastal or river plain. Notably, the landward continuation of this LGM alluvial reflector could be traced also on land. This study offered the first 'direct' evidence for the identification and reconstruction of submerged Aegean palaeo-surfaces, foreshadowing the importance of seismic reflection profiles – among other marine geophysical techniques – as a powerful tool for coastal palaeogeography. Regressional or transgressional surfaces of the Late and late Middle Pleistocene have also been identified as unconformities that provide important chronostratigraphic markers. In the Gulf of Gokova (southeastern Aegean, off Turkey), four major erosional unconformities were correlated with the transgressions of MIS 9, 7, 5 and 1, when deltas rapidly retreated landward and little sedimentation took place on the shelf (Uluğ *et al*. 2005). Elsewhere, unconformities that were identified in seismic stratigraphic units indicated subaerial surfaces during glacial lowstands, as it is the case in the North Euboean Gulf (van Andel and Perissoratis 2006). In cores retrieved from South Euboean Gulf and the Gulf of Kavala, such unconformities have been identified as truncated B horizons of soils that were formed when land was exposed during lowstand(s) (Perissoratis and van Andel 1988; 1991). Similar soil horizons have been identified in boreholes onshore at the Argive Plain and on seismic reflectors offshore (van Andel *et al*. 1990b). These Middle and Late Pleistocene paleosols were formed on transgressive marine deposits when the shelf was exposed during glacial (regressive) intervals and they could be traced also

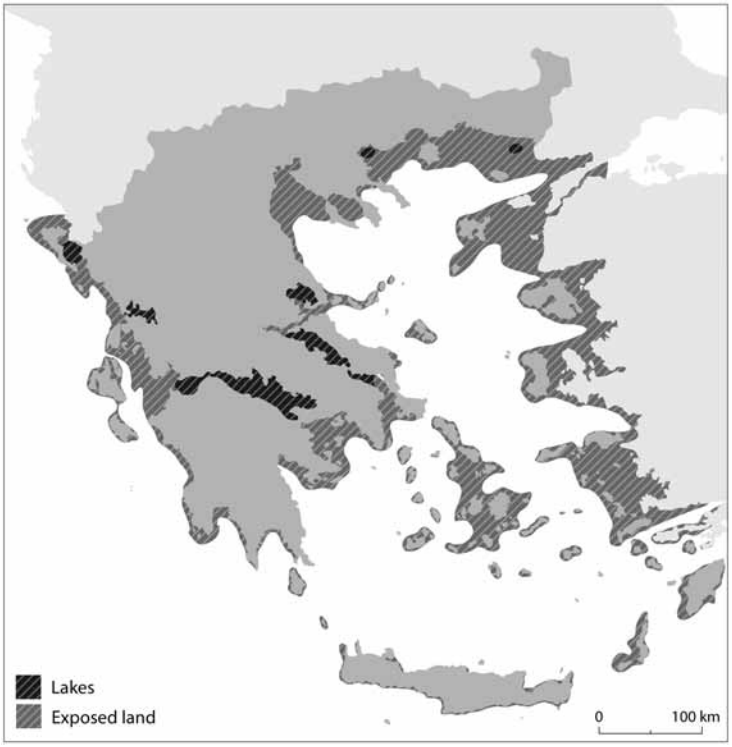

Fig. 6.17 Reconstruction of coastal configuration at the sea-level lowstand of MIS 2 (sea-level at -120 m). Modified after Perissoratis and Conispoliatis 2003: fig. 3

inland (ibid). Their landward extension facilitates correlations of sea-level changes deduced from the marine depositional record with Pleistocene surfaces inland; if they can be dated, they may offer valuable insights on the time-spans during which the shelf was exposed.

The exposure of coastal areas would attain its maximum extent only during glacial maxima and hence for only a few thousand years. Lambeck (1995, 1996) produced the first thoroughly-studied curve of Aegean sea-level change and shoreline evolution for different time-slices of the last 20 kyr, correcting for the glacio-isostatic and tectonic effects. For the period of the last glacial, Lambeck showed that coastal plains were fully exposed from about 20 to *ca*. 16 ka, whereas from *ca*. 70 ka up to the onset of the LGM, sea-levels fluctuated between about -50 and -80 m (1996, 606). Notably, the latter values (from -50 to -80 m) seem to be characteristic for the longest parts of the Middle and Late Pleistocene (Shackleton 1987). Lambeck's results (ibid) were generally confirmed by the study of Perissoratis and Conispoliatis (2003), who used a global eustatic curve for the last 20 kyr, corrected for the glacio-isostatic and tectonic contributions with the aid of sedimentary, bathymetric and seismic profile data (Fig. 6.17).

This map offers a first visual approximation of the maximum extent of exposed coastal areas during the Last Glacial Maximum, and hence it provides also a means of envisaging how much of the (potential) ar-

Depth of Continental Shelf	**Area (km^2)**	**% in relation to mainland Greece**
0 – 50 m	20,159	15.3
50 – 100 m	21,240	16.1
100 – 120 m	7,496	5.7
0 – 100 m	41,399	31.4
0 – 120 m	48,895	37.1

Table 6.1 Estimates of the extent of exposed coastal areas at different depths of lowered sea-level during the last glacial period (see text). The last column on the right shows the percentage of exposed areas when compared with the total extent of mainland Greece, which is 131,957 km^2. Data provided by V. Kapsimalis, personal communication 2009

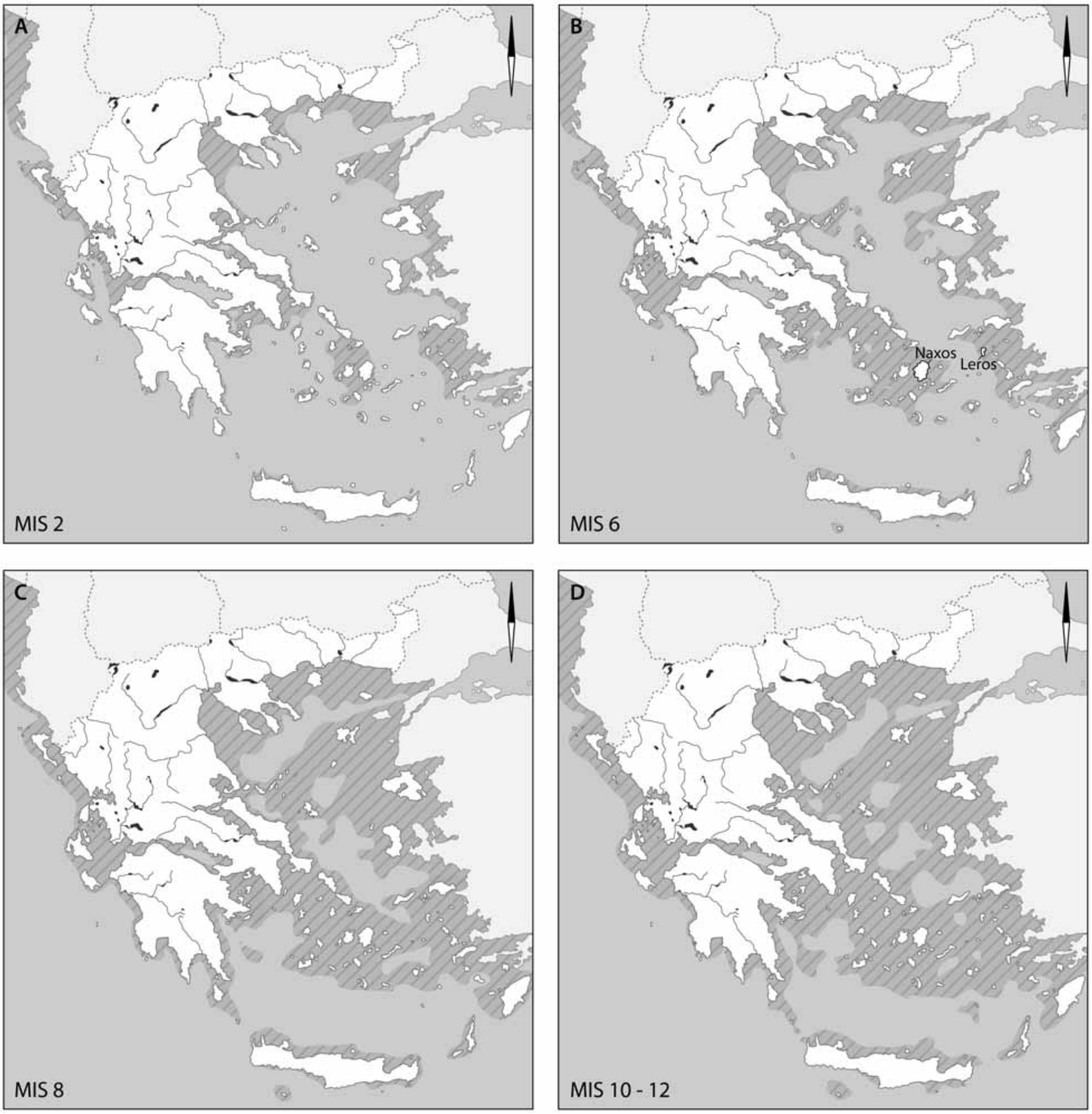

Fig. 6.18 Palaeogeographic reconstruction of the Aegean and the Ionian Sea for MIS 2, 6, 8 and 10-12. Hatched areas denote emerged land. Modified after Lykousis 2009: fig. 5

chaeological record lies submerged since the inundation of those areas. Table 6.1 gives a numerical appreciation for the loss of land during the last glacial period, though considering only the eustatic contribution and without accounting for the glacio-isostatic and tectonic effects. Nevertheless, these values give a fairly close approximation of the true extents of exposed surfaces, being also *minimum* estimates for many of the areas depicted here, as the glacio-isostatic effect would give a correction in the order of only a few meters and because the subsidence that has occurred up to the present has not been taken into account (*cf.* Perissoratis and Conispoliatis 2003, 149-150; Lambeck 1995, 1996).

Area	Sites	MIS 6 - 2 (m ka^{-1})	MIS 8 - 6 (m ka^{-1})	MIS 10 - 8 (m ka^{-1})	MIS 12 - 10 (m ka^{-1})
NORTH AEGEAN	Thermaikos margin	0.86	1.43	1.61	1.88
	Northern margin	0.85	1.31	1.49	
	Southern margin	0.83	1.34	1.46	
CENTRAL AEGEAN	South Limnos margin	0.49	0.67		
	North-west Lesvos	0.53	0.71		
	West Lesvos	0.50	0.74		
	East Skyros	0.45	0.59		
	Eastern Cyclades	0.34	0.57		
	North-eastern Cyclades	0.40	0.55		
	Saronic Gulf	0.42	0.60		
	Corinth Gulf	0.70	1.28		
	Gulf of Patras	0.90			
	Izmir Bay	0.45	0.55		

Table 6.2 Subsidence rates for the Central and North Aegean margins during the last 400 kyr, after Lykousis 2009

Thus, considering -120 m as a mean value for the LGM lowstand (Lambeck 1995, 1996; Perissoratis and Conispoliatis 2003; van Andel and Lianos 1984), we can estimate a total of 48,895 km^2 of exposed coastal areas, or, certainly no less than 41,399 km^2. Even when regarding those as maximum and minimum values respectively, both of them correspond to more than a third of Greece's continental (mainland) coverage. This is admittedly a coarse estimation, but arguably within reasonable confidence limits (*e.g*. see van Andel and Shackleton 1982, 448; van Andel 1989, 735), bearing also in mind the accuracy levels discussed in the beginning of this chapter. The amount of exposed areas remains noteworthy (an equivalent of *ca*. 10% of the extent of the mainland) even when considering the value of -40 m, which would be the level that best defines the coastal palaeogeography of *ca*. 110 to 30 ka (van Andel 1989, 739); and this amount of exposed areas would again be raised appreciably if we consider the levels of -60 to -70 m as better representatives for the most severe stadials of that time-span (ibid).

Lower and Middle Pleistocene Aegean palaeogeography

Notwithstanding the importance of these Late Pleistocene reconstructions, the Aegean palaeogeography changes dramatically in the Middle and Early Pleistocene, when the subaerially exposed areas were three to five times more extensive than those of the Last Glacial Maximum. The key-factor behind this large difference is the greater intensity of the extensional tectonism and the associated subsidence rates that affected the region during the earlier parts of the Pleistocene (Lykousis 2009).

After an early Pleistocene episode of compressional tectonics (Mercier *et al*. 1976), the extensional tectonic regime continued to be prevailing in the Aegean during the Middle and Late Pleistocene (see section 6.3). The subsidence rates that were associated with this latter regime during the last 400 kyr have been recently investigated by high-resolution seismic reflection profiles taken from selected sites in the Aegean (Lykousis 2009). The seismic profiles enabled the identification of successive peak glacial delta prograding sequences (Low System Tracts, 'LST'), which indicate the positions of glacial sea-level lowstands and palaeo-shelf edges with an error margin of less than 10-15 m (ibid). Assuming similar magnitudes for the different sea-level lowstands, and considering that the lack of hiatuses and the conformable development at increasing depths suggests continuous subsidence, the rates of subsidence were estimated by comparing the vertical displacement from the topset-to-foreset transitions (*i.e*. shelf-breaks) of every two successive LST prograding sequences (ibid; Table 6.2). Accordingly, the recognition of those palaeo-shelf break delta deposits together with the derived estimates of subsidence allowed the reconstruction of the Aegean palaeomorphology for the most pronounced glacial periods of the last 400 kyr (Fig. 6.18).

Lykousis notes that "during the isotopic stages 8, 9, 10, 11 and 12, almost 50-60 % of the present Aegean Sea was exposed to subaerial conditions [....] while marine conditions prevailed in the southern Aegean throughout the study period" (ibid, 2041). One first important implication that can be derived from those reconstructions is that, for the studied period (*i.e.* back to *ca*. 500 ka), the emergence of subaerial surfaces was not restricted only to the glacial stages but occurred also during some interglacials, notably those of MIS 9 and 11, but possibly also 7.

The lower subsidence values for the central parts of the Aegean are largely attributed to the relatively milder tectonic activity experienced in those areas, in comparison to the northern and southern parts (Lykousis 2009). Therefore, there are spatial differences in the Aegean subsidence rates (Table 6.2), with the highest values occurring in the seismotectonically active margins of the Gulfs of Patras/Corinth and the North Aegean. Importantly, there are also temporal differences: mean subsidence rates were reduced more than 50% over a period of *ca*. 300 kyr, reflecting a gradual decrease of the mean subduction rates (ibid). This is explained as the result of a gradual decline in the intensity of extensional tectonism and/or a decrease in the isostatic rebound after the early Pleistocene compressional phase in the Aegean domain (ibid).

On account of the results from this research as well as of correlations with evidence from commercial boreholes, Lykousis argues that the first marine transgression in the North Aegean probably occurred during the interglacial period of MIS 9 (Lykousis 2009, 2041). In the central Aegean, the absence of delta prograding sequences older than *ca*. 400 ka probably indicates that *before that time the area was a subaerially exposed land* (ibid). An earlier study by Lykousis and colleagues identified in the South Ikaria basin three delta progradation sequences, correlative to MIS 2, 6 and 8; a possible fourth (and probably a fifth) deformed and eroded deposit was then assumed to be of Early-Middle Pleistocene age (MIS 16-22), indicating "progradation in a shallow marine environment" (1995, 69). In light of the new data discussed above, this latter sequence of the South Ikaria basin should be viewed as probably reflecting lagoon-lacustrine conditions (or, at least not open marine) at some time during the Early Pleistocene. This is most likely because directly to the south of the South Ikaria basin there is an important marine corridor between approximately the islands of Naxos and Leros. As we can see in Fig. 6.18 (B, C) this passage is closed (*i.e.* becomes subaerially exposed land) for the last time in MIS 8. As a result of the 'closure' of this marine corridor, the sea would not penetrate into the central and northern parts of the Aegean prior to about 400-500 ka (MIS 10-12) and all major depressions (basins) would have turned into lakes. Overall, the absence of prograding sequences deeper than those correlated to MIS 10-12, together with other sedimentological and biostratigraphic data, points to the conclusion that *at around MIS 8 and 10 as well as before ca. 500 ka the central and northern Aegean would have been a subaerially exposed land, with shallow water lake environments occupying the Aegean basins* (Lykousis 2009). Certainly, more data are needed in order to securely confirm this assessment and to constrain a lower limit for the time-span that most of the Aegean region was subaerially exposed. Nevertheless, most of the evidence available up to date seems to support this conclusion; yet, the low resolution of the chronological estimates prevents at the moment any substantial qualitative refinement[57] (*cf*. Anastasakis *et al.* 2006; Piper and Perissoratis 2003).

57. As noted above, Lykousis argues that prior to *ca*. 500 ka the central and northern Aegean areas would have been subaerially exposed land, without pinpointing the lower age-limit for the time-span of this land-exposure. His conclusion would be falsified if marine deposits dating to, for example, 700-800 ka were identified in the places where he denotes subaerially exposed land (*e.g.* as in Fig. 6.18: D). To assess this at present is difficult, because: (1) Lykousis' maps are (inevitably) published in small scale (low resolution), thereby hampering comparisons with other published local seismic sections (published in larger scale, better resolution) (2) in lack of absolute dating, regional seismic stratigraphies are often chronologically bracketed by indirect methods: for instance, Piper and Perissoratis (2003) use sedimentation rates with error margins that may be as great as ± 50% on ages >500 ka (3) unconformities are often used as both stratigraphic markers and chronostratigraphic tie-points for local/regional correlations and comparisons, but the time-span of erosion/non-deposition that those unconformities may represent is in most cases ill-defined or indefinable (4) ascribing a precise classificatory term of sedimentary facies to a reflector may be a difficult task, and thus a reflector characterized as "shallow marine" may actually be a lagoon facies.

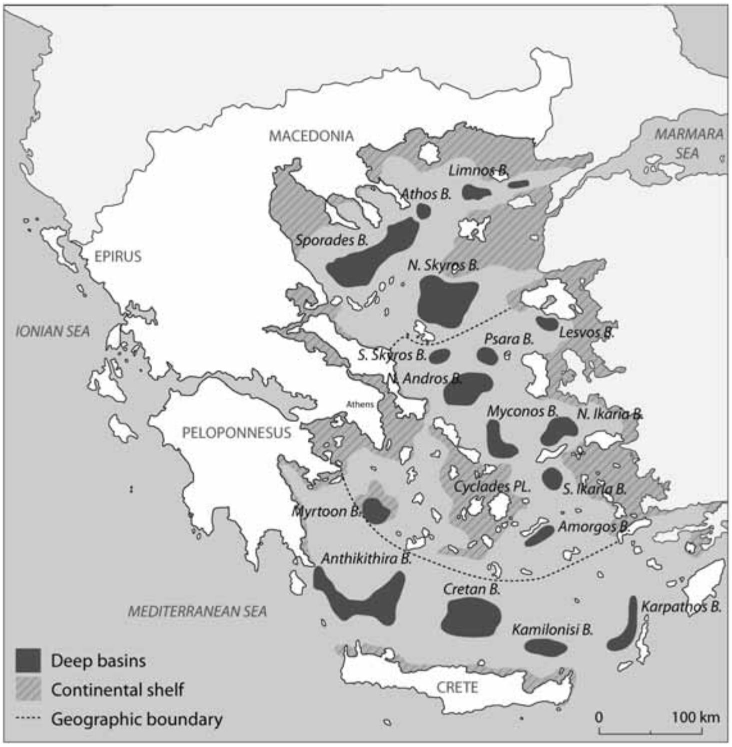

Fig. 6.19 Bathymetry of the Aegean Sea. B = basin, Pl. = Plateau. Dotted lines denote the boundaries of north, central and south Aegean. Modified after Poulos 2009: fig. 1

6.4.4 Prospects of underwater and terrestrial investigations of the Aegean Lower Palaeolithic record

The geomorphological and depositional setting

For both onshore, coastal areas, and offshore, seafloor and sub-bottom deposits, a geoarchaeological prospecting of the most promising locations for retrieving Lower Palaeolithic evidence needs to take into account two critical factors: (1) sediment thicknesses and (2) geomorphological processes related to the construction/destruction of depositional landforms. In order to evaluate both of these parameters, the Aegean region can be assessed on the basis of a geomorphological division into the following main settings: basins and troughs (depressions), plateaus (flat-topped sea floors) and ridges (long, narrow elevations with steep sides and irregular topography) (*cf*. Stanley and Perissoratis 1977).

From a large-scale perspective and on the grounds of the region's geodynamic/structural setting, physiographic/bathymetric morphology and sedimentological composition, the Aegean Sea can be divided in three main sectors (Mascle and Martin 1990; Poulos 2009): the northern, central and southern Aegean (Fig. 6.19). The *North Aegean* sector is characterized by almost aligned and relatively deep depressions, of which the North Aegean Trough is the most prominent and with the thickest sediment accumulation (Mascle and Martin 1990). Thinner sequences that tend to thin landward cover the three major plateaus of Thermaikos, Thasos-Samothraki and Limnos-Imvros, which extend up to a water depth of *ca*. 150 m and do not present any significant internal deformation (Stanley and Perissoratis 1977; Poulos 2009). The main morphostructural unit of the *Central Aegean* domain is the Cyclades Plateau, which is a largely aseismic and shallow region with water depths of less than 250 m. (Mascle and Martin 1990; Lykousis 2001; Poulos 2009). Here, the thickest sections of unconsolidated sediments occur in the basins, whereas sedimentary sequences are relatively thin (*ca*. 250 m) on the two plateaus (Cyclades and Samos-Kos), but more variable on the ridges of Andros-Chios and Amorgos-Leros (Stanley and Perissoratis 1977; Fig. 6.19). The *South Aegean* region includes the deepest depressions in the Aegean. Apart from the inner Argolikos Gulf (depths <600 m), the rest of this domain is composed of separate sub-basins which altogether belong to the Cretan Sea (Poulos 2009).

The continental shelf of the Aegean is wide (*ca*. 25-95 km) along the northern and eastern coasts, where large rivers from the Balkans and Asia Minor have formed broad alluvial plains with thick terrigenous sedimentary sequences that make up a smooth morphology of low slope gradients (Perissoratis and Conispoliatis 2003). Those coastal alluvial plains extend seaward until water depths of about 120-140 m, where a distinct shelf-break occurs (ibid). In contrast, on the coastal zone of Western Greece[58], the eastern coast of Peloponnesus and the fringes of *ca*. 200 islands, the continental shelf is mostly narrow (<10 km) and rocky; in those cases the shelf break is largely controlled by major bounding faults and it occurs at depths of 130-150 m, beyond which very steep slopes (up to 1:20) lead into deep basins (Aksu *et al*. 1995).

Overall, the Aegean region has been intensely fractured due to tectonics, resembling now "a tectonic puzzle made up of relatively small pieces" (Muscle and Martin 1990, 276), with a complex topographical structure and an irregular bathymetry (Stanley and Perissoratis 1977). At places, notably in the central Aegean, depositional trends are not following closely those of topography, and -for example, in the Myrtoon Sea- the relationship between sedimentary/depositional patterns and physiography is not everywhere self-evident, with sedimentary infills varying markedly in terms of facies and thickness even in adjacent basins (ibid; Anastasakis *et al*. 2006). Nevertheless, the distribution of Pliocene/Quaternary deposits generally conforms to topography, so that the thickest sequences occur in depressions, whilst thinner deposits are found on plateaus and ridges and between basins and topographic highs (Stanley and Perissoratis 1977). As with their terrestrial counterparts, the now-submerged depressions and basins that served as sedimentary receivers are expected not only to contain the bulk of the preserved archaeological evidence, but also to carry the best potential for yielding evidence in primary contexts. Yet, since the latter have subsided and are the places with the thickest sedimentary sequences, Early and Middle Pleistocene deposits are practically inaccessible by any contemporary technological means, as they lie in great depths and are usually buried under hundreds of meters of sediments (see further below). Thinner depositional sequences are to be found on plateaus and ridges, and areas with shallower water depths such as the Cyclades Plateau or the continental shelf of northern Aegean should be considered as main targets for future underwater investigations, especially in localities where the Holocene sediment tracts are of minor thickness (*cf*. Perissoratis and Conispoliatis 2003). As in the case of continental Greece (see chapter 6.3), areas that were basinal during Early Pleistocene may have turned into positive features (*i.e*. ridges, horsts) during Middle and Late Pleistocene due to basin inversion (*e.g*. see Piper and Perissoratis 2003), a fact possibly resulting in erosion of overlying deposits and exposure of the deeper (Early and/or Middle Pleistocene) strata. This kind of localities could also be deemed as potential targets, although their occurrence would be rather rare and their recognition difficult.

A geoarchaeological perspective

After pinpointing the most promising areas, submarine explorations need to apply certain oceanographic techniques, such as high-resolution reflection surveying, side-scan sonar surveying, bathymetric mapping, magnetometer surveys, etc (*e.g*. Ballard 2008). Yet, in all probability, these methods help only to identify the appropriate deposits and provide geomorphological indications for places where sites are likely to occur. Remote-sensing techniques are still not able to distinguish cultural (Palaeolithic) remains from natural features, and for this purpose visual methods must be used (ibid). Diving is, then, indispensable, and it commonly needs to be combined with extensive coring (Flemming 1998). Flemming (1985, 1998) reports on Levallois-Mousterian lithic debitage identified by diving in shallow waters (3 to 30 m) about 200 m off the coast of Kerkyra (Ionian Islands). Nonetheless, diving with the aim of identifying Lower Palaeolithic artefact scatters is virtually not yet a feasible research task for most of the Ae-

58. Strictly speaking, the coastal zone of western Greece belongs to the Ionian Sea. Following the researchers cited here, it is included in this discussion for the sake of meaningful comparisons.

gean, because the targeted deposits have been tectonically subsided to great depths. Moreover, erosion and partial or total disturbance of coastal zones and/or submerged depositional landforms should also be kept in mind: submarine mass movements, ranging from debris flows to large landslides, fault-related over-steepening of slopes and subsequent slope instabilities and failures, are altogether a wider and relatively well-documented reality in the Aegean (*e.g*. Lykousis *et al*. 2002, 2009; Xeidakis *et al*. 2007). This kind of phenomena that overall induce destruction or at least fragmentation of the marine geoarchaeological archive would have been accelerated by the alternating transgressive and regressive cycles. Considering also the complex tectonic history of the region (Jolivet and Brun 2010) –a fact that has obscured, biased and hampered Lower Palaeolithic research first and foremost on terrestrial Greece- one would be forced to conclude that the exploration of submerged terrestrial Pleistocene sites in the Aegean is far from being realistic in the near future; let alone the difficulties in convincing funding agencies for investing in surveys with admittedly little prospects for delivering solid results.

In that respect, perhaps the most promising and/or most realistic research objective would be the development of underwater research plans as parts of broader land-based investigations, or alternatively, as the seaward extensions of terrestrial explorations. These can include the regions of wide continental shelf mentioned above, namely the northern and eastern Aegean coastal areas, as well as the shallow-water area of the Cyclades Plateau and associated islands. The Cyclades have already been the target of geoarchaeological explorations, although with a primary focus on the late Upper Palaeolithic and later prehistoric periods (for a recent study and review see Kapsimalis *et al.* 2009). The latest approach (ibid) concluded that the Cyclades Plateau cannot be considered as an area of high archaeological potential with respect to the Middle and Upper Palaeolithic, because preservation of material is not favored due to erosion, and access is impossible due to deep burial. Acknowledging the limitations outlined above, it is believed here that the Cyclades landmass (*i.e*. the C. Plateau *sensu stricto* plus the surrounding islands) should be regarded as an area with an archaeological potential that ought not to be overlooked, for the following main reasons: (1) for a vast amount of time -that may prove to reach even a million years- this was a terrestrial landscape, favorable to be used as not only a landbridge for animal and human migrations between Asia Minor and continental Greece, but also for human occupation *per se* (2) it is a largely aseismic region (Lykousis 2009) and, generally, it lacks major/widespread evidence of submarine failures/instabilities (*e.g*. Piper and Perissoratis 2003) (3) it has relatively shallow water depths (4) geoarchaeological research has already been carried out in various islands of the region, and, even if it involved projects that dealt with the later prehistory, significant knowledge and experience has already been collected (see for example Sampson 2006).

In the last few decades, new research on the Aegean islands has yielded some important contributions to our understanding of human occupation in this region, and, to a large extent, this new data is already in position of addressing crucial questions regarding early human maritime activity and crossings (for a review see Broodbank 2006; Sampson 2006). Instructive examples of a geoarchaeological study can be drawn from the relatively well-attested Middle Palaeolithic evidence from the Sporades (Panagopoulou *et al.* 2001), whereas the finds from Milos, although still inconclusively studied and without a securely datable context, do show the potential for further explorations in the central Aegean (Chelidonio 2001). Artefacts made on quartz and quartzite, including handaxes and cleavers, have recently been reported from Crete: whilst the geoarchaeological study of the associated contexts is still ongoing, a significant number of the finds has been attributed to the Lower Palaeolithic on the basis of typo-technological traits and the preliminary analysis of the geomorphological contexts (Strasser *et al*. 2010). Considering that, according to current knowledge, Crete has been an insular land for most (if not all) of the Pleistocene, a confirmation of the suspected Palaeolithic age of the finds would be a scientific discovery of prodigious significance, especially if the finds prove to date to a period before the appearance of anatomically modern humans. As shown in Fig. 6.18 of the Aegean palaeogeography (Lykousis 2009), before MIS 8, and most notably during and before MIS 10-12 (and for an unknown time-span up to some point in the Early Pleistocene), access to Crete would have been possible, at least during time-windows of lowered sea-level, through two main routes: from the western side, by way of the emerged terrestrial platforms which would have connected the southeastern end of Peloponnesus with Kithira and Antikithira; from the eastern side, by 'island hopping' through Rhodes and the adjoined islands of Karpathos-Kassos. The western route would have entailed a see crossing of less than a nautical mile, which is a distance that does not require the employment of any sophisticated watercraft (the simplest raft or even a natural mat would suffice), whilst it falls well within the documented swimming abilities of most mammals. Consequently, although a tentative *seagoing* across small distances can be credibly assumed for both aforementioned routes, adept *seafaring* needs not be invoked, unless further refinement of the Aegean palaeogeography shows that the marine crossings involved significantly larger distances; or, unless other lines of evidence are able to demonstrate a direct link with the coastal areas of Africa or the Near East. In light of the evidence from Flores (Morwood *et al*. 1998), a modest sea-crossing capacity can indeed be envisaged for hominins as early as the Early Pleistocene, but the Cretan evidence cannot yet neither confirm nor falsify such a scenario for the Mediterranean Basin. Although the artificial character of the specimens appears to be beyond doubt (Strasser *et al*. 2010; personal observation of a sample of the material in 2008), the mechanical properties of the raw material (*i.e*. a bad knapping quality) pose immense difficulties when assessing patterned human action or -even worse- reduction strategies and technological attributes; and this is an extra perplexity that researchers need to resolve, particularly with regard to cases where there is overlap in the use of the Cretan sites in the Palaeolithic and the Mesolithic, since quartz is the dominant raw material used in the latter period, too. Nonetheless, what needs to be further investigated and most convincingly established is the age of the depositional contexts with which the implements are inferred to be associated. The contexts of the Palaeolithic finds are Bt horizons of paleosols, beach conglomerates of marine terraces and debris flow fans; on the basis of the Maturity Stages of the paleosols and age estimates extrapolated from radiometric dates on the lowest Pleistocene marine terraces, an age of *ca*. 130-190 ka is suggested as a *terminus ante quem* for the Palaeolithic artefacts, although the researchers do stress that "these are rough approximations" (Strasser *et al*. 2010, 186). In short, provided that a Pleistocene age is confirmed, the Cretan material illustrates the high potential of Aegean insular sites for yielding important contributions to the unraveling of early human sea-crossing capabilities and -in extent- cognitive abilities, maritime adaptations, and exploitation of marine resources (*e.g*. see discussions in Broodbank 2006; Stringer *et al.* 2008; Joordens *et al*. 2009). Middle and Lower Palaeolithic surface finds have been reported recently from Gavdos, a small island situated 21 nautical miles off the southwestern coast of Crete (Kopaka and Matzanas 2009). Their attribution to such an early age is essentially based on typological/technological characteristics and the degree of surface weathering and patination (ibid). The two implements that are shown in the publication (and reported as of Lower Palaeolithic age), a handaxe and a chopping tool, are made on limestone. Overall, there is no doubt that further research is needed in order to substantiate claims for a Lower Palaeolithic human presence on Gavdos. Last but not least, the Ionian Islands should not be overlooked, as they have already yielded promising results (see above, section 4.4). As part of an extremely active tectonic domain, human habitation on the Ionian Islands (and the preservation of its vestiges) would have been constrained and influenced by a complex tectonic history of emergence and submergence. However, it is likely that for much of the Early and Middle Pleistocene, most of the islands would have been joined to the opposite mainland. The wealth of Palaeolithic finds from mainland Epirus; the relatively well-known Middle Palaeolithic evidence from the Ionian Islands, which includes also artefacts from underwater contexts, and the existing indications for cultural connections with Italian sites, altogether emphasize the encouraging state of research and the potential of this region for yielding Lower Palaeolithic material. Importantly, for the time-spans during which the islands were joined to the mainland, a wide coastal route would emerge, connecting the western parts of Greece with the Albanian and Dalmatian coasts (see Fig. 6.18).

In view of all the above and considering the wider implications of the newly published palaeogeographic maps of the Aegean, what needs to be stressed is that future research in the region needs not be engaged primarily in underwater investigations, as there are numerous Aegean islands that remain virtually unexplored. When results are being produced by terrestrial surveys, then submarine investigations may follow in the offshore vicinities, if they are deemed both feasible and necessary. An example of such a strategy can be given with regard to paleosols, of which the importance has already been stressed here (chapter 4). The association of Palaeolithic finds with paleosols has been attested not only in continental Greece, but also on Aegean islands (*e.g*. Sporades: Panagopoulou *et al.* 2001; Gavdos: Kopaka and Matzanas 2009[59]; Crete: Strasser *et al.* 2010), and it has already been noted that there may be cases where paleosols recognized on land can be traced offshore by acoustic reflection surveys. All in all, geoarchaeological and geomorphological strategies that have already been (or need to be) applied on continental Greece, with regard to site location models, assessments of site/assemblage formation, and ultimately interpretation, can also be employed in research on insular and/or coastal locations, modified according to the constraints imposed by the latter landscapes. From a large-scale geomorphological perspective, we can divide the Aegean into two basic landscape categories: firstly the basins, which serve as 'sink areas' ultimately receiving sediments and material remains; secondly, positive features (ridges, horsts, etc) that essentially serve as 'source areas' from which geo/archaeological material is transported into the basins. The land masses that occur today as islands in the Aegean belong to this latter group: they are the topographic highs which would have always been above the level of the sea, the 'suppliers' of the now-subsided and submerged depressions. As such, they carry much less potential for preserving material, in comparison to the topographic lows of basinal features. Yet, from a finer-scale perspective and by zooming into the physiography of each one of the islands, similar (albeit of smaller-scale) depressions can be identified *on the islands* as well. Consequently, these should be the starting points of any future Lower Palaeolithic investigations in the Aegean domain.

Similarly, localities that have so far yielded palaeontological finds can also serve as points of reference for contextual geoarchaeological assessments, let alone that earlier appraisals of the uncovered assemblages may need to be reconsidered, because at the time of their discovery the palaeogeography of the Aegean was less understood. Together with any possible future improvements in the existing chronological schemes, the acknowledgment of extensive land-bridges between Aegean islands and continental Greece and between the latter with Asia Minor will advance our understanding of issues such as the degree of insular endemism and the presence of unbalanced faunas in many Aegean islands (*e.g*. Dermitzakis and Sondaar 1978). Faunal structure, diversity, migration and dispersal, as well as biogeographical distribution, all depend largely on changes of the physical environment and all are highly relevant to human dispersal and colonization (*e.g.* Koufos *et al*. 2005; Spassov 2002). In that sense, it might be of crucial importance whether the Early and Middle Pleistocene configuration of the Aegean region served more as corridor and/or filter, rather than a sweepstake or pendel route, especially for east-west directed hominin migration/dispersal events and faunal expansions (*cf.* Dermitzakis and Sondaar 1978; Kostopoulos *et al.* 2002; 2007).

A final note on issues of depositional contexts

For the time-period that much of the Aegean was an emerged, terrestrial landscape, we can assume that the now-submerged archaeological material was initially (*i.e*. upon discard) subjected to the same main processes (geomorphological, sedimentological, taphonomic, etc) that apply to continental areas. It follows that alluvial and fluvial sedimentation would have been the prime processes for the potential transportation, burial and preservation (or, instead: erosion and reworking) of human remains and artefacts accumulating on the emerged palaeo-surfaces. In fact, during times of subaerial exposure, sediment river discharge in the Aegean would have been appreciably higher than today's values, especially in adjacent depressions and lakes, close to the regions where prominent rivers debouche large quantities of

59. The researchers note the association of lithic finds with "zones of red soils/*terra rossa*"; my personal observations on Gavdos (2008) can confirm that those "red soils" are paleosols of potentially great maturity.

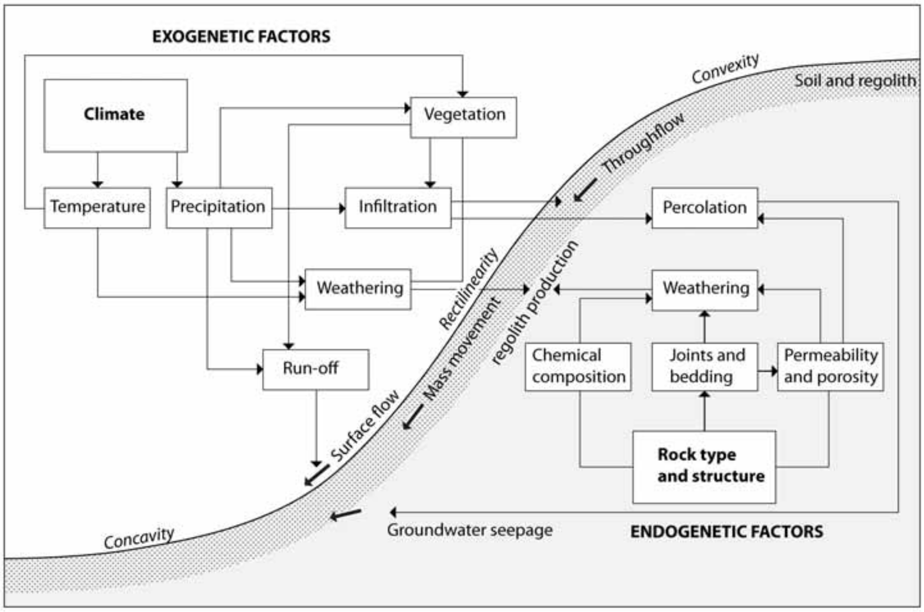

Fig. 6.20 Simplified flow-chart of the slope system with a convex-rectilinear-concave slope as an example. After Clark and Small 1982: fig. 1.1

terrestrial sediments (Lykousis 2009; Stanley and Perissoratis 1977).

A major conclusion that has already been drawn after the examination of the extant Greek Lower Palaeolithic (continental) evidence is that the archaeological record essentially involves findspots associated with secondary contexts. Considering how dynamic the Greek landscapes have been and continue to be, secondary contexts should be expected to dominate the Pleistocene archaeological archive in the future. The now-emerging onshore and -most prominently- offshore Aegean geoarchaeological record will most likely follow this inconvenient trend. Far from being a pessimistic view, dealing with reworked material from low resolution data-sets does not render it useless, as long as solid and coherent approaches can be developed. For instance, with regard to the Lower Palaeolithic record of the English Channel and the North Sea, Hosfield (2007) exemplified possible methodological means and conceptual frameworks that have already been applied in the terrestrial records; with the appropriate adaptations, such analytical tools can also be fruitfully employed in the maritime realm.

6.5 SURFACE PROCESSES

6.5.1 Introduction

In any kind of environments, the archaeological record has been shaped by the interaction and cumulative effects of syn- and post-depositional taphonomic processes induced by cultural and non-cultural agents (Kirkby and Kirkby 1976; Butzer 1982; Schiffer 1987; Nash and Petraglia 1987; Waters and Kuehn 1992; Ward and Larcombe 2003; Fanning *et al.* 2009).

Here, we focus mostly on the role of the natural processes and specifically those that are designated as *slope* or *surface processes*. Exactly because these operate on land surfaces, namely at the interface of the atmosphere with the lithosphere, their mechanisms and their impact on the geoarchaeological resources integrate and transect elements, systems and structures that belong to interconnected domains of the biosphere: climate, tectonic forces, geological formations and topography (Fig. 6.20). Thus, whilst an emphasis is given on the latter realm (topography), this chapter builds upon the previous discussions on climatic, tectonic and sea-level controls, in order to explore aspects of Quaternary landscape evolution that are likely to have biased the preservation and visibility of the Lower Palaeolithic record in Greece.

In Mediterranean environments, erosion and land degradation have long been considered as significant driving factors in landscape evolution (Inbar 1992; Conacher and Sala 1998; Vandekerckhove *et al*. 2000; Geeson *et al*. 2002; Butzer 2005; Thornes 2009; Wainwright 2009). Grove and Rackham (2001, 8-17) convincingly disprove a long-lasting notion of the Mediterranean as a 'ruined' and 'degraded' landscape, which has suffered extensively from human misuse. However, pointing out the problems of the 'degradationist theory' or the confusion related to the terms 'desertification' and 'degradation', does not mean that the authors deny the importance of erosion: rather than presenting evidence to show that erosion "does not matter", their goal is to "demonstrate that it is not simple" (ibid, 267). Importantly, they also note (ibid, 246): "It is a common fallacy that all erosion represents the loss of soil […] Sheet, wind and rill erosion, working from the top downwards, do indeed remove soil, but that is not true of the other kinds. With gullying, sidecutting, marine and karst erosion, most of the sediment removed is not soil". As such a discussion falls outside the scope of this section, for the definitions of these terms and the related debate the reader is referred to the abovementioned work, as well as to Imeson 1986; Romero Diaz *et al*. 1992; Conacher and Sala 1998; Thornes 2002; Bloom 2002; Aksoy and Kavvas 2005; de Vente *et al*. 2008. From an archaeological perspective, what matters is that erosion is inherently related to sediment transport, and that desertification and land degradation promote the erosion of both soils and sediments (hence also the disturbance, reworking or destruction of archaeological material). Under this perspective and in line with numerous researchers, who, although acknowledge the limited use of soil erosion *sensu stricto*, they still discuss erosion in its broadest meaning (*e.g*. de Vente *et al*. 2008), erosion is meant here to include all major eroding processes and agents that operate in slope denudation and landform development (*e.g*. Della Seta *et al*. 2009; Verheijen *et al*. 2009).

6.5.2 Erosion measurement and modeling

Studies of landscape evolution and/or erosion apply surface process models in two broad theoretical and methodological frameworks: *slope processes*, usually expressed by short-range hillslope transport using the diffusion equation (*e.g.* Martin and Church 1997; Jimènez-Hornero *et al*. 2005) and *fluvial processes* (*e.g*. Montgomery and Dietrich 1992; Coulthard *et al*. 2005; Aksoy and Kavvas 2005; Brown *et al*. 2009), both of which are viewed within the overarching setting of a drainage network[60] (*e.g*. Horton 1945; Codilean *et al*. 2006). Factors that need to be parameterized when modeling erosion include lithological properties, soil characteristics, climatic variables, topographic factors (surface roughness, elevation, mean relief, slope angle/length/aspect), runoff attributes, vegetation cover and land use (de Vente and Poesen 2005). Erosion, notably soil erosion, can be measured in four main ways, each of which is commonly associated to a particular spatial scale: (1) change in weight (point scale, 1 m^2) (2) change in surface elevation (hillslope scale, <500 m^2) (3) change in channel cross section (field scale, <1 ha) (4) sediment collection from erosion plots and watersheds (plot scale, <100 m^2 and small watershed scale <50 ha) (Stroosnijder 2005). For larger spatial and temporal scales, catchment-wide erosion can be estimated from the transport of fluvial sediment and its deposition in reservoirs; in this case, erosion is inferred from measurements of sediment yield (*e.g.* Langbein and Schumm 1958; Brown *et al*. 2009). The latter may not be an accurate measure and it most likely underestimates slope erosion mainly because sediment travels intermittently and may be stored in traps within the fluvial system for unknown time-spans (Grove and Rackham 2001; Bloom 2002); moreover, extremely episodic sediment delivery due to catastrophic erosional events will almost certainly be underestimated and/or overlooked during the measuring period (Kirchner *et al*. 2001). Finally, erosion can also be expressed as rates of slope retreat and/or as denudation rates, which estimate the depth of rock/sediment removed from an area in a specified time interval (Bloom 2002). Recently, the use of isotope geochemistry and specifically the application of techniques based on cosmogenic nuclides have revolutionized erosion assessments for

60. Considering drainage basins as three-dimensional landsurface entities, one is forced to identify a continuum rather than a sharp distinction between 'slope' and 'stream' processes (Conacher and Dalrymple 1977). Hence the difficulty in discussing here those processes separately.

Fig. 6.21 Vegetation cover map of Greece. Shaded areas represent forest, shrubs and vegetated pastures; the remaining areas are cultivated and bare lands. After Kosmas *et al*. 1998a: fig. 5.5 (based on the forest map of Greece, Greek Forest Service)

large temporal and spatial scales (*e.g*. Schaller *et al*. 2001; von Blanckenburg 2005). Nevertheless, erosion studies that span 10^3-10^6 yr^{-1} time-scales at equivalent spatial scales (*e.g*. 10^1-10^6 km^2) are overall scarce due to many reasons, if not because they are bounded to be employed within broader frameworks that address long-term landscape evolution with sophisticated conceptual and numerical modeling (see Mather *et al*. 2002 and Balco and Stone 2005 as examples of notable exceptions; and Bishop 2007 for a recent review).

Problems with investigating erosion and deducing widely applicable generalizations from erosion studies -and especially in Mediterranean landscapes- largely arise due to: (1) the complexity of the interplay between the controlling processes and the nonlinear character of their behavior, (2) the heterogeneity of landscape forms (*e.g*. runoff surfaces) and land-use histories (3) the difficulty in the parameterization and modeling of the involved processes, and the problems of scale (de Vente and Poesen 2005; Thornes 2009). The latter issue, namely the intricacies related to the different spatial and temporal scales used when recording, modeling and extrapolating data on erosion, has acquired a great importance and has been addressed repeatedly by various scholars (Phillips 1995; Imeson and Lavee 1998; de Vente and Poesen 2005; Stroosnijder 2005; Heimsath and Ehlers 2005; Codilean *et al*. 2006). Most models, be they conceptual, empirical, or physically-based, have limited applicability outside the range of conditions for which they were developed, and many authors argue that erosion rates measured at one scale cannot be regarded as representatives at another scale level (*e.g*. de Vente and Poesen 2005). Whilst third generation erosion models come to provide valuable solutions to such drawbacks by combining deterministic mechanics and stochastic equations (*e.g*. Sidorchuk 2009), other recent studies show that it *is possible* to extrapolate current erosion measurements over longer time-scales (*e.g*. see Peeters *et al*. 2008). For instance, Phillips (2003) explores the reasons behind the fact that a significant number of studies show general consistency or overlap between shorter-term (*i.e*. modern, or Holocene) and longer-term (Quaternary, 10^5 time-frame) rates of erosion, denudation or fluvial sediment yield.

Largely due to their threshold-dominated nature (Schumm 1979; Vanderkerckhove *et al*. 2000; Bloom 2002), geomorphic systems are typically

Formations	Frequency of landslides (A)	Area (B)	Ratio (A) / (B)	Relative frequency of landslides
Quaternary	16.229	15.87	1.022	0.104
Tertiary	30.212	24.00	1.258	0.128
Flysch	35.581	8.48	4.196	0.428
Transition zone to flysch, cherts, schist-cherts etc	2.996	1.22	2.456	0.250
Limestone	3.621	19.50	0.186	0.019
Phyllites-Schists	8.614	18.35	0.469	0.048
Volcanic rocks	2.746	12.58	0.218	0.022
Total	99.999	100.00	9.805	0.999

Table 6.3 Frequency distribution of landslides in different lithological formations of Greece. After Koukis and Ziourkas 1991: Table 21

non-linear, exhibiting complex behaviors that include dynamical instability and deterministic chaos (Thornes 1985; Phillips 1993). Under such conditions, small and short-lived disturbances may have disproportionately large and long-lived effects, thereby inhibiting predictability of future -and reconstruction of previous- geomorphic responses (Phillips 2006). However, as Phillips puts it, "geomorphic systems are evolutionary in the sense of being path dependent and historically and geographically contingent"; in that sense, they are "governed by a combination of 'global' laws, generalizations and relationships that are largely (if not wholly) independent of time and place, and 'local' place- and/or time-contingent factors" (ibid, 739-740). It is in this respect that the following sections discuss the main aspects of past and present erosional processes, which have contributed in landscape development.

6.5.3 Vegetation

The main relationships between vegetation cover/type, climatic variables (*e.g.* precipitation) and erosion were outlined in section 6.2. Here, suffice it to recall that accelerated erosion dominates in sloping landforms when vegetation cover falls below a value of 40% or even 70% (Macklin *et al*. 1995; Kosmas *et al*. 2000). Nowadays, under interglacial conditions (and largely reflecting anthropogenic interference), about 50% of Greece is bare or cultivated land; total forest area covers about 20% of the land, whereas the rest is covered by shrubs, phrygana and other kinds of maquis vegetation (Fig. 6.21; Kosmas *et al*. 1998a). As discussed earlier, major contractions of tree populations were not restricted only to glacial maxima but occurred also during shorter (and relatively milder) glacial intervals and cold spells, even within interglacial complexes (Tzedakis *et al*. 2006). Reduced and highly seasonal precipitation, disappearance of woodlands and prevalence of open vegetation with herb and steppic taxa of low biomass would have promoted soil erosivity and runoff (Macklin *et al*. 1995; Leeder *et al*. 1998; Roucoux *et al*. 2008, 1392; but see also Rogers and Schumm 1991, and section 6.2).

6.5.4 Lithology

Different rock types will present different erodibility attributes according to, for instance, chemical composition, joint spacing, porosity and permeability (Clark and Small 1982). Yet, slope development is controlled by many other variables capable of masking or outweighing the lithological constraints, so that it is unrealistic to expect that a particular rock type will consistently produce a specific slope-form or erosional behavior (ibid). Nevertheless, a few simple and broad generalizations can still be made. For example, massive, coherent rocks will erode less rapidly than weak, incoherent rocks; the former will tend to produce free faces in a slope, whilst the latter will likely form more rounded, convex-concave slope elements (Clark and Small 1982). Generally, in the case of 'hard rocks' chemical weathering is needed in order to break down the rock to erodible soil or saprolite; whereas with 'soft rocks' mechanical weathering and/or direct erosion by runoff or gravity prevails, without the necessity of soil formation for erosion, even if a soil will most likely be present none the less (Leeder *et al*. 1998). An inspection

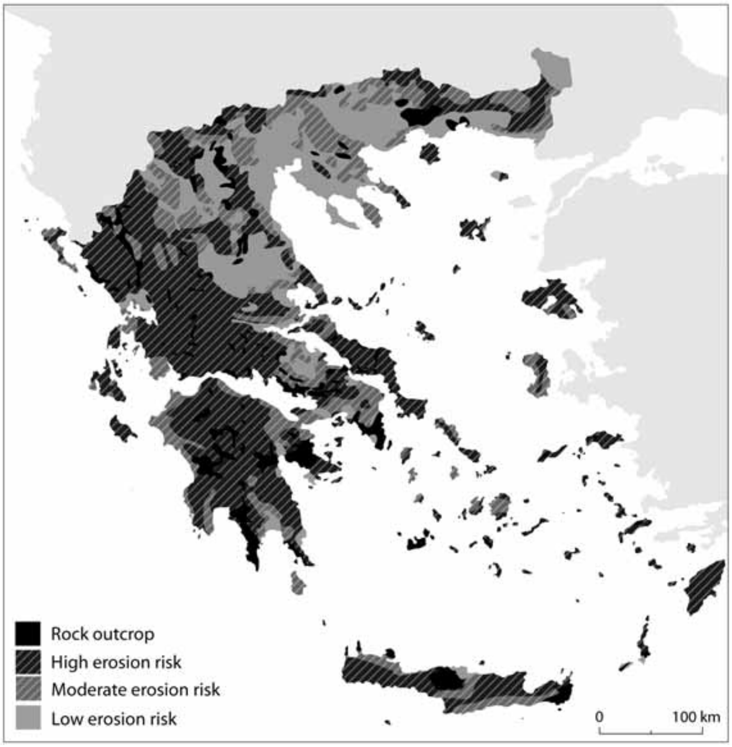

Fig. 6.22 Potential soil erosion risk map of Greece. Modified after Conacher and Sala 1998: fig. 12.3

of the engineering geological map of Greece reveals that the bulk of Quaternary sediments are classified as 'loose', with 'Quaternary cohesive' rocks being a minority. An indication of the erodibility of the superficial, loose Quaternary deposits comes from the frequency distribution of the various types of slope failures in Greece: after rotational slides (slumps) which affect mainly Neogene argillaceous formations, the next most frequent types of movement, namely earthflows and creep, involve Quaternary sediments (Koukis and Ziourgas 1991). Furthermore, statistical investigation of landslide occurrences in different lithological formations demonstrates that Quaternary sediments have the highest frequency distribution after flysch, a transition zone to flysch, and Neogene sediments (Table 6.3; Koukis and Ziourkas 1991).

6.5.5 Soils

Soils on the uplands and mountainous areas of Greece are poor, shallow and deeply eroded, mainly due to the presence of steep slopes and because of the scarcity and/or destruction of natural vegetation cover; in contrast, soils are deeper, better structured and more productive in lowlands (Kosmas *et al.* 1998a). Soil depth has a great effect on vegetation cover and performance, and it increases with humidity (Kosmas *et al.* 2000). The erosion of soil volume capable of supporting vegetative cover is one of the main causes of land degradation in Greece (Fig. 6.22; Conacher and Sala 1998). Decline of soil structure, loss of organic matter and reduction in aggregate stability and salinization are the key agents of soil degradation, which in turn increases the susceptibility to erosion (ibid). Soils formed on hilly Tertiary and Quaternary consolidated landforms typically have a restricted soil depth and limited subsurface layers, such as petrocalcic horizons, gravely and stony layers or bedrock (Kosmas *et al.* 1999b, 27). Extensively degraded areas with skeletal soils are primarily restricted to limestone and secondarily to acid igneous and metamorphic rocks (Conacher and Sala 1998). Particularly limestone formations, which are rather extensive in Greece, produce shallow soils with a relatively dry moisture regime and slow vegetation recovery that makes them highly erodible (Kosmas *et al.* 2000).

When the texture of the parent material is silty and mostly structureless, as with marls, or when the developed soils are thin, as on limestone, then once the topsoil, rich in organic material and well-structured, is removed, the exposed substrate typically becomes subjected to high erodibility in sloping areas, unless mature, strong and cohesive horizons are present (Bk/Bca or K calcic/calcrete horizons). The upper A and E horizons are both unconsolidated and thus it is not surprising that in most of the buried and/or relict Pleistocene paleosols in Greece these horizons are absent: usually, the assumption is that they have been either stripped off by erosion in the past or removed by farming; in some cases, it can be hypothesized that surface lithic artefacts derive from those fragile, now-vanished horizons, or from the erosion-resistant and preserved B and C or K profiles (Pope and van Andel 1984; Demitrack 1986; van Andel *et al*. 1990; van Andel 1998; Karkanas 2002; Runnels and van Andel 2003). For example, the characteristic autochthonous (residual) red soils of Greece (primary *terra rossa*: van Andel 1998) commonly formed on limestones and other calcareous formations, only rarely present a complete Rhodoxeralf profile and in most cases it is merely remnants of their Bt and/or Bk horizons that are preserved (Yassoglou *et al*. 1997). In contrast, allochthonous red soils (van Andel's re-deposited *terra rossa*), much of which are of Tertiary or Quaternary age, are more abundant, exactly because they have formed on lowland, gently sloping or flat landforms, such as lacustrine sediments, alluvial deposits and palaeo-floodplains (ibid). This latter category of *terra rossa* is in itself an indication of past erosion, because the parent materials of those soils have been transported from (mainly) limestone slopes to lower sink-areas (Yassoglou *et al*. 1997). In agreement with previous researchers cited above, my personal observations during fieldwork in Thessaly, Epirus, Macedonia, Peloponnesus and Zakynthos confirm the conclusion that Pleistocene paleosols are almost always to be found best preserved and commonly buried on landforms of low elevations and low gradients (*e.g*. valley bottoms, coastal plains). On upland and high-relief regions, these soils are very rare and are found as eroded, typically truncated remnants; they are preserved only in topographic depressions and/or stable landforms such as plateaus. Soil formation typically requires time-spans of hundreds to thousands of years, and soil development can only be attained in relatively stable landscapes that are neither rapidly aggrading nor eroding (Retallack 2001). Consequently, the relatively thin, spatially discontinuous and patchy, and always truncated Pleistocene paleosols of Greece either indicate relatively short time-windows (in the geological time-scale) of geomorphic stability, or alternatively, they point to past erosional events affecting the most fragile soil horizons.

6.5.6 Land use

The effects of anthropogenic landscape modifications and land use in the biography of erosion and landscape development have long been a primary focus in geoarchaeological, geomorphological and ecosystem management studies in the Mediterranean and in Greece, which is still well-known as a region with a long history of human land use and abuse (van Andel *et al*. 1986, 1990; Runnels 1995; Kosmas *et al*. 1999; Allen 2001; Butzer 2005; Thornes 2009). The long-lasting debate over the human *vs*. climatic impact as major drivers of alluviation and soil erosion during post-glacial times was mentioned briefly earlier and has been thoroughly reviewed and discussed by Bintliff (1992, 2002, 2005), van Andel *et al*. (1990a) and Zangger (1992). Van Andel and Zangger (1990) evaluated a number of geoarchaeological indications from three major projects (in Southern Argolid, Argive Plain and Larissa Basin); they concluded that all three case studies point to landscape instability due to -and after- the spread of farming populations and the onset of woodland clearances, resulting in extensive slope erosion and valley aggradation. Zangger (1992) stressed the time-transgressive nature of this human-induced landscape instability, showing also that such disturbances differ from region to region according to physiographic characteristics and settlement history. All the same, most of these studies entail some recurrent themes with respect to both causal factors and resultant effects. For example, triggering parameters and processes inferred include mainly a combination of the following: the expansion of farming settlements; forest clearance; intensive cultivation and/or over-exploitation of fertile soils; pastoralism and related vegetation disturbance by animal husbandry (grazing, etc); inadequate soil conservation and ineffective or total abandonment of terrace maintenance and gully check-dams; economic and demographic circumstances and policies (Pope and van Andel 1984; van Andel *et al*. 1986; van Andel and Zangger 1990; van Andel et al. 1990a; Zangger 1992). Erosion and destabilization or, alternatively, stability, is inferred from geomorphological, sedimentological and pedological indications from the depositional archives, generally encompassing evidence and conclusions such as: soils indicate periods of stability (*e.g.* Pope and van Andel 1984); erosion events are envisaged as catastrophic sheet erosion on slopes (*e.g.* van Andel *et al.* 1986); gully erosion stripped off soils, which were then deposited in valley bottoms as stream-flood deposits, whilst debris flows have also been feeding many of the Holocene alluvia (Fig. 6.23; van Andel *et al.* 1990a); shifts to instability, signaled either by mass movements (*e.g.* debris flows) or channel aggradation, can be abrupt (Pope and van Andel 1984); the degradation of vegetation by cultural or natural processes (or both) is a general prerequisite in all explanatory schemes (*e.g.* van Andel *et al.* 1990a; James *et al.* 1994; Lespez 2003).

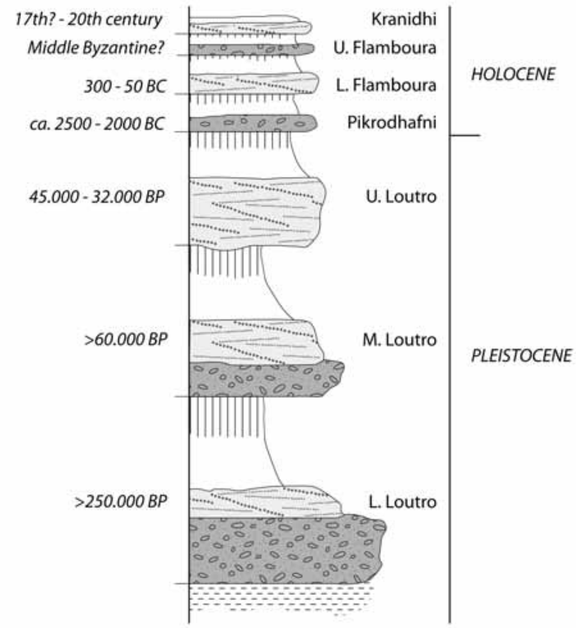

Fig. 6.23 Quaternary alluvial sequence from southern Argolid (Peloponnesus). Cobbles with a chaotic structure indicate debris flows, whilst stratified gravels are streamflood deposits; blank zones are overbank loams and vertical lines denote soil profiles. Although the dates give only approximate ages, they do highlight an apparent discontinuity in alluviation events. Modified after van Andel *et al.* 1986: fig. 4

Another recurrent feature in studies of the relation between land use and erosion is the apprehension of the importance of *slope gradient*. For example, neglect or abandonment of terraces on steep slopes is often thought to have been followed by erosion and downslope sediment redistribution, which in cases results in different artefact densities between steep and gentle slopes (*e.g.* James *et al.* 1994; French and Whitelaw 1999). As expected, slope inclination becomes more significant mainly on upper and middle reaches of the hillslope-channel system (evidenced as *e.g.* debris flows and streamflood deposits in alluvial/colluvial units), whilst it may be less effective in the lower valleys and coastal plains where flooding predominates during erosional episodes (evidenced as *e.g.* overbank loams in floodplain deposits) (van Andel *et al.* 1990a; Harvey 2001). Investigating the implications of polycyclic (prehistoric and historic) terracing in Kythera Island, Krahtopoulou and Frederick report that terrain above the 12º-gradient is preferentially terraced (2008, 559). Moreover, they document that after terrace abandonment, the main types of erosion are sheet wash and gully erosion, the latter affecting mainly unconsolidated formations on steep topographies, occasionally causing almost complete denudation of the slope, which in turn produces significant sediment flux to the local fluvial system. Similarly, an investigation of contemporary erosion on terraced lands showed that sediment loss after abandonment of cultivation increases according to slope gradient, which -in the case of very steep slopes- constitutes the main determinant of erosion irrespective of land use (*i.e.* cultivation, short- or long-time abandonment) (Koulouri and Giourga 2007). Single-storm interill erosion on cultivated soils formed on slopes of 6% to 24% inclination, on Pliocene and Quaternary alluvia in Thessaly, was found to be influenced by surface crusting, whereas sediment delivery was mainly conditioned by rainfall intensity, rain kinetic energy and runoff amount (Dimoyiannis *et al.* 2006). In another study-area, this time in Northern Greece, high degree of erosion was documented for slopes greater than 6% inclinations (Anthopoulou *et al.* 2006). Soil survey data from Lesvos Island show that slope angle has a variable effect on erosion, depending on annual rainfall: in the semi-arid zone, soils on slopes greater than 12% are severely eroded, whereas under the same slope class, the soils of the sub-humid zone were found to be moderately eroded (Kosmas *et al.* 2000). Fields with rain-fed cereals (which is a rather widespread land use in Greece) show the largest values of soil loss in the period from early October to late February, when rainfalls are of high intensity and long duration and the soils are almost bare (Kosmas *et al*. 1997). This kind of studies provide direct empirical support to the decisive role of slope gradient and extreme erosional episodes, such as catastrophic storm events (see also Vandekerckhove *et al*. 1998; López-Vermudez *et al*. 1998; Kosmas *et al*. 1999b; Grove and Rackham 2001, 218-311, and Casana 2008 for examples of similar results from other Mediterranean countries).

Although research on Holocene erosion in Greece continues to provide evidence for a strong correlation between erosional episodes and cultural rather than natural processes (*e.g*. Fuchs 2007), there is at the same time a growing awareness of the driving role that climate, tectonics and geology acquire in preconditioning human-induced erosion (ibid; Allen 2001; Casana 2008; Thornes 2009). Bintliff (2002) suggests that instead of focusing on climate or anthropogenic causation as monocausal or deterministic alternatives, there is much stronger evidence for viewing geomorphic and environmental parameters as setting up a 'pre-adaptation scenario' of a sensitive landscape, where cultural and natural trajectories intersect to create 'windows of opportunity' during which extreme erosion events are likely to occur. This viewpoint is best understood when considering Mediterranean landscape evolution as a "punctuated equilibrium rather than a uniformitarian process of prolonged change" (Bintliff 2002, 418). From this perspective we can evaluate the evidence for prolonged times of stability interrupted by erosional events, the latter being rare and brief but extreme enough to have caused large-scale disruptions that are now manifested in depositional sequences as discontinuous alluvia and colluvia (*e.g*. see Fig. 6.23; Pope and van Andel 1984; van Andel *et al*. 1990a; Bintliff 2002; *cf.* chapter 6.2 and see discussion below).

Anthropogenic slope alteration, agricultural practices, land abandonment or change of land use, all may result in contrasting effects for the ability of soils to resist erosion: in some cases such parameters lead to deterioration of soils whilst in others they tend to improve soil stability (see papers in Conacher and Sala 1998 and in Geeson *et al*. 2002; Grove and Rackham 2001). For example, there is ongoing discussion about the role of terraces in erosion control and whether abandonment of terraced cultivation results in increased erosion or not (Allen 2009). Likewise, terracing affects artefact mobility and site integrity, and terrace construction methods may either impede or improve archaeological preservation and visibility (Krahtopoulou and Frederick 2008). That being said, the general trend is that slope erosion and instability have indeed been accelerated by anthropogenic processes, as the studies mentioned above on prehistoric, historic and contemporary erosion make clear (*cf*. Macklin *et al*. 1995; Wainwright 2009). Dedkov and Moszherin (1992) classified river basins into natural or modified states by using land use type and extent, in an attempt to evaluate the relative importance of natural and human contributions to the sediment yield of mountain river basins across the world. Their results show that present sediment yields (and hence averaged erosion rates) in the Mediterranean have the highest anthropogenic component of any other climatic-vegetation zone, and although the premises of this finding may be questioned, their conclusion is still cited as generally solid (*e.g*. Macklin *et al*. 1995, 16; Thornes *et al*. 2009, 245).

6.5.7 Topography

In simple terms, landscape evolution is geomorphologically expressed mainly by the transfer of mass from source areas to accumulation areas (Cendero and Dramis 1996). In effect, a fundamental contrast in erosional style and magnitude is between flat and steep terrains, with erosion being weaker in the former and stronger in the latter (Stallard 1995; Bloom 2002; Huggett 2003). Thus, physical denudation depends primarily on regional relief and secondarily on runoff and other environmental factors: physical (mechanical) weathering processes, mass wasting and erosion are accelerated by steep relief, especially in areas where orogeny is active (Schumm 1963 cited in Stallard 1995, 20; Bloom 2002, 338), as it is the case for Greece (*e.g*. Reilinger *et al*. 2010). Elaborating further on this assessment, Pinet and Souriau (1988) divide river basins that drain in young orogenies into two parts: below 600 m there is a region of uniform sedimentation, residing in subsiding foreland basins, whereas above 600 m lies a region of erosion. Mountainous regions erode at rates of at least 10 to 100 cm/1000 yr, an order of magnitude more rapid, regardless of climate, than lowlands with typical rates of 1 to 10 cm/1000 yr (Bloom 2002, 338). In other words, denudation rates are generally positively correlated with both altitude and relief: erosion is promoted by greater elevation and steep relief, and deposition is advanced in lower altitudes and on gentle/low relief (Bloom 2002; Phillips 2005; Wainwright 2009).

It is thus not surprising that in many large-scale studies that link tectonics and surface processes, hillslope erosion is modeled *as a function of either elevation or slope* (Montgomery and Brandon 2002 and references therein; Bishop 2007). Ahnert (1970) reported a linear relation between erosion rate and mean local relief for mid-latitude drainage basins. Montgomery and Brandon show that Ahnert's linear correlation holds well "in tectonically inactive low-erosion-rate landscapes, but provides only a lower limit for erosion rates in tectonically active landscapes" (2002, 485). The latter researchers demonstrate the presence of two main types of landscapes in tectonically active settings, with two distinctive geomorphological controls on landscape-scale erosion rates (ibid, 487):

1. low-relief, low-gradient landscapes where erosion rates relate linearly to mean slope or local relief; here, climatic or tectonic controls direct erosion rates through changes in hillslope steepness
2. high-relief, high-gradient landscapes: here, changes in tectonically-driven rock uplift rate influence erosion rates through adjustments in the frequency of slope failure (primarily, landslides).

Effectively, for the second type of landscapes, any increase in mean local relief (*e.g.* due to deglaciation of over-deepened valleys) to values approaching the limiting relief, necessitates a substantial increase in erosion rates (Montgomery and Brandon 2002). The study of Korup and colleagues corroborates these conclusions, at the same time emphasizing that the contribution of catastrophic slope failures to total erosion is not restricted to areas of highest relief and/or strong rocks: hillslope and relief adjustment by means of large events (landslides) occur also in tectonically active areas of moderate relief (*e.g.* at 300-700 m asl, along fault-bounded, mountain range fronts), and include also soft rocks, usually where "discontinuities and low rock-mass strength control slope stability, whether or not threshold slopes have developed" (2007, 589).

Kirchner *et al.* (2001) estimated erosion rates at 10 kyr time-scales in mountainous regions by use of cosmogenic nuclides (^{10}Be), and found a large mismatch when these are compared to conventional sediment yield measurements over years or decades. They interpret this as the result of large, albeit rare and extremely episodic events of sediment flux that are not reflected in the short-term measurements. Importantly, they also demonstrate that the driving factors of episodic erosional events (*e.g.* deluges) must be highly correlated in space (*e.g.* in both large and small catchments). In sum, they conclude that mountainous landscapes exhibit an erosional regime which entails two distinct forms of sediment delivery: incremental erosion, which prevails most of the time but accounts for a small part of the total sediment yield; and catastrophic erosion events that are rare and brief but dominate the long-term erosion rates (Kirchner *et al.* 2001, 593). Further support to the above assessments are given by other studies, which yielded similar qualitative results on the actual and potential destructiveness of such low-frequency large-scale events and the role of landslide phenomena (rock and debris falls and avalanches, debris and earth flows, slides) for landscape evolution in Europe (Cendrero and Dramis 1996). Wainwright, for example, concludes (2009, 190): "It can be argued that major landslides are, at least in the longer term, likely to be the most important landscape-modifying process in the Mediterranean landscape".

6.5.8 Geomorphological opportunities for the preservation of Lower Palaeolithic material: a working hypothesis for the Greek landscapes and the role of topography

As far as relief is concerned, Greece is dominated by mountainous areas with steep slopes (Kosmas *et al.* 1998a). The mountainous parts occupy 64.4% of the country, while lowlands with elevations less than 200 m cover the remaining 34.6%, of which "the real

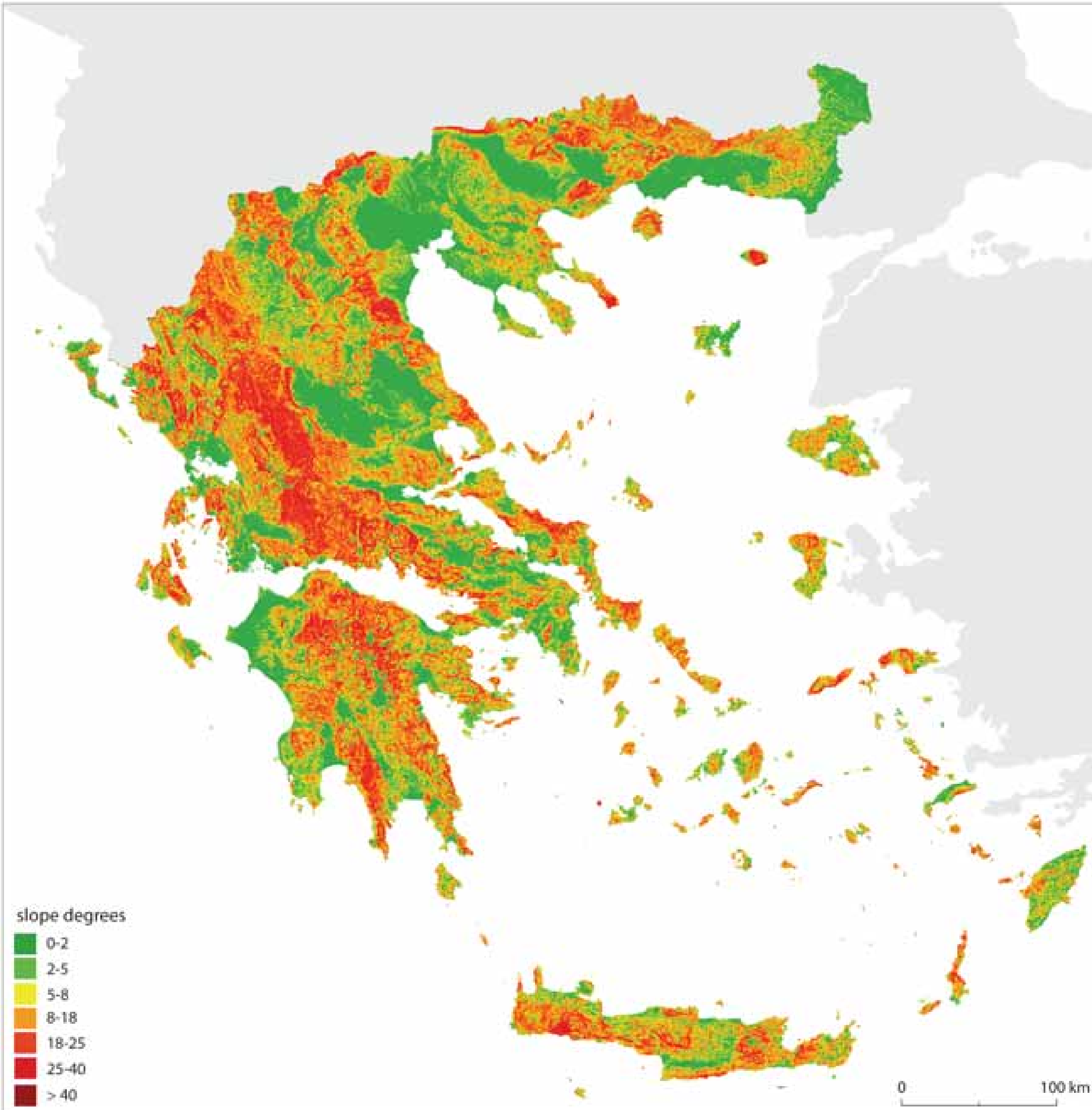

Fig. 6.24 Slope map of Greece. The map was produced in ArchGIS 9.2 and it is based on 1:250,000 topographic maps with a 20-m countour interval

plains reach only 25%" (Koukis and Ziourgas 1991, 48).

As a working hypothesis, I argue here that it is only in this latter 1/4 (or 1/3, at best) part of the country, where Lower Palaeolithic evidence may have survived up to the present in a primary and/or secondary context. The remaining *ca*. 60-70% of hilly-to-mountainous terrain, with the exception of caves and rockshelters, will hardly yield any evidence, and if it does, the material will be either context-less, or from a secondary and -most probably- tertiary context. Obviously, the basic argument behind this assessment is that low gradient favors the persistence of deposi-

Slope class	Slope Angle (degrees)	Area (km^2)	% of total land	Geomorphological classification
1	0 – 2	37411.65	28.3	plain to slightly sloping
2	2 – 5	20938.85	15.9	gently inclined
3	5 – 8	16694.50	12.6	strongly inclined
4	8 – 18	34468.60	26.1	strongly inclined to steep
5	18 – 25	14197.23	10.8	steep
6	25 – 40	8069.57	6.1	steep to precipitous
7	> 40	199.61	0.2	precipitous to vertical
	Total	**131980.01**	**100.00**	

Table 6.4 Numerical attributes of the slope map of Greece

tional landforms, whilst high gradient will be predominated by erosional ones.

Indications that the Greek topography has been and still is unfavorable for the preservation of the geoarchaeological archive were outlined above in relation to vegetation, lithology, soils, topography and land use, whereas the relevant role of climatic, tectonic and sea-level controls have been discussed in the corresponding chapters. Based on the aforementioned relationships between relief/slope and erosion, slope angle is used below as a surrogate for mean local relief and as a proxy for assessing long-term erosion at the landscape-scale, in order to support the argument presented above about the preservation of the Lower Palaeolithic record. In the absence of erosion models at the landscape-scale; and under the premise that "even the limited existing potential for spatial and temporal syntheses should no longer be ignored" (van Andel and Tzedakis 1996), I produced a slope map to use as a morphologic measure for evaluating past and actual biases in archaeological preservation and visibility introduced by erosion (Fig. 6.24). Slope classes, their geomorphological classification and equivalent total spatial extent are given in Table 6.4.

The slope map provides an immediate visual impression of patterns of surface steepness in Greece. Clearly, the areas covered by the first slope class (0º-2º, shown in dark green) mainly represent the sedimentary basins and lowland coastal plains of Greece –the "real plains" of Koukis and Ziourgas (1991), which are also the places where the vast majority of Pleistocene sediments have accumulated (compare Fig. 6.11 with Fig. 6.24). Apart from basins, this slope class (plain to slightly sloping) should also correspond to either smaller-scale, inter- and intramontane topographic depressions, or to plateaus: an example of the former type would be the numerous karst and solution basins of Epirus (dolines, poljes, 'loutses'; see section 4.5), whilst the latter are well exemplified by the plateaus of Macedonia. In all of these settings, deposition prevails over erosion, which, in the case of the larger alluvial plains is essentially restricted to lateral fluvial erosion.

The second slope class represents gently inclined slopes and typically shows a clustering mostly at the fringes of the aforementioned basins/depressions, but also occasionally within them. The Middle Thessalian Hills offer a good example of such a gently sloping landscape (Fig. 6.25, B: 1-3). Those fluvio-lacustrine landforms, dividing the Karditsa and Larissa basins, were either uplifted or did not subside during the Early and Middle Pleistocene, when the adjacent basins were subjected to subsidence (Caputo *et al*. 1994; see section 4.6). This divergence in tectonic history largely explains the difference in morphology and erosional history, and hence also archaeological preservation and visibility: whereas the Middle and Early Pleistocene sediments of the Larissa basin lie now in great depths and are covered by later (Late Pleistocene and Holocene) alluvia, their equivalents on the Middle Hills were always subaerially exposed and thus subjected to erosional processes (see 6.3 above for more details); erosion may have been slow and imperceptible, like in the case of soil creep, but it did accelerate in prehistoric and historic times due to human intervention, today reaching values as much as 1.7 cm of soil removal per year, with rill and interill erosion affecting the cultivated slopes (Kosmas *et al*. 1999b; van Andel *et al*. 1990a; Dimoyannis *et al*. 2006). Other types of landforms that would be fairly

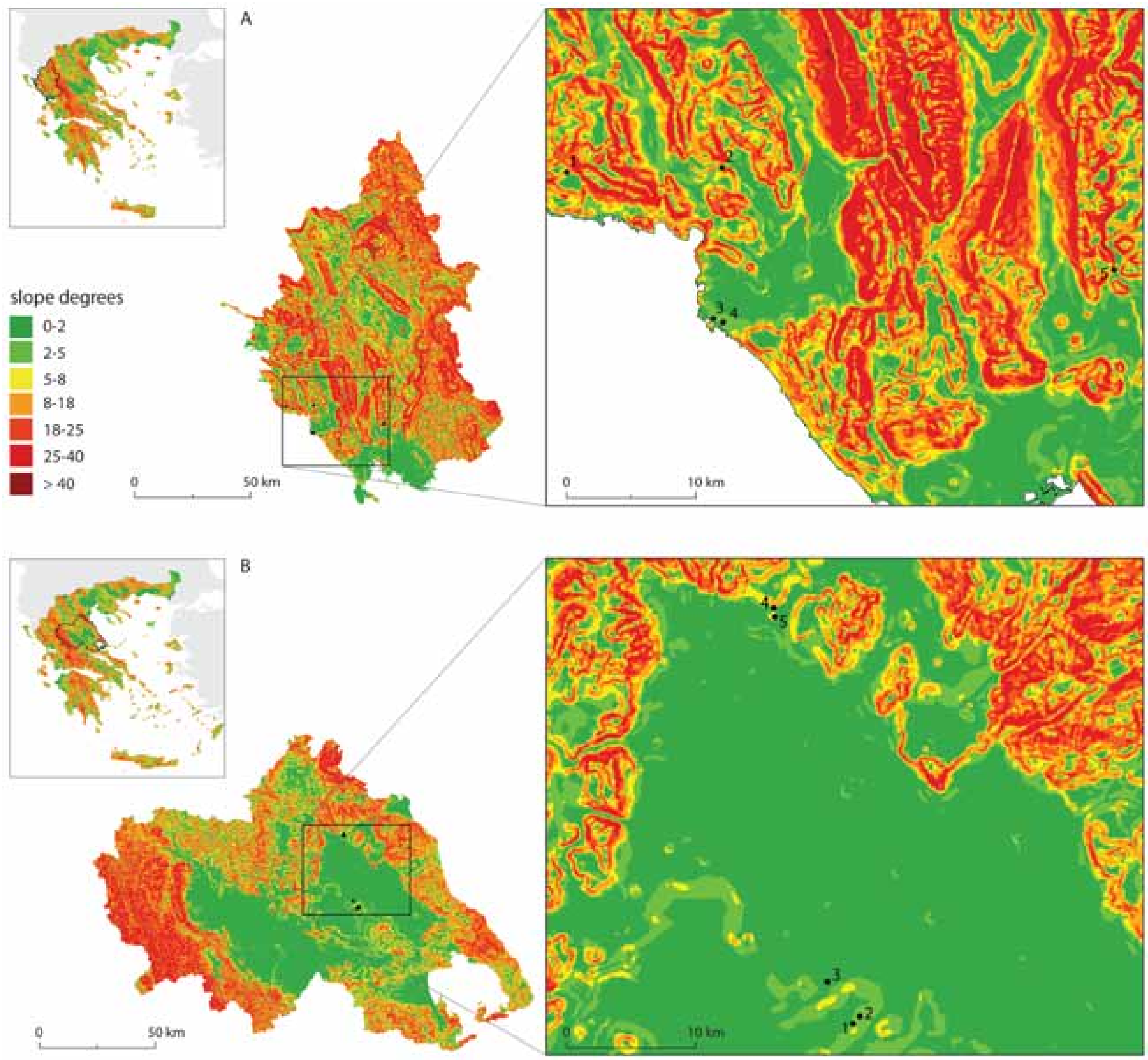

Fig. 6.25 Slope maps of Epirus (A) and Thessaly (B), showing the locations of some archaeological sites and localities with Pleistocene outcrops surveyed by the author and discussed in sections 4.5 and 4.6, respectively. (A): 1) Ayia (see App. I: 28, 29); 2) Morphi (see App. I: 30-32); 3) Ormos Odysseos; 4) Alonaki; 5) Kokkinopilos (most figures in App. I). (B): 1) N. Karyes, outcrop of 'upper Hochterrasse'; 2) N. Karyes, outcrop of 'upper Hochterrasse' in a quarry (see App. II: 6-8); 3) Nikaia, outcrop of 'lower Hochterrasse' (see App. II: 5); 4) Rodia, Kastri Quarry, outcrops of 'upper Hochterrasse' and/or Rodia Formation (see App. II: 10-14) 5) Rodia, site FS 30 (see App. II: 15a)

well-represented by this class involve alluvial and colluvial fans, such as those formed at the fringes of the Thessalian plain (Demitrack 1986), or those of the Argive plain mentioned earlier (Pope and van Andel 1984). Generally, it is exactly because they develop at the interface of a mountainous terrain with a lowland plain or a valley floor that landforms such as alluvial fans present a dual character with respect to resistance to erosion, persistence in time and preservation.

The rest of the slope classes refer to strongly inclined up to almost vertical slopes, which altogether make up the characteristic rugged landscape of Greece.

This mountainous regime is best exemplified by the Pindus mountain chain, which forms the orogenic backbone of Greece with a NW-SE direction and is easily recognized on the map as a belt of red-hued colours (Fig. 6.25, A). From the perspective of geomorphology, the mountainous realm epitomizes the effects of erosional surface processes, and it can be collectively understood as an assemblage of residual landforms, each in the process of being consumed or transformed by erosion. As mentioned previously, this is the high-gradient relief where erosion rates are regulated by the frequency of slope failure. Indeed, studies of engineering geological conditions in Greece with a particular emphasis on landslide distributions show that unstable zones involve mainly areas with either steep slopes and influences by strong deformation stresses in the past, or areas that are still affected by active geodynamic regimes; Central and Western Greece are in that respect the most sensitive regions (Koukis and Ziourkas 1991; *cf.* chapter 6.3). Particularly, Koukis and Ziourkas (ibid, 54) show that landslide frequency in Greece increases linearly with elevation, whilst the frequency distribution according to relief shows that the majority of landslides (65.6%) appear in areas of intense relief.

Yet, how secure is the assumption that the patterns of surface steepness visualized in the map of Fig. 6.24 can be regarded as 'representative' for a time-period that reaches up to the Early Pleistocene? First of all, the initial argument about the topographic control on the preservation of archaeological material is put forth to explore *general trends* dictated by geomorphic conditions, and as such it is not meant to reconstruct palaeotopography, let alone at the hillslope-scale. Nonetheless, we can safely assume, as a general trend, that the mountain chain of Pindus (to use an example) was a mountainous region of high-relief also during the Pleistocene. Similarly, the sedimentary basins were 'always' terrains of low relief during the Pleistocene; in fact, many of them were formed in the Miocene and continued to subside in the Pleistocene: low gradients were in those cases already in place much before the extensional tectonics of the Pleistocene[61] (see section 6.3). In short, at the particular level of generalization used here, if we consider the onset of the Pleistocene as *ab initio* datum for landscape development, then the 'Pleistocene landscapes' inherited a topography in which the Thessalian plain was already a plain and the Pindos mountain range was already in place[62]. Even if we consider that parts of today's basins were not as flat as they appear now but have been flattened in the course of the Pleistocene; and relief in mountainous regions was not as high as it is now but slopes have gradually steepened (during orogenic development); then, the argument above is further reinforced, because in the former case, sediment accumulation (and hence burial of archaeological material) is required for the 'flattening', and erosion/denudation (hence sediment transport and disturbance of archaeological deposits) is assumed for the 'steepening' of mountainous relief. The situation of basin inversions, discussed more thoroughly in section 6.3, is in this respect the most notable exception, wherein a relief that used to be of low gradient in the Early Pleistocene (hence including depositional landforms), has been steepened in the course of time and it is now depicted as a strongly inclined and/or steep terrain. Again, the very same processes inferred (uplift, inversion, drainage diversion or rejuvenation, denudation) exhibit an erosional character that does not undermine the validity of the principal assessment.

Apart from this latter case, and besides situations where a once-low-gradient hillslope may have steepened by tectonic activity (*e.g.* due to a fault), all major and widely used models of slope evolution assume a development where high gradients decline into lower values, and not *vice versa*. In the model of *slope decline*, the upper part of the slope erodes faster than the lower segments of the profile, eventually leading to an overall decline of the slope, whilst in the *slope replacement* model, each slope unit retreats until it is replaced by a lower-angle unit

61. In the geological time-scale and time-span considered, localized and/or sporadic occurrences of steep slopes and/or slope segments *within the sedimentary basins* is regarded here negligible.

62. Landscape evolution can only be viewed as a continuum; the employment of a datum (the onset of Pleistocene) for *ab initio* considerations of landscape development is used here only to clarify this specific statement on the inherited palaeotopography.

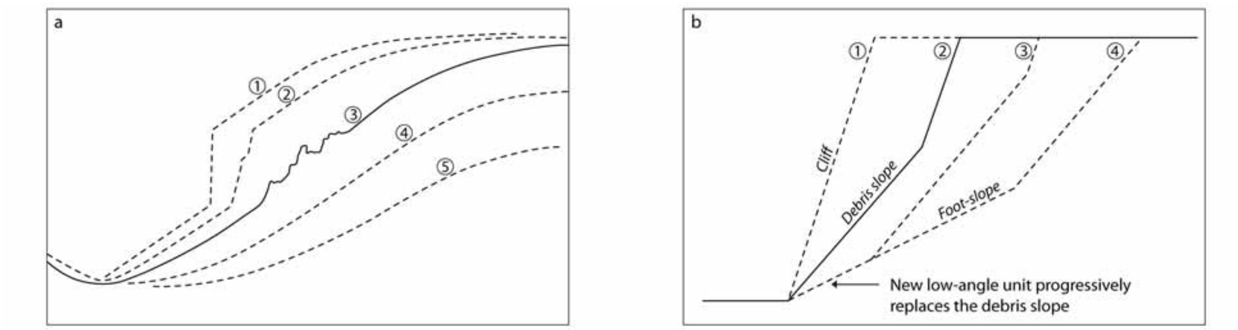

Fig. 6.26 Models of slope retreat. a) slope decline, b) slope replacement. After Clark and Small 1982: fig. 5.11, 5.12

growing up from below[63] (Fig. 6.26; Clark and Small 1982, 73).

Therefore, although we cannot evaluate in which stage of their evolution the slopes (*i.e.* basically slope classes 3 to 7) occur at present, we can still consider three main possibilities:

1. According to slope evolution models, whatever the gradient is now, the slopes were even steeper before (*i.e.* in previous stages of their evolution trajectories).
2. Some slopes may have been of lower gradient and at some point were steepened by an erosional episode (*e.g.* a landslide due to a heavy rainstorm, or a collapse due to an earthquake or fault activity); in this case, their denudational character remains, and it is reasonable to assume that archaeological material resting there would have been removed, either during the episode itself or afterwards, during the (erosional) adjustment of the slope towards a new graded profile (*cf.* Kirkby 1971; Montgomery 2001).
3. Some slopes may have retained their (high) gradient unchanged for a long period as:
   a. 'high cohesion slopes', *e.g.* with maximum slope angles within the range of 40º-50º or even higher; but these are typically free of debris, the latter collecting at the slope base as talus.
   b. 'repose slopes': these are also denudational forms, commonly appearing within the range of 30º -35º; they are strongly controlled by the angle of repose of the overlying particles that mantle the slope and are being slowly removed by gravity-controlled processes (Clark and Small 1982). Loose, surface material can indeed remain at the angle of repose for considerable time-spans, but it is unlikely to do so in the time-scale of a complete glacial-interglacial cycle (see argumentation below).

Numerous field data and experimental studies indicate that erosion intensity increases linearly with slope angle until a threshold value; when this value is exceeded the relationship inverses and erosion decreases instead (*e.g.* Liu *et al*. 2001). Because this critical slope gradient (the threshold value) depends on many interrelated factors (*e.g.* grain size, soil bulk density, surface roughness, runoff length, net rain excess, friction coefficient of soil), it differs in the various relationships, but it generally appears in the range of 41.5º-50º (ibid, and references therein). However, a decrease in the values of erosional parameters (*e.g.* runoff depth, overland flow velocity, scouring ability of overland flow) or total erosion, does not eliminate the disturbance effects of erosion: to put it simply, erosion rates may be decreased in a slope of >45 degrees as compared to a range of lower angles (*e.g.* 10º-45º), but erosion is definitely there,

63. A third model, that of *parallel slope retreat* "suggests that each of the upper slope units retreats by the same amount so that the whole profile (not just the individual units) retains its form but leaves an extending concave unit (a pediment) at its foot" (Clark and Small 1982, 74). This model, where slope development is essentially controlled by the rate of retreat of the free face, necessitates particularly resistant tocks and is best fitted in landscapes of arid and semi-arid climates. If it should be considered applicable to landforms of hard rocks in, for instance, southern Greece, it still does not affect the validity of the argument.

and capable of eventually removing detritus resting on the slope until a new threshold gradient is attained.

Equally numerous are the studies on clast entrainment and transport by slope and fluvial processes (*e.g*. Carrigy 1970; Church and Hassan 1992; Poesen *et al*. 1998). Geoarchaeologists and geomorphologists have used basic principles of gravity-induced transport and hydraulic mechanics to investigate the modes in which geomorphic processes affect archaeological material and site formation (Rick 1976; Schick 1987; Wainwright 1992; Fanning and Holdaway 2001; Morton 2004; Hosfield and Chambers 2004; Fanning *et al*. 2009). Overall, the general patterns observed by these two bodies of research seem to accord fairly well, which is expected if we consider artefacts as another type of clast, exhibiting similar mechanical behavior as natural angular material does. Thus, Rick (1976) found that lithic artefact density increases near the bottom of the studied site where slope angle is smaller, and average weight of lithic artefacts increases where slope decreases. The point to be stressed is that artefacts do move downslope under the influence of erosional processes, if not gravity alone; their transport distances and rates are largely controlled by slope gradient and may or may not vary according to size, shape and weight (Petraglia and Nash 1987; Morton 2004). Fanning and Holdaway cite the work of Poesen, which indicates that "the steeper the gradient, the longer the potential transport distance", and that "stones resting on very gently sloping surfaces are likely to move only short distances" (2001, 671). The latter scholars demonstrated that inter-rill entrainment does not affect artefacts larger than 2 cm (in maximum dimension) *on gradients less than 5º*, concluding that on such low-gradient surfaces, geomorphic processes like surface wash do not have significant effects on artefact distributions (Fanning and Holdaway 2001, 681). Another important result comes from field experiments and simulations of artefact movement, which show that most of the transport occurs in the period immediately following 'site abandonment' (Wainwright 1992).

The nine-unit land-surface erosion model and its geoarchaeological significance

One of the most widely used models of slope form and evolution is the 'nine-unit land-surface model' of Dalrymple and colleagues (1968). As French remarks (2003, 30), this model "is probably one of the best ways of envisaging erosion and landscape change...It forces one to visualize what is going on in each part of a landscape *at whatever scale of investigation is being used*" (emphasis added). Moreover, French suggests that the model "gives potential foreknowledge of where there may be good and poor archaeological and palaeoenvironmental preservation of sites and deposits" (ibid, 32). Furthermore, its significance to the exploration presented here lies also in its process-response defined units and the spatial correlation between soil processes, water movements and gravity (*cf*. Conacher and Dalrymple 1977)

Essentially, the model presents an idealized cross-section through one-half of a valley, from watershed (top) to river valley (bottom) (Fig. 6.27). The uppermost unit (1), the interfluve, is almost flat (0º to 1º) and is characterized by pedogenic processes (soil formation). Unit 2, the seepage slope, has modal slope angles between 2º and 4º and the dominant process is eluviation associated to lateral subsurface water movement. Unit 3 is a convex slope segment that can be regarded as the upper part of the fall face (French 2003, 31), with a gradient of 35º-45º and with creep being the dominant process. The next unit (4), the fall face, is the steepest part of the profile with gradients higher than 45º (and commonly >65º), "characterized by the exposure of the parent material and the general absence of soil and vegetation" (Dalrymple *et al*. 1968, 64); main processes here are rock falls and slides. Unit 5 is the transportational mid-slope segment with 26-35 degrees of slope; this may be the most actively eroding of all units, characterized by movement of material downslope by flow, slump, slide, creep and surface wash (ibid, 65). Unit 6 is essentially the locus of redeposition of colluvial material derived from higher up the slope profile by mass movement processes or by surface and subsurface water action. Unit 7 is the alluvial toeslope in the floodplain, it usually exhibits 0-4 degrees and is characterized by alluvial deposition; its main difference with Unit 6 is that material here

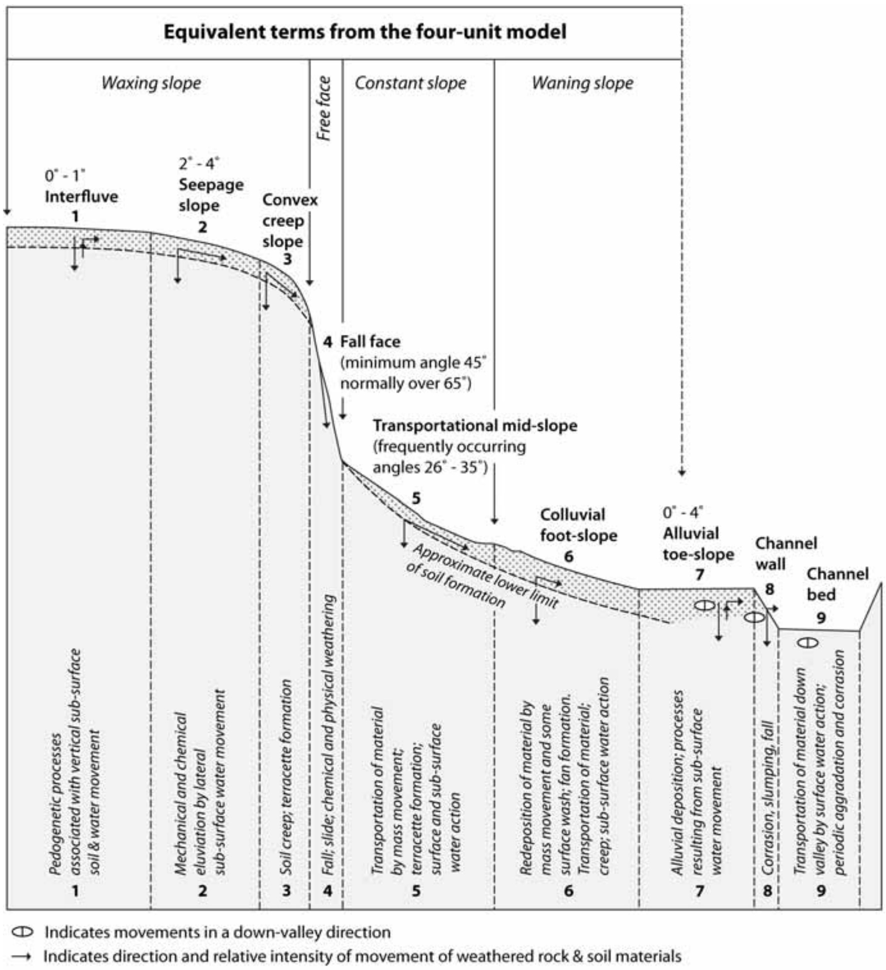

Fig. 6.27 The nine-unit land-surface model of Delrymple *et al*. (1968). Modified after Clark and Small 1982: fig. 5.10

is derived from upvalley rather than from upslope. Finally, Unit 8 is the bank of the river or stream, subjected to stream corrosion and slumping of the channel wall, whilst Unit 9 is the active channel, marked by bed transport down valley, aggradation and erosion. It is important to realize that "in any one land-surface, only unit 1 must occur, it must be first in the profile, and it must occur only once. The remaining eight units may or may not all be present; they may occur in any order; and they may occur more than once" (Dalrymple *et al*. 1968, 73).

What is made immediately clear from the model is that erosion and sediment transport (or, to use a more general term: instability) prevails in units 3, 4, 5 and -needless to say- units 8 and 9. Accumulation and re-deposition are the main geomorphic processes operating in units 6 and 7, whereas units 1 and 2 are characterized by neither deposition nor erosion, but rather, soil formation. Interpreting this in geoarchaeological terms (Table 6.5), one reasonably assumes that:

1. archaeological material is most likely to be found, and is most likely to be in primary context, only in units 1 and 2, namely at a gradient of 0° to 4°, where soil formation may engulf and protect sediments and artefacts. A primary context may be preserved also in unit 7 (the alluvial toeslope, with slope angles again between 0º and 4º) but, due to the dynamic nature of floodplains, a high degree of preservation is adequately advanced only under the combination of specific circumstances, for instance rapid burial associated with sedimentation of alluvial fines, followed by long-term stability and soil formation; if such conditions are not met, water movement and alluviation will most probably result in some sort of

Unit	Slope Angle (degrees)	Geomorphic Process	Potential for Preservation	Archaeological Context	Archaeological Visibility
1, 2	0 – 1 2 – 4	soil formation	high	primary	low, medium? high?
7	0 – 4	alluvial sedimentation	medium	?primary? secondary	low, medium
4	45 to >65	slope failure (landslides)	none	none	—
3, 5, 8, 9	26 – 45, possibly even higher	sheet erosion, soil creep, mass movements, channel corrosion, streambed transport	low	secondary, tertiary	high
6	modal angles	redeposition of colluvial material by mass movement, surface wash	medium	secondary	low, medium

Table 6.5 Summary of main geoarchaeological conclusions deduced from a hypothetical classification of the landscape according to the nine-unit land-surface model. Assessments on archaeological context and visibility are to some degree of speculative nature and are noted here only as 'most probable possibilities'. Artefact visibility, for example, can be high on lagged surfaces, low on depositional surfaces (because of burial) and somewhere in between (medium) on erosional surfaces. See text for discussion.

disturbance and hence the formation of a secondary context.

2. archaeological material is not to be found in unit 4, the 'fall face'.
3. archaeological material is least likely to be found in unit 3 (the 'convex creep slope' of 35º-45º), in unit 5 ('transportational mid-slope' of 26º-35º, the most actively eroding) and in units 8 and 9 (channel-wall and -bed, respectively). If there is any material preserved in those units, it will most certainly be from a derived context (secondary or tertiary).
4. archaeological material is likely to be found in unit 6 ('colluvial foot-slope'), but it will commonly be in a secondary context, as material coming from upslope. Of course, there is also the possibility that material was discarded on a colluvial foot-slope and was quickly buried by colluvial sediments; hence a primary context cannot be excluded, but it would be rather exceptional.

6.5.9 Discussion and conclusions

The most compelling conclusion of the examination unfolded in this chapter is that the controlling factors behind erosion in Greece, namely the nature and extent of vegetation cover, the lithology of surface materials, the characteristics of the soils, and the effects of a long history of land (mis)use, all are accentuated by the topographic setting and, essentially, the high relief of the Greek landscape.

Most -if not all- of the lithic scatters found today *on the surface* were probably once buried, unless they are found in very stable, low-relief landforms that were most of the time unaffected by stream and slope processes. Where it can be shown or safely assumed that they were indeed once buried, their exhumation inherently implies an episode of disturbance, the degree of which depends in turn on the processes involved, hence greatly on slope gradient. For instance, on steep slopes, it is difficult to envisage a low-energy winnowing of the finer sediment matrix leaving the artefacts behind as 'lag deposits'. If they were not buried, it is again only in the case of very specific erosion-protected landforms that artefacts can be considered to have escaped disturbance. Artefacts that were once buried on low-gradient surfaces that have experienced relative stability in the long-term, as in the case of some plateaus in karst areas, may have been deflated on those surfaces, although they otherwise remain more or less close to the original place of discard. In this instance, while the vertical integrity of the original artefact deposits may be lost, the lateral integrity remains relatively undisturbed (*cf.* Fanning et al. 2009, 127).

Similarly, all artefacts and sites that are now found *buried,* were once surface scatters. Because most disturbances occur in the period immediately following artefact discard (Wainwright 1992, 235), it is unlikely that artefacts would have escaped disturbance if they were discarded on land-surfaces of high gradient, for the same reasons that apply to the case of the non-buried-artefacts.

Therefore, it is only in *low-gradient* and/or *depositional* environments that we can assume:

1. little disturbance after discard and before burial
2. little disturbance after discard and upon burial (*e.g.* with sediment accumulation by low-energy depositional process as in a lacustrine context)
3. little disturbance after burial and upon exhumation (*e.g.* by deflation)
4. little disturbance after discard and with no or minimal burial (*e.g.* on karstic, flat-floored plateaus, where water-related processes are mostly endorheic)

Yet, both in a large-scale (landscape-scale) perspective and in the meso-scale of the hillslope, there are landforms and slope-classes that fall into an 'intermediate' qualitative category. This would include mainly slope gradients of 2-4 degrees or more, and landforms formed *e.g.* as hilly alluvial or fluvio-lacustrine units and alluvial/colluvial fans (although the latter may display even greater gradients). Existing badlands, exhibiting inclined to strongly inclined slope angles, can also be viewed as potentially belonging to these 'transitional' realms, with respect to their role in archaeological preservation and visibility. For instance, during the times when early hominins were moving around the ephemerally lacustrine environment of Kokkinopilos, the topography of this tectonic basin (polje) most probably ensued a depositional environment of generally low-gradients and low-energy geomorphic processes (hence, in accordance with the assessment about preservation- and context-wise advantages of basins). When the polje was subjected to the tectonic activity triggered by the adjacent fault, uplift caused stream incision and dissection of the basin-floor, gradually developing the characteristic badland landscape that we see today. The present-day micro-environment of Kokkinopilos is one of constant and rather vigorous erosion, which changes the local topography in time-scales of years or even months; and yet, there are considerable opportunities for recovering *geologically in situ* artefacts from non-reworked deposits. In short, landscapes and landforms with a biography such as that of Kokkinopilos, and perhaps also alluvial fans that are being -or, have recently been- dissected by incising streams, may overall serve as 'windows of opportunity' for recovering material from a geologically relatively undisturbed context. In that respect, they should not be underrated because of their steep gradients. Such 'windows' may potentially combine those two highly-valued and usually ill-assorted geoarchaeological conditions: a good degree of preservation and a high grade of visibility.

This is mainly why a geomorphological approach acquires a crucial role in the armory of archaeological methodology, for geomorphological indicators are capable of dictating both a qualitative and a quantitative allocation of efforts in archaeological research, by pinpointing:

1. areas that are promising for surface survey, *e.g.* extant stable surfaces of low-gradient, such as karstic plateaus
2. areas that may be appropriate for subsurface investigations, *e.g.* places where aggradation was predominant during and/or after the time-span of interest, such as a floodplain or a lacustrine setting, especially if recent (Holocene) erosion has exposed sections and palaeo-surfaces
3. areas that should be deemed unpromising for neither of the above, as places where neither stability nor aggradation, but, rather, erosion predominated during and/or after the time-span of interest.

A slope map is but one morphological measure that can be used to assess archaeological potential for preservation and recovery as a function of surface stability. At the regional level, stability classification is a powerful tool for both predicting and interpreting artefact and/or site distributions. To this end, the application of morphostratigraphic and/or allostratigraphic mapping assists in considering forms and processes in conjunction, thereby "classifying the landscape into surfaces whose form indicates a dominant depositional or erosional process for a discrete period of time" and "combining the surface morphology with the subsurface stratigraphy" (Wells 2001, 110-13). In this line, the slope map of Greece that I present here, as well as the discussion around it, can be regarded as an example of the methodological approach advanced in this study not only when assessing the current status of the Greek Lower Palaeolithic record, but also in addressing pathways for 'the way forward' in future investigations.

6.6 CONCLUSIONS

In accordance with the most up-to-date assessments of erosion in the Mediterranean in general and Greece in particular (*e.g*. Grove and Rackham 2001; Thornes 2009; Wainwright 2009), the examples and evidence outlined in this section altogether suggest that the terrestrial environment of Greece is largely prone to erosion, and it has most probably been even more so in the course of the Quaternary. The reasons behind such a high susceptibility to erosional processes are tightly interconnected by various generic dependencies and feedback mechanisms, and so the challenge here was to examine those complex parameters both in conjunction and as separately as possible.

Thus, section 6.2 highlighted seasonality, precipitation and the seasonality of precipitation as the major climatic attributes influencing slope and fluvial erosion. The present-day mode of seasonal river flow fluctuations and the seasonality of runoff would have been even more pronounced during the Pleistocene (*viz*. mostly during cold stages), accentuating the steepness of river hydrographs, *i.e*. increasing the ephemerality of flow regimes and decreasing the recurrence intervals of discharge peaks, thereby intensifying both frequency and magnitude of short-lived extreme events (*e.g*. floods). The geo-archaeological implications would be reflected in an enhanced coupling of the slope-channel systems, very rapid time-spans of incision/erosion and alluviation episodes, and hence reworking of artefact accumulations in alluvial/fluvial sediments every few thousand years. In the long-term, phases of landscape instability would mainly coincide to climatic transitions at both orbital (millennial) and sub-orbital (centennial/decadal) time-scales, and most notably during cold-to-warm transitions. In sum, although vegetation and rain-erosion maps defined also a spatial aspect, this chapter stressed primarily the *temporal dimension* of landscape (in)stability and erosion, namely the temporal windows of erosion/deposition, their possible durations, as well as the frequency of recurrence.

Section 6.3 explored tectonic controls putting the emphasis on the *spatial dimension*, as revealed in different scales of analysis: in a large-scale perspective, extensional tectonics and associated subsidence provided the necessary accommodation spaces for the bulk of the sedimentation occurring during the Quaternary, configuring the nature and extent of drainage basins. In the meso-scale, uplifting regions in compressional regimes (*e.g*. Western Greece) and areas that were parts of footwall-blocks in extensional regimes (*e.g*. Central Greece) have been mostly subjected to erosional processes, and basin inversion was stressed in this regard with particular reference to two case-studies (Gulf of Corinth and Megara Basin). Finally, the tectonic control on syn- and post-sedimentary processes was considered with regard to specific examples of drainage diversion/incision, as well as with reference to the effects of faults and earthquakes on slope development, landslide triggering and mass failures.

Whereas climate and tectonics constitute the main exogenetic and endogenetic driving factors in landscape evolution, their various interactions are manifested on the land-surface as slope processes. Main aspects of the latter were examined in section 6.5, where a slope map of Greece was presented as a means of visualizing and extracting general spatial trends of erosion, with the aim of exploring geomorphological and geoarchaeological patterns of preservation and visibility. Notwithstanding the coarse resolution and the speculative nature of this appraisal, it was argued that geomorphological opportunities for a relatively high degree of site preservation (also context-wise) is restricted to some one fourth (or, one third, at most) of the total area of the country, where low-gradient depositional settings occur. As it was shown, such an assessment finds strong support when considering the landscape under the perspective of the nine-unit land-surface model.

As a whole, these analyses largely explain why the Greek Lower Palaeolithic record is scanty, mainly surficial and rarely found in buried deposits: this picture is most probably a consequence of the prevalence of erosional rather than depositional landforms and land-surfaces. In addition, it can be argued that the fragmented nature of both the geological and the archaeological record before around MIS 6 is not a coincidence: it can be seen as reflecting an eco-geomorphic system dynamic and/or unstable enough to have considerably prevented adequate preservation of landforms and associated archaeological material *before the last interglacial-glacial cycle*. It is well-known that in north-western Europe, the severity of (some) glaciations erased the geological traces of earlier glacial and/or interglacial stages, biasing the geological archive in favor of the most recent stages. For Greece, it has already been shown that the geomorphological implications of the most severe glaciations would have been disproportional to the small size of the equivalent glaciers (*e.g.* Woodward *et al.* 2008). Even if we neglect the advance and retreat of glaciers as potential biasing agents as in the case of north-west Europe, the argumentation developed in this chapter (and most notably in sections 6.2, 6.3. and 6.5) indicates that there is still much evidence to suggest that short-lived but high-amplitude extreme erosional events, mainly accompanying climatic transitions, would have been capable of eradicating and/or *reworking* sedimentary units of previous glacial-interglacial cycles -to such an extent that the geomorphological imprint of warm and cold stages before the last interglacial are only discontinuously preserved today, mostly as erosional rather than depositional landforms.

Triggered either by climatic (*e.g.* deluges), tectonic (*e.g.* earthquakes) or, lately, anthropogenic agents (*e.g.* land use), it is indeed principally those episodic but catastrophic events that induce most of landscape instability, with landslides, slope failures and river incision being some of the main geomorphic responses. The validity of this assertion has acquired a wider acceptance for active tectonic settings with a susceptible terrain of high-relief and an eco-geomorphic system that responds rather fast to climatically generated disturbances, and it has been supported by studies of both past and present erosional phenomena in Greece and landscape development in the Mediterranean (*e.g.* Cendero and Dramis 1996; Martin and Church 1997, 278; Mulligan 1998; Kosmas 1999b, 19; Kirchner *et al*. 2001, 593; Grove and Rackham 2001, 247-252; Montgomery and Brandon 2002; Bintliff 2002; Macklin *et al.* 2002, 1638; Goldberg and Macphail 2006, 77; Koukis *et al*. 2009; Thornes 2009; Wainwright 2009, 190).

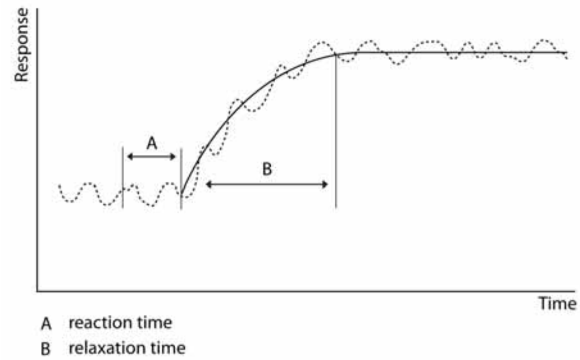

Fig. 6.28 Simplified graph showing the response of a geomorphic system subjected to disruption (after Imeson 1986: fig. 5). If a new disturbance occurs before the relaxation time is completed (*i.e.* within the time-span represented by 'B' in the graph), transient (unstable) forms will prevail.

Much as it is conceptualized in theoretical frameworks (*e.g.* Brunsden and Thornes 1979) Quaternary landscape evolution in Greece can be viewed "as a series of short adjustments between constant process-characteristic form states" (ibid, 481), where landscape change was mainly episodic and marked mostly by temporally discontinuous large-scale disruptions. Overall then, erosion may have not been prevailing for most of the time, but the combination of tectonism, increased seasonality, steep relief and erodible lithologies, altogether pre-conditioned 'windows of opportunity' for extreme erosional episodes to occur.

Referring to the impact of vegetation change on the Mediterranean fluvial systems, Macklin and colleagues (1995, 12) note that "a steady-state can neither be reached, nor maintained, if the relaxation time of the system is longer than the mean recurrence time of the disturbance to it" and that "an abrupt or step-functional change of this nature [*i.e.* millennial-centennial climatic fluctuations] would undoubtedly have exceeded both vegetation and soil system's capacity for adjustment, triggering a period of landscape instability with high erosion rates, valley floor aggradation and river metamorphosis". The researchers note also that although this scenario appears very likely, the low resolution of fluvial chronologies prevents a direct confirmation. Essentially, Macklin and

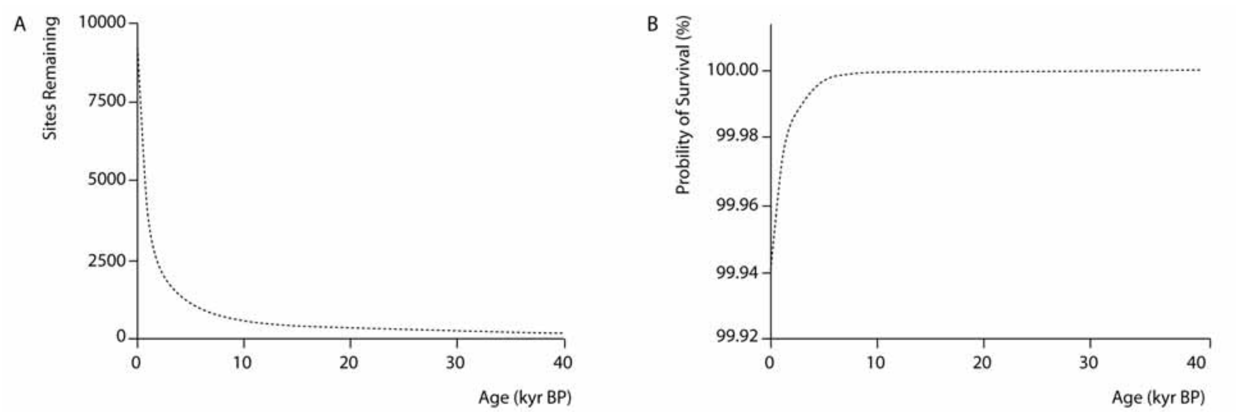

Fig. 6.29 Main results of the model of taphonomic bias, after Surovell *et al.* 2009: fig. 2. A: the predicted loss of archaeological sites over time assuming 10,000 sites at time zero; B: the annual likelihood of survival as a function of site age. Note that, in both (A) and (B), a marked change in the curve occurs at around the time-point of 10,000 years

colleagues' statement addresses the issue of landscape sensitivity (Fig. 6.28). The latter is concerned with whether transient or persistent (stable) landforms prevail under given circumstances and can be assessed by comparing landform or ecosystem relaxation time with the recurrence interval of vegetation or geomorphic disturbances (Brunsden and Thornes 1979, 480; Phillips 1995, 338-340). According to the transient form ratio (TF_r) introduced by Brunsden and Thornes (ibid), $TF_r = t_a / t_f$, where t_a and t_f are the mean relaxation and recurrence times, respectively. A ratio greater than unity indicates the prevalence of transient forms, whilst a ratio less than unity suggests that characteristic, stable forms will prevail. In other words, the system cannot attain stability (and/or steady-state) if a new disturbance occurs before the relaxation time has elapsed (the latter denoted as B in the scheme of Fig. 6.28). As it is explicitly stated by Macklin *et al*. (see above), this could have been the norm in the Mediterranean during the Quaternary due to the perturbations induced by climatic fluctuations at millennial, centennial, or even decadal scales. Also, tectonics would have contributed to such disturbances with a short recurrence time[64]. Although the task of mapping the sensitivity of landscape components in Greece during the Quaternary will remain unrealistic for the near future – hence hindering a confirmation of this assumption, it is tempting to assume a marked unsteadiness or transient behavior of the landscape in Greece over periods of 100 to 10,000 years during most of the Quaternary (*cf.* Brunsden 2001; Phillips 2005, 2006; Bishop 2007).

In a study of how taphonomic bias affects temporal frequency distributions of archaeological sites and dates, Surovell and Brantingham (2007) explored a basic geoarchaeological argument, namely that the probability that something will be removed from the geological/archaeological records is a function of time, which in turn causes an over-representation of recent events (or, sites) relative to older events. Building on the idea that taphonomic bias can be evaluated by examining the ratio of archaeological to geological contexts, Surovell and colleagues (2009) demonstrate that, in contrast to the results of their previous work, the rate of site loss does not remain constant through time, but instead declines with site age (Fig. 6.29).

The latter finding agrees well with the notion mentioned previously that a site is most at risk of being destroyed by surface processes during or directly after the time of abandonment, when material rests at or near the surface (Surovell *et al*. 2009, 1718). According to the model of Surovell *et al*., "if a site can survive its first 10,000 years of existence, its annual probability of destruction is reduced to approxi-

64. As an indication, let us recall here that the recurrence frequency of strong earthquakes (Magnitude >6 in the Richter scale) was estimated to be one every 1.7 years for the past 2,500 years (Caputo *et al*. 2006).

mately 0.01 %" (ibid). This 10^4-years sort of 'threshold' emerging from that model brings to mind the landscape-sensitivity argument discussed above: if the vast majority of landscape components in Greece acquired a transient (unsteady) behavior over periods of 100 to 10,000 years for the most part of the Quaternary, then archaeological sites would have fewer chances for surviving their first 10,000 years of existence -and hence improving thereafter the likelihood of survival up to the present.

In order to assess taphonomic bias on the spatial extent of the Lower Palaeolithic geoarchaeological record, we must consider the total area of Greece as the *potentially initial extent* of that record. To this, we would have to add the areas emerging during glacial sea-level lowstands (as well as during some interglacials *e.g.* that of MIS 11, and earlier ones; see Fig. 6.18 above); those areas have a total extent that approximates that of the current mainland (*ca.* 130,000 km^2). Since those latter areas are now submerged, we can consider that only about half of the initial record is theoretically today at the disposal of archaeologists for investigations. According to the arguments presented in section 6.5, potential areas for preserving the record are restricted to about 40% of half (50%) of the initial record, namely some twenty percent altogether. This twenty percent mainly represents the low-gradient, sedimentary basins, which have been hosting the bulk of sediment accumulation during the Quaternary. However, much of the latter value (the total area of the basins) corresponds to Holocene surficial sediments, since a large part of Pleistocene sediments lies now at great depths, as it is the case with the stacked alluvial sequence of Thessaly, where exposed Early and Middle Pleistocene sediments occupy only 0.8% of the total extent of the basin. Moreover, for another fraction of the basins it is pre-Pleistocene deposits that are exposed on the surface, as in the case of many depressions in northern Greece. According to Koukis and Ziourkas (1991), Quaternary formations occupy 15.8% of the total area of Greece. In the most optimistic assessment, Pleistocene sediments would thus cover about 10% of Greece, the rest 5.8% representing Holocene deposits. In effect, Pleistocene sediments cover today a spatial extent which corresponds to 10% of the 50% of the potentially initial extent of the record, namely some 5% of the latter; note that this assessment is independent of the geoarchaeological / geomorphological appraisal on the spatial coverage of potentially promising areas. If we do consider the geomorphological argument, then areas promising for preserving the record would amount to:

> 10% (= Pleistocene deposits) of the
> 40% (= low-gradient depressions including pre-Pleistocene and Holocene sediments) of
> 50% (= current continental extent of Greece)
> = 2% of the potentially initial extent of the Lower Palaeolithic geoarchaeological record.

The premise of an 'initial extent' is not entirely arbitrary, but it *is* dependent on time: the initial extent of double the size of Greece is best applied to time-periods before *ca*. 400-500 ka, when emerged areas in the Aegean attained their maximum coverage during sea-level lowstands. After MIS 10-12, this 'initial' value would be smaller (*viz*. the emerged areas were more restricted) but the final outcome of the estimate would not be much bigger (*i.e*. a larger percentage left (as promising) for investigations). When considering the geomorphological argument based on relief, the concluding amount is still small, even if we regard the 'initial extent' as equaling that of Greece's current area:

> 10% (Pleistocene sediments) of the
> 40% (low-gradient, same as above) of the area of Greece
> = 4% of the 'initial record' (and of the area of Greece in this scenario).

I believe that possible miscalculations due to the 'roughness' of the assessment would be averaged and cancelled out, and anticipate that more accurate calculations based on higher-resolution data would probably yield even smaller percentage values. In fact, it can be argued that for both of the above scenarios (pre- and post-500 ka assessments) the above-estimated final amounts could be viewed as maximum evaluations. For example, Early and Middle Pleistocene sediments, which are now either exposed on the surface or buried by relatively thin overlying sequences, would cover some 5%-10% of the total area of the basins; put differently, most of the spatial extent of exposed Pleistocene sediments in Greece refers to Late Pleistocene deposits.

In sum, from a hypothetical initial extent of the Lower Palaeolithic geoarchaeological archive in mainland Greece and the Aegean at almost any given datum in the Early and Middle Pleistocene, what has been (potentially) preserved until the present covers no more than five percent of the Greek mainland. This is admittedly a coarse estimate, but it provides a good indication of 'how much has been lost'; in that sense, it allows for a more objective explanation of why the Lower Palaeolithic record of Greece is so scarce. Furthermore, it is, in the main, a quantitative assessment with a semi-qualitative additional character: it quantifies in spatial terms the potential for artefacts to have been preserved in a primary and/or secondary context. However, this evaluation does not necessarily include all areas that combine *both* high preservation potential *and* good archaeological visibility (exposure). Examples of the latter situation were given with regard to inverted basinal settings: the Corinth Gulf and the Megara basin were discussed as case-studies of the meso-scale, whilst Kokkinopilos exemplifies this sort of 'window of opportunity' in a smaller-scale. For all of those cases, it was stressed that the critical factor lies in the mode, onset (timing) and duration/intensity of erosion since the initiation of uplift and the resultant basin inversion/drainage diversion, incision and exposure of low-gradient depositional settings that -until then- remained buried and protected. Differences in this critical factor may be able to explain the paucity of the Greek Lower Palaeolithic record as opposed to that of, for instance, Italy or Spain –a hypothesis that is discussed below, in the final chapter of this thesis.