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Constraints on the timing and regional conditions at the start of the present phase of crustal extension in western Turkey, from observations in and around the Denizli region

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Abstract

The chronology of extension of the continental crust in western Turkey has been the subject of major controversies. We suggest that these difficulties have arisen in part because of past misuse of dating evidence; and in part because the assumption often made, that deposition of major terrestrial sedimentary sequences implies crustal extension to create the necessary accommodation space, is incorrect. We report evidence that the present phase of extension began in the Denizli region at ~ 7 Ma, around the start of the Messinian stage of the Late Miocene. This timing matches the estimated start of right-lateral slip on the North Anatolian Fault Zone, and corresponds to a substantial increase in the dimensions of the Aegean extensional province to roughly its present size: beforehand, between ~ 12 Ma and ~ 7 Ma, extension seems to have only occurred in the central part of this modern province. In some localities, terrestrial sedimentation that began before this start of extension continued into this extensional phase, both within and outside normal fault zones. However, in other localities within the hanging-walls of normal faults, the start of extension marked the end of sedimentation. Relationships between sedimentation and crustal extension in this region are thus not straightforward, and a simple correlation should therefore not be assumed in structural interpretations. During the time-scale of this phase of extension, the Denizli region has also experienced major vertical crustal motions that are unrelated to this extension. The northern part of this region, in the relatively arid interior of western Turkey, has uplifted by ~ 400 m since the Middle Pliocene, whereas its southern part, closer to the Mediterranean Sea and with a much wetter climate, has uplifted by ~ 1,200 m since the Early Miocene, by up to ~ 900 m since the Middle Pliocene, and by an estimated ~ 300 m since the Early Pleistocene. This regional uplift, superimposed on the local effects of active normal faulting, is interpreted as a consequence of lateral variations in rates of erosion. A reliable chronology for this phase of extension in western Turkey, in relation to changes in the geometry of motions of adjoining plates and Late Cenozoic environmental change, is now in place.
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Keywords: Turkey; Miocene; Pliocene; Quaternary; Extension; Volcanism; Uplift; K-Ar dating

1. Introduction

The aim of this study is to investigate the timing of and conditions at the start of the present phase of extension of the continental crust in western Turkey. Preparation of this synthesis has been motivated by the realisation that much of the recent literature on this topic has reached incorrect conclusions due to misuse of dating evidence (see below), and thus gives a false impression about the relative timing of events. The availability of a reliable chronology will facilitate future

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studies along the lines of recent work testing regional kinematic models (e.g., [1, 2]), and undertaking rheological modelling of crustal deformation in this region in response to flow in its highly mobile lower-crustal layer (e.g., [3, 4]), for which reliable age estimates are necessary to test computer predictions of deformation rates.

Western Turkey forms the eastern part of the Aegean extensional province (Fig. 1). It is now generally accepted that the continental crust in this region is extending in response to forces exerted on it by subduction of the African plate beneath its southern margin (e.g., [5]). Forces related to this subduction also appear to be responsible for pulling southwestward the block of continental crust forming the small Turkish plate (e.g., [5]), whose motion relative to the Eurasian plate to the north requires right-lateral slip on the North Anatolian Fault Zone (NAFZ) (Fig. 1). Until recently, it was thought that the NAFZ became active at ~ 5 Ma (e.g., [6, 7, 8]). More recent analysis (e.g., [1, 2]) suggests instead that it initiated at ~ 7 Ma, and can be explained as a result of the change in the regional state of stress that accompanied the dramatic fall in water level in the Mediterranean basin in the Messinian stage of the Late Miocene (e.g., [9]).

The timing of the start of extension in western Turkey has been controversial. In the 1980s it was accepted that this extension began in the late Middle Miocene or early Late Miocene (~ 12 Ma), at the same time as slip on the NAFZ was thought to have begun (e.g., [10]). Subsequently, as better evidence emerged, the initiation of the NAFZ was placed *later*, at ~ 5 Ma (e.g., [6, 7, 8]). However, around the same time the start of extension in western Turkey was adjusted *earlier*, to somewhere in the range ~ 18 Ma (early Middle Miocene) to ~ 23 Ma (early Early Miocene), following

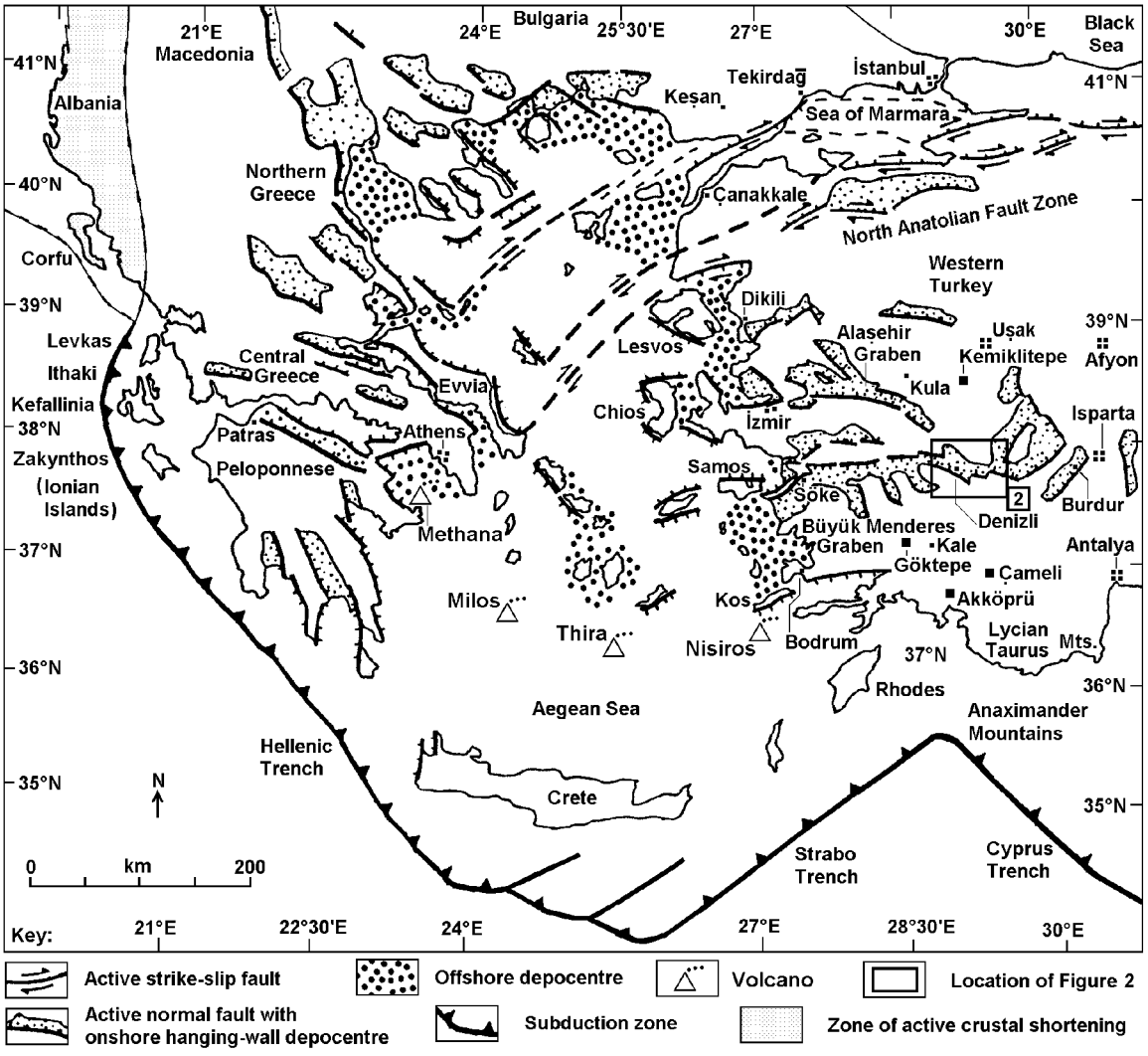


Fig. 1 Map of the Aegean region, showing active faulting and related sedimentation, adapted from Fig. 1 of [25] and Fig. 1 of [29], based on results of earlier studies (e.g., [48, 54]). The right-lateral North Anatolian fault zone enters this region from the northeast, its strands terminating against northeast-dipping normal fault zones which bound the northeast coasts of Evvia and adjacent islands. Box indicates the location of Fig. 2. Toygar is located SW of Kula adjacent to the normal fault at the northern margin of the Alaşehir Graben. The volcano symbol only indicates Quaternary subduction-related volcanism.

reports of apparently extension-related sedimentation and volcanism with biostratigraphic and isotopic dates of this age (e.g., [11, 12, 13]). It has subsequently been realised (e.g., [14, 15, 16, 17, 18, 19, 20, 21, 22, 23, 24]) that the presence of Early-Middle Miocene sediments beneath the younger fill at some localities within actively-extending grabens is fortuitous; it simply indicates that when the present phase of extension began, some normal faults cut through pre-existing depocentres. Many previous studies have also assumed that sedimentation in western Turkey is an indicator of crustal extension; that significant bodies of sediment can only accumulate in the presence of active normal faulting to create the necessary accommodation space, this faulting relating either to the present phase of extension or to some earlier phase. However, it is now clear that, with the lower crust highly mobile, sediment loading can dynamically create accommodation space for more sediment (e.g., [25, 26, 27, 28, 29]). As a result, there need be no requirement for sedimentation and extension to accompany each other in this region. The first issue addressed in this paper thus concerns how to distinguish sediment related to this present phase of extension from older sediment deposited under different conditions. A related controversy, which is also addressed, concerns whether the vertical crustal motions revealed in western Turkey by the present-day altitudes of sedimentary units relative to sea-level are (e.g., [30, 31, 32]) or are not (e.g., [3, 4, 25, 33, 34, 35]) simply predictable as the isostatic consequences of active normal faulting.

The second major issue concerns dating techniques. Pollen biostratigraphy has been extensively used to date the Neogene terrestrial sequences of western Turkey (e.g., [11, 13, 36, 37]). A series of pollen biozones was defined many years ago (e.g., [38, 39]), and their ages have since been calibrated using isotopic dating (e.g., [13]). However, this work has not considered whether the resulting numerical age spans are each consistent with the stratotype originally used to define each biozone, raising the possibility of systematic error. A second, more serious, problem with this technique has arisen, because studies (e.g., [20, 40]) have noted instances where pollen used to date sedimentary units is not from *in situ* lignite but instead from clasts of lignite reworked into younger deposits, and thus gives an apparent age greater than the true age of deposition of this sediment. Other techniques that can instead be applied to date this region's sediments include magnetostratigraphy, mammalian, molluscan, and ostracod biostratigraphy, and isotopic dating of associated volcanics. Only limited magnetostratigraphic work and isotopic dating have been undertaken on the Late Miocene and younger sequences in this region, and there is thus plenty of scope for refining its chronology by additional studies in future. The available evidence will be summarised, and supplemented with some new K-Ar dating of our own. Previous work (e.g., [37]) has dismissed the quite extensive mammalian biostratigraphic evidence from this region as unreliable. On the contrary, it now seems possible to make quite specific correlations between existing data-sets and established Neogene mammal (MN) biozones (see below), whose age spans have been calibrated in absolute terms (e.g., [41]).

Some recent studies have suggested that the present phase of extension in western Turkey was preceded by an earlier phase in the Early-Middle Miocene and/or the Late Oligocene (e.g., [14, 16, 17, 18, 19, 20, 24, 42]), these phases possibly being separated by an interval of crustal shortening or an interval with no significant crustal deformation. However, there is no consensus about the timing or geometry of any such earlier phase of extension, and other recent studies (e.g., [43]) have argued that no extension was occurring in the Late Oligocene to Early Miocene, at least in the eastern part of the modern extensional province where our study is concentrated. Resolution of what was occurring at this earlier stage is a major task, beyond the scope of this study, and readers are therefore referred to the recent literature (e.g., [44]) for details. We will instead concentrate on the timing of the start of the present phase of extension, and what was happening immediately beforehand.

The reason for concentrating on the eastern part of this actively extending region (Figs. 1. 2) is that where not much extension has occurred during the past few million years, evidence of what was occurring beforehand is expected to be better preserved. In addition, where only limited vertical crustal motions have occurred due to normal faulting, it is expected to be easier to demonstrate the existence (or not) of any regional component of surface uplift, as was first tentatively identified more than a decade ago [33]. Many studies have indeed investigated the structure of, and sedimentary record in, in the major normal fault zones farther west in western Turkey, where more extension has occurred, notably the Alaşehir and Büyük Menderes Grabens in Fig. 1 (e.g., [14, 16, 17, 20, 21, 22, 25, 34, 45, 46, 47]). However, no precise conclusion has been reached in these studies as to the timing of the start of their present phase of extension. The particular localities covered in this study will include the Denizli Basin, the Kale-Tavas Basin SW of Denizli, the Karaçay Valley SE of Denizli, and the Baklan Graben NE of Denizli (Fig. 2), the latter structure being also sometimes known as the Çivril Graben (e.g., [48]). We will also briefly touch upon the Neogene sequences in the Uşak-Güre and Selendi Basins to the north and NW of Denizli, notably around Kemiklitepe and Kula (Fig. 1), drawing upon the conclusions of recently published work (e.g., [4]).

2. The Karaçay Valley

The Karaçay valley, ~ 30-50 km ESE of Denizli city (Figs. 1, 2, 3), and now highly dissected by erosion (Fig. 4a) was formerly the site of extensive lacustrine sedimentation, many hundreds of metres thick. These sediments, dominated by limestone and marl in the upper part of the section, are capped by thin flows of basalt (Fig. 4a). Although exposure is limited, yellow-brown sandstone and conglomerate

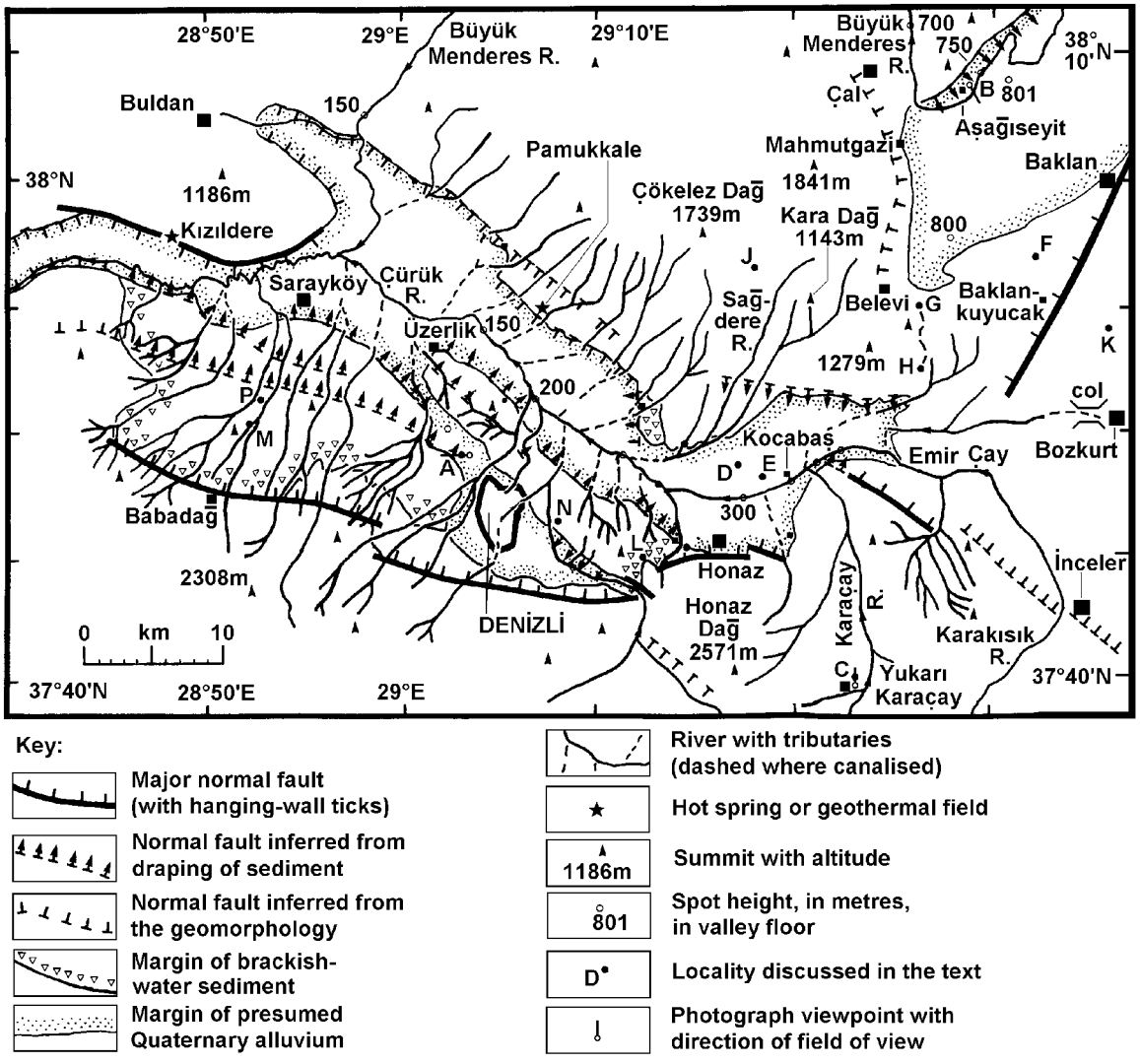


Fig. 2 Map of the Denizli region, adapted from Fig. 2 of [33]. See Fig. 1 for location. A, B, and C mark the viewpoints of Figs. 4a-c. The col at the eastern end of the Denizli Basin near Bozkurt marks the palaeo-outlet from the Acıgöl lake basin farther east, at times of wetter climate when the lake level is higher than at present (after Erol [55]). See text for discussion of other localities.

appear to predominate in the lower part of the sequence, exposed in river gorges south of Kocabaş (Fig. 2). Although other names also exist in the literature, we adopt the name Belevi Formation for this sequence (after [43]). At present, most of the basalt outcrops in this area form hilltops or ridge crests, indicating that where basalt is present it has protected the underlying lake sediment from erosion. However, the disposition of this basalt suggests that it erupted after lacustrine sedimentation ceased and the landscape was already slightly eroded by river channels (see below). Some of the basalt flows indeed seem from their geometry to have followed river channels for considerable distances, notably around the village of Yukarı Karaçay (Figs 5, 6a).

These Denizli or Karaçay volcanics were depicted on the regional geological map of this area [49]. They were first mapped in detail and subjected to petrological and geochemical analysis by Ercan *et al.* [50], who reported that they consist of shoshonitic basalt, latite, and trachyte, with phenocrysts of augite, biotite and apatite, and considered this volcanism to be extension-related. Subsequent analysis [51] of a typical sample confirmed this view: although not particularly mafic, with SiO_2 51.9 wt%, it is strongly potassic, with K_2O 5.5 wt%, and formally alkaline, with 10.5 wt% of normative olivine and 1.5 wt% of normative nepheline. Ercan *et al.* [50] noted that these volcanics cap the local terrestrial sedimentary sequence, which was then thought to be Pliocene, and thus tentatively assigned them a Late Pliocene age. These volcanics were later re-examined by Paton [52], who classified them as lamproites, although this study made no reference to the work by Ercan *et al.* [50]. It thus involved re-mapping the area, which led to the identification of some new basalt localities (Fig. 6), although some of the basalt outcrop previously mapped by Ercan *et al.* [50] (Fig. 5) was missed. Paton [52] also reported five Ar-Ar dates (Table 1), the only isotopic dating of this volcanism prior to the present study. The resulting ages are concentrated around 6-5 Ma

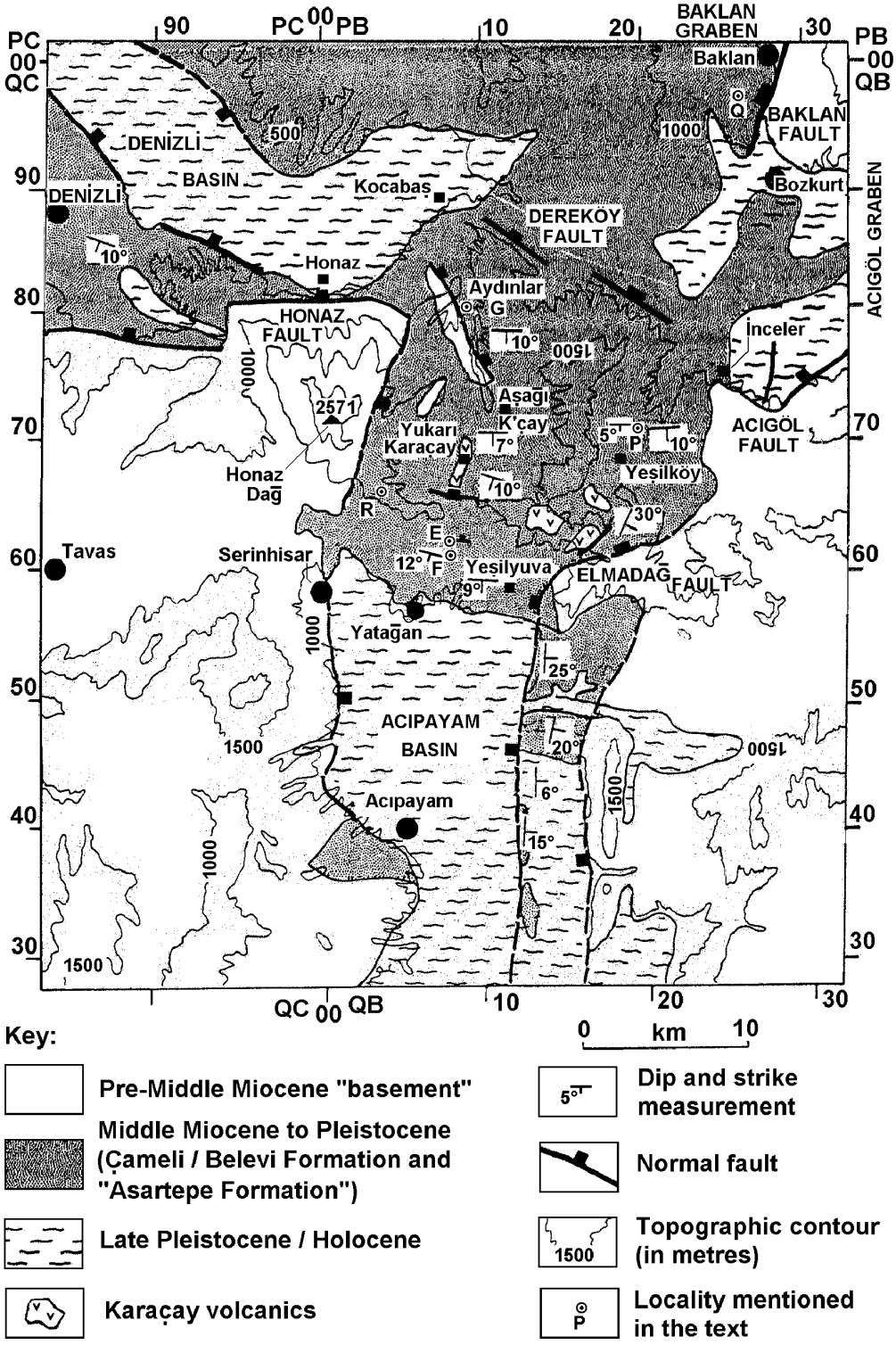
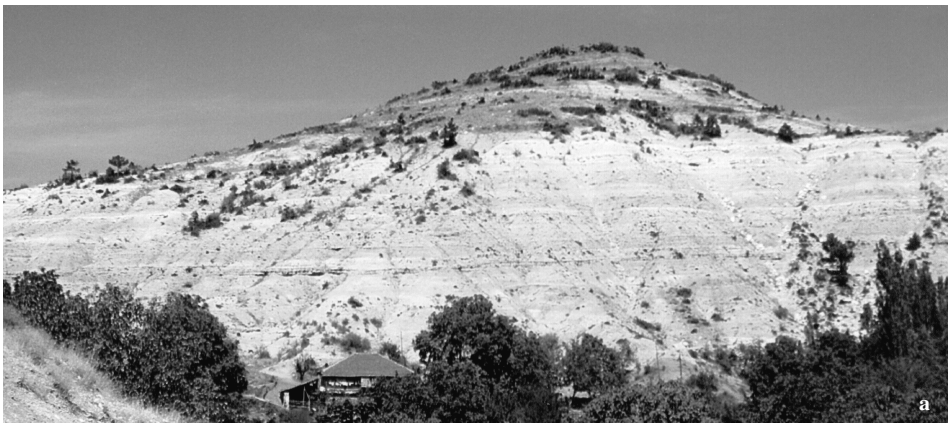


Fig. 3 More detailed map of the eastern part of the Denizli region, showing the Karaçay Valley and the location of its Messinian age volcanism, adapted from Fig. 5.2 of Paton [52], showing information from [49], with other detail added from [33] and from recent original fieldwork. Faults interpreted as of lesser importance are labelled using thinner symbols. This map and Fig. 6 are included here as a courtesy to the original author, Stuart Paton, who—to the best of our knowledge—did not publish this material after completing his Ph.D. [52] more than a decade ago. As Table 3 indicates, much of this material has nonetheless subsequently entered the literature through citation of it as "unpublished" or as "personal communications", without mentioning the original author. However, there are considerable problems with these maps, for instance this Figure (like maps in many other studies) has lumped together the pre-extensional sediments (which we correlate with the Muğla Group of the neighbouring Kale-Tavas Basin, of presumed Middle Miocene to Tortonian age) and the younger syn-extensional sediments in the various grabens (of (?) Messinian to Quaternary age) as a single sequence, which has all been presumed to be extension-related. The original map for Fig. 3 is also for some reason drawn to a distorted projection, with the vertical scale smaller than the labelled horizontal scale. See text for discussion.

Fig. 4 Field photographs from the vicinity of Denizli.



(a) View of Huykıran Tepe at the southern end of a N-S-trending ridge, in subhorizontally-bedded marl and lacustrine limestone, with some interbedded conglomerate near the top, capped by ~ 4 m of basalt, to the northeast of Yukarı Karaçay (also known as Kocapınar). The summit is at [QB 0954 7021], at ~ 1,350 m (A in Fig. 6a, where our dated basalt sample 01TR60 in Table 2 was collected); the viewpoint is at [QB 0932 6964] (B in Fig. 6a; C in Fig. 2), ~ 120 m lower. Note that the uppermost ~ 30 m of the lacustrine sequence, where the clastic input is greatest, has weathered to a slope away from its basalt cap, whereas the underlying almost pure carbonate sequence forms a near vertical cliff.



(b) The view SSE from the vicinity of Aşağıseyit near the northern margin of the Baklan Graben (B in Fig. 2), showing thin, subhorizontally-bedded sand, silt and marl (part of the Kızılören Member of the Çameli Formation) underlain by Menderes Schist basement, and incised to a depth of ~50 m by the Büyük Menderes river gorge. In the far distance, the Beşparmak Dağ mountain range, the footwall escarpment of the Baklan Fault (Fig. 2), is visible. Baklan town is near the left edge of the field of view, below the highest point on the skyline: the 1,446 m summit of Topocağı Tepe within this range. To the right is the extensive outcrop of the Çameli Formation around Baklankuyucak (F in Fig. 2; Q in Fig. 3), which has been uplifted to ~ 1,000 m above sea-level, ~ 200 m above the typical level of the Baklan Graben interior, possibly by a component of vertical slip on a normal fault that splays from the Baklan Fault near Baklan town.



(c) The exposure of calcareously-cemented marine beach rock between Denizli city and Sarayköy reported by Westaway [33] at A in Fig. 2 (at [PB 8065 8847], ~ 420 m altitude). This exposure consists of cemented sand and pebbles, containing abundant marine molluscs (predominantly of the genus *Cerastoderma*) and shell fragments. It dips at ~ 15° towards S10°W and is overlain by marl. It is also underlain by marl, as is exposed in the foreground.

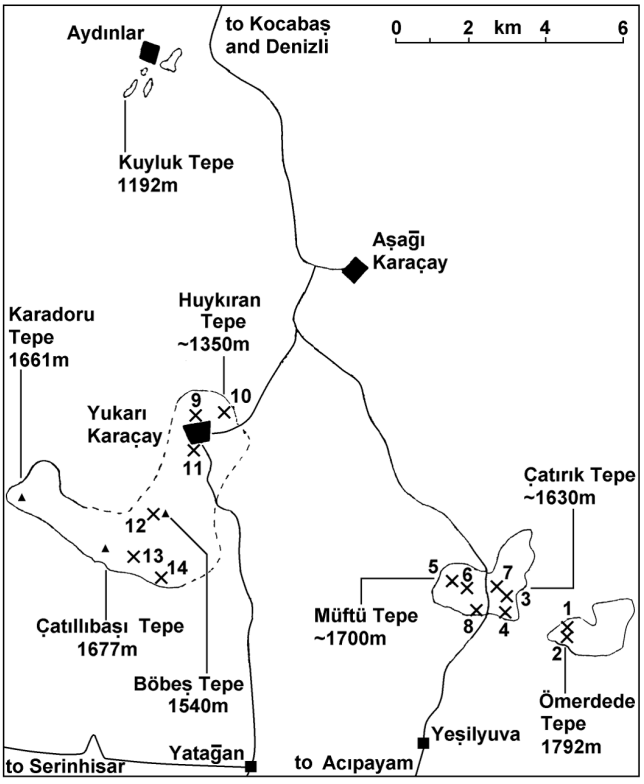


Fig. 5 Map of the Karaçay volcanic field, adapted from Fig. 1 of Ercan *et al.* [50], showing the three groups of basalt outcrops: north of Yeşilyuva, around Yukarı Karaçay, and around Aydınlar. Cross symbols with numbers indicate the sampling sites for the geochemical analysis by Ercan *et al.* [50]. This relatively inaccessible region can be reached using the road from Kocabaş to Serinhisar via Yukarı Karaçay and Yatağan, which is now surfaced throughout, or the road from Acıpayam to Yeşilyuva. At the time of our fieldwork in 2000 and 2001 the track shown linking Aşağı Karaçay and Yeşilyuva was not passable using normal motor vehicles.

(Table 1), in the latest Miocene. However, there is considerable scatter between samples (including pairs of samples from adjacent sites in the same flow units; Fig. 6b) and between individual splits of some samples, such that the overall range of dates spans from ~ 8 Ma to ~ 1 Ma, the weighted mean of all these dates being 5.96 ± 0.54 Ma (± 1σ), or 5.96 ± 1.08 Ma (± 2σ).

Paton [52] also Ar-Ar dated samples from the Kula volcanic field farther north (Fig. 1). However, most of these sample splits are now known to have yielded apparent ages that are much too old ([3, 4]), apparently as a result of dating amphibole phenocrysts that often contain inherited argon. As Table 1 indicates, the dating of the Karaçay volcanics instead utilised biotite phenocrysts and glass. These materials have less likelihood of retaining inherited argon, but glass may instead lose argon due to its amorphous structure, possibly yielding dates that underestimate the true age of any sample. Another difficulty affecting the potential reliability of any dating is that much of the Karaçay basalt is highly weathered and altered. However, in September 2001 we noted that the basalt capping Huykıran Tepe (Figs 4a, 5, 6a) northeast of Yukarı Karaçay had recently been excavated, apparently to construct foundations for a forest fire watchtower. An *in situ* sample from this atypically fresh exposure was thus collected, to verify the earlier dating results.

This sample was crushed to 1 mm size to allow magnetic separation of groundmass from the phenocrysts. Splits of this groundmass were then analysed for argon isotopes using the unspiked (or Cassignol) variant of the K-Ar technique. For technical reasons discussed in detail elsewhere [53], this method can achieve high precision when dating young rocks (Table 2). The resulting 6.71 ± 0.20 Ma (± 2σ) date for this sample is older than the weighted mean of Paton's [52] Ar-Ar dates. However, these two results are almost concordant at the ± 1σ level, and are readily concordant at the ± 2σ level. The main conclusion from Paton's [52] dating, that this instance of small-scale volcanism is latest Miocene, is thus confirmed.

Paton [52] also noted that, although somewhat variable (Figs. 3, 6a, b) the dips of the sediments in this former lake basin indicate overall tilting to the south by ~ 10°. This tilting has presumably resulted from slip on north-dipping normal faults farther north: the Honaz and Dereköy faults (Fig. 3; see also [33]). Near the SE margin of the depocentre, the lake beds tilt more steeply towards the SE, presumably indicating some normal component slip along the Elma Dağ escarpment that defines this basin margin (Figs 3, 6b). This escarpment is thus presumed to be another normal fault, for which the name Elmadağ fault is suggested (Fig. 3). Many minor normal faults are also evident within this sedimentary sequence.

In this summary of field evidence, 8-digit UTM co-ordinates are GPS fixes from our own fieldwork, 6-digit co-ordinates being from Paton [52]. Locality A [QB 0954 7021] (Fig. 6a) is the southern end of the N-S-trending ridgeline north of Yukarı Karaçay (Fig. 4a), rising to ~ 1,350 m at Huykıran Tepe (Fig. 5), where a ~ 4 m thickness of basalt caps the underlying lacustrine sequence of marl and limestone with occasional thin interbeds of well-cemented water-lain conglomerate with small (~ 5 mm to ~ 5 cm) matrix-supported clasts of basement lithologies (but not of the local basalt). B, [QB 0932 6964] (Fig. 6a), at ~ 1,230 m; is the viewpoint for this photograph, at the NE edge of the village. By locality C, at [QB 0968 7078] (Fig. 6a), the basalt has risen to a small summit, at ~ 1370 m. At several localities between A and C, steps in the basalt surface are interpreted as the edges of preserved remnants of individual basalt flows. The surfaces of these separate flows are observed to dip gently southward. No textures could be observed in the basalt that would indicate the direction of flow. However, it is presumed that this basalt erupted farther south (see below) and flowed northward. Locality D (Fig. 6a), ~ 300 m farther north at [QB 095 712], is the site of Paton's [52] sample 91-27 that yielded a 6.00 ± 1.54 Ma Ar-Ar date. This date (Table 1) is concordant with our new K-Ar date (Table 2) from the same flow unit, but with a much wider margin of uncertainty. Paton [52] designated this flow unit as the "west group" of basalt. We suggest the name "Yukarı Karaçay volcanics" instead.

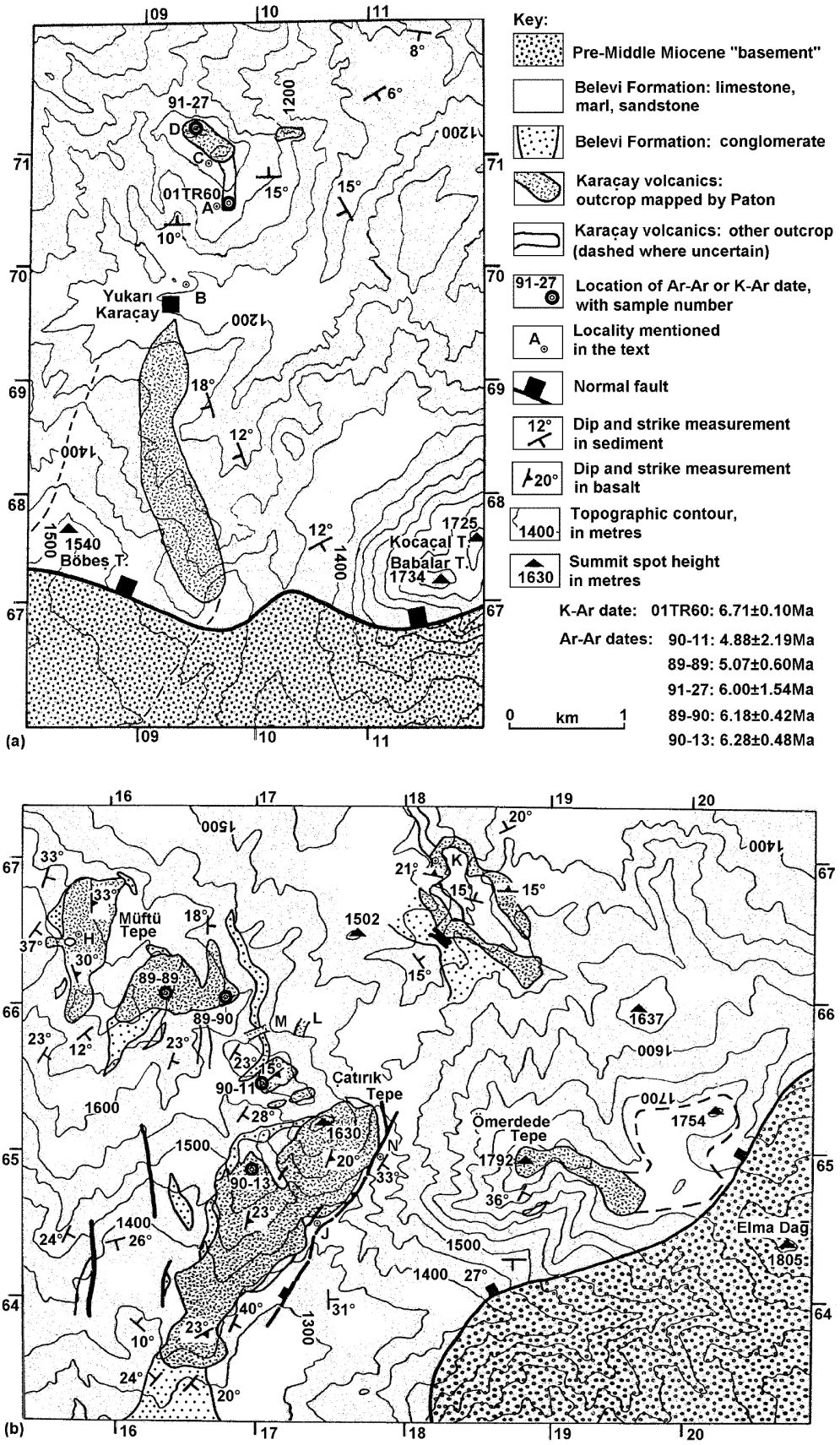


Fig. 6 Maps of the southern part of the Karaçay Valley, showing outcrops of basalt, terrestrial sediments that are assigned to the Belevi Formation [43] and interpreted here as of Middle-Late Miocene age, and normal faults. **(a)** Showing the Yukarı Karaçay volcanics. **(b)** Showing the Yeşilyuva volcanics farther east. Adapted from Figs 5.3 and 5.4 of Paton [52]. One kilometre squares of the UTM system (quadrangle QB) are used to define co-ordinates.

Table 1
Ar-Ar dating of the Karaçay volcanism.

Sample	Co-ordinates	Type	J	^{40}Ar (pl)	^{39}Ar (pl)	^{38}Ar (pl)	^{37}Ar (pl)	^{36}Ar (pl)	^{40}Ar*/^{39}Ar (± 1σ)	Age (Ma) (± 1σ)	Overall age (Ma) (± 1σ)
Main Group (Yeşilyuva volcanics)											
90-11	QB 170 655	biotite	$750 \pm 4 \times 10^{-7}$	19.79173	0.267552	0.049465	– 0.032	0.03425	36.15 ± 16.20	4.88 ± 2.19	4.88 ± 2.19
89-89	QB 164 661	biotite	$750 \pm 4 \times 10^{-7}$	219.7068	0.883002	0.213551	1.898887	0.641995	33.97 ± 4.21	4.59 ± 0.57	
		biotite	$750 \pm 4 \times 10^{-7}$	141.712	0.801111	0.11866	0.992643	0.366037	41.88 ± 4.64	5.66 ± 0.63	5.07 ± 0.60
89-90	QB 168 660	biotite	$670 \pm 4 \times 10^{-7}$	314.8247	0.351583	0.243238	0.200533	1.031799	28.24 ± 18.05	3.41 ± 2.18	
		biotite	$670 \pm 4 \times 10^{-7}$	353.3585	1.003935	0.253238	0.210937	1.051557	42.46 ± 9.40	5.12 ± 1.14	
		biotite	$670 \pm 4 \times 10^{-7}$	269.3875	1.615824	0.139238	0.456219	0.644786	48.80 ± 3.35	5.89 ± 0.41	
		biotite	$670 \pm 4 \times 10^{-7}$	471.0304	1.780452	0.301238	0.33042	1.268292	54.06 ± 2.62	6.52 ± 0.33	6.18 ± 0.42
90-13	QB 168 649	biotite	$750 \pm 4 \times 10^{-7}$	274.9273	1.184749	0.29431	2.0084	0.743876	46.52 ± 3.58	6.28 ± 0.48	6.28 ± 0.48
West Group (Yukarı Karaçay volcanics)											
91-27	QB 095 712	glass	$750 \pm 4 \times 10^{-7}$	20.86229	0.278426	0.062129	0.569384	0.028779	44.39 ± 11.43	6.00 ± 1.54	6.00 ± 1.54

Data are from tables III-1a and III-1b of Paton [52], except the sample co-ordinates that have been measured in this study from Figures I.12 and I.13 of this reference. Mineral separates were irradiated with neutron fluence J in the Imperial College reactor at Ascot, England, then ablated using a Nd/YAG laser at the Open University, Milton Keynes, England. For details of the technique see [127]. The argon released by laser ablation was analysed using a MAP 215-50 mass spectrometer, its counts being converted into equivalent volumes of gas measured at s.t.p.. ^{40}Ar*/^{39}Ar is the ratio of radiogenic ^{40}Ar to ^{39}Ar, estimated from the ^{40}Ar and ^{36}Ar concentrations in the sample, assuming a ^{40}Ar/^{36}Ar ratio for atmospheric argon of 295.5 (from [67]). The overall age for each sample has been determined in this study as the mean of the ages from individual sample splits, each inversely weighted by the square of its standard deviation.

Table 2
New unspiked K-Ar date for the Karaçay volcanism.

Sample	Site	K (wt%)	M (g)	P (%)	[^{40}Ar*] (pmol g^{-1})	Age (± 2σ) (Ma)	Overall age (± 2σ) (Ma)
West Group (Yukarı Karaçay volcanics)							
01TR60	QB 0954 7021	4.201 ± 0.040	0.95865	65.110	49.378	6.77 ± 0.14	6.71 ± 0.20
	Yukarı Karaçay		0.52225	70.458	48.542	6.65 ± 0.14	

Potassium content was measured by atomic absorption at the Centre de Recherches Petrographiques et Geochimiques, Nancy, France, on a split of groundmass separate. K is the resulting percentage by weight of potassium, expressed as potassium oxide. Argon isotopes were measured on other unspiked splits of the same groundmass separate at the Laboratoire des Sciences du Climat et de l'Environnement, Gif-sur-Yvette, France. M is the mass of each of these sample splits. P is the percentage of the ^{40}Ar present in each split that is estimated as radiogenic. [^{40}Ar*] is the amount of radiogenic ^{40}Ar estimated to be present. Calculations assume the decay and isotopic abundance constants from [67].

Farther south, at locality E (Fig. 3) [QB 1053 6528], an excavation for the new surfaced road from Yukarı Karaçay to Yatağan (built in 2001) has revealed a ~ 15 m high exposure of a north-dipping normal fault plane. At two points, striations in the recrystallised limestone of the footwall are visible. These indicate strike 275°, dip 50° and slip vector azimuth N12°E, and strike 256°, dip 40° and slip vector azimuth N6°E. This part of Turkey is known to accommodate biaxial extension, which is typically resolved in localities with more than one fault set as components of N-S and E-W extension (e.g., [25, 54]). This particular fault has evidently taken up N-S extension, like the Honaz and Dereköy faults farther north [33]. However, the sense of extension accommodated on the Elmadağ fault farther east remains unclear. Paton [52] suggested that it continues southward to form one strand of the N-S-trending normal fault zone forming the eastern margin of the Acıpayam Basin (Fig. 3). Figure 3 also indicates that its other end seems to link end-on into the Acıgöl Fault that appears to accommodate E-W extension [54]. Both these features thus suggest that the Elmadağ fault accommodates a component of E-W extension, although this cannot yet be confirmed by direct evidence. About 800 m farther south, another fresh road cut at locality F (Fig. 3) at [QB 1057 6447], exposes water-lain conglomerate interbedded with lacustrine marl, dipping at 12° towards S20°W. This footwall remnant of the Neogene succession of the Karaçay basin indicates that the Miocene depocentre was not bounded south of Yukarı Karaçay, as depicted in Fig. 6a; it was evidently continuous between the Karaçay and Acıpayam basins, as is indicated schematically in Fig. 3.

We are puzzled by Paton's [52] mapping of a north-dipping normal fault striking westward between Kocaçal Tepe and Babalar Tepe in the east and Böbeş Tepe in the west (Figs 3, 6a). We did not observe the southern margin of Neogene sediment depicted here, nor did we observe any evidence of a major normal fault, nor is any such fault depicted in the geological map of this area [49]. Furthermore, the mountain comprising Kocaçal Tepe and Babalar Tepe (Fig. 6a) is not part of the Neogene sedimentary sequence - it is instead made of Mesozoic limestone [49]. The Neogene sediment appears to be banked against it at up to ~ 1,500 m above sea-level, with no evidence of faulting, suggesting that it is a remnant of the pre-depositional landscape that was largely buried beneath this lacustrine sediment. The alternative mapping by Ercan *et al.* [50] (Fig. 5), showing basalt erupting from the vicinity of Karadoru Tepe/Çatıllıbaşı Tepe (R in Fig. 3), then flowing eastward to Böbeş Tepe and then northward to link up with the fragment preserved on Huykıran Tepe (Fig. 4a), seems more accurate. The eastward-flowing part of this outcrop caps lacustrine sediment at more than 1,600 m above sea-level, which forms the drainage divide between the Karaçay River (a tributary of the Büyük Menderes; Fig. 2) to the north and the drainage into the Acıpayam Basin (in the headwaters of the Dalaman River) to the south. The source of this basalt was evidently close to the east-dipping normal fault bounding the western margin of the Karaçay Valley, with Honaz Dağ in its footwall (Fig. 3), making the overall dimensions of this basalt outcrop ~ 6 km N-S and ~ 6 km E-W.

At Aydınlar, ~ 10 km north of Yukarı Karaçay on the west flank of the Karaçay gorge (Fig. 3), Ercan *et al.* [50] mapped several additional small basalt outcrops (Fig. 5), with dimensions of no more than a few hundred metres. The clearest of these forms the summit of the small hill Kuyluk Tepe, which rises ~ 10 m above the surrounding land surface (G in Fig. 3, at [QB 0752 7893]), ~ 700 m SSW of and ~ 100 m above the village that is ~ 1,100 m above sea-level, and which may mark the eruption neck. These outcrops, which are surrounded by many basalt boulders, may be the eroded remnants of a single small-volume flow; they adjoin another east-dipping normal fault (Fig. 3). Paton [52] seems to have missed this basalt locality, and so did not name it. We suggest the name "Aydınlar volcanics" for it.

The final area of basalt, which Paton [52] called the "main group", crops out over an area with dimensions ~ 3.5 km N-S and ~ 4.5 km E-W, centred ~ 4 km NE of Yeşilyuva and ~ 10 km SSE of Aşağı Karaçay (Figs 3, 5, 6b). As elsewhere, this basalt caps the underlying lacustrine sediment, creating uplands that form the present drainage divide between the Acıpayam Basin to the south, the Karaçay River to the NW, and the Karakısık River to the NE. The latter joins the Emir Çay near İnceler, which then flows NW along the hanging wall of the Dereköy Fault, before flowing into the Çürük River that forms the axial drainage of the Denizli Basin (Figs 2, 3). We suggest the name "Yeşilyuva volcanics" for the basalt in this locality, whose three principal outcrops cap Ömerdede Tepe to the east, Çatırık Tepe to the SW, and Müftü Tepe to the NW (Figs 5, 6b). Ercan *et al.* [50] reported the Ömerdede Tepe outcrop much larger (Fig. 5) than is depicted in Fig. 6b: it continues east for ~ 1 km and includes the summit area of the unnamed 1,754 m hill that is shown.

At locality H in Fig. 6b (at [QB 157 664]) Paton [52] reported an outcrop of Neogene sediment that is surrounded by lava. This was interpreted as a small palaeo-hill, existing at the time of eruption, which formed an "island" in the lava flow. At locality J (at [QB 174 645]), the basalt reaches the maximum observed thickness, of ~ 30 m, all in a single flow. This suggests ponding of the lava, again implying some palaeo-relief on the surface of the underlying sediment at the time of eruption. In contrast, at locality K (at [QB 182 670]) three superimposed flow units are evident, each ~ 5 m thick, with interbedded soil horizons. Elsewhere, the basalt is typically ~ 5 m thick. It is frequently observed to have a rubbly base, interpreted as a chilled margin, and to be directly underlain by a ~ 0.2 m thick red or ochre brown layer containing small igneous clasts, interpreted as a palagonitic tuff, which erupted subaerially at the start of the volcanism. In contrast, clasts of the basalt are not observed in the underlying sediment, suggesting that it post-dates all the lacustrine sedimentation in the region. At two localities, L and M (at [QB 172 659] and [QB 170 659]), Paton [52] reported dykes, which may have fed the flows in the vicinity. These were ~ 5-10 m thick and each could be traced laterally for ~ 100 m.

Paton [52] estimated the total volume of these Karaçay volcanics as ~ 0.014 km^3, consistent with covering an area of ~ 2.7 km^2 with an average thickness of ~ 5 m of lava. Although he clearly missed some outcrop (notably, at Aydınlar, east of Ömerdede Tepe, and SW of Yukarı Karaçay), these estimates seem reasonable and indicate the exceptionally small scale of this volcanism. Notwithstanding the scatter in numerical ages of individual sample splits in Table 1, the evidence suggests that a single, small-scale, short-lived, phase of volcanism occurred in this region.

Minor normal faults offsetting the basalt flows and underlying sediments are abundant. The most important of these, passing through locality N ([QB 178 650]) with a SW strike, appears to have downthrown the Çatırık Tepe flows in its hanging-wall relative to the Ömerdede flows in its footwall by several hundred metres (Fig. 6b). Given the characteristic southward tilt of the sedimentary sequence (Fig. 3) we attribute the overall southward increase in altitude of the basalt to components of northward downthrow on other minor normal faults (not shown). We presume that at the time of basalt eruption the upper surface of this lake sediment was subhorizontal, although the incipient incision into it by the Karaçay River had presumably already established a northward river gradient.

In contrast with the typical carbonate sedimentation in the Yukarı Karaçay area (Fig. 4a), there is a substantial proportion of conglomerate in the eastern part of the depocentre (Fig. 6b). A notable instance is at locality P (at [QB 199 706]; Fig. 3) where a ~ 100 m thickness of coarse conglomerate has been observed, apparently dipping at ~ 5° to the

south. Paton [52] considered this concentration of clastic material as evidence that the Elmadağ Fault was active during deposition of the Neogene sequence in the Karaçay basin, which thus was regarded as accumulating in its hanging-wall. However, it could simply reflect proximity to fluvial sediment sources entering the basin margin, in contrast with the more distal Yukarı Karaçay area. Paton [52] also reported evidence of soft-sediment deformation, which he took to indicate that seismicity, and thus, active normal faulting, was occurring during deposition. However, the only locality where he documented an instance of this was in a different part of western Turkey, not in the Karaçay Valley. Our own observations have revealed no such evidence: the carbonate sedimentation in the Yukarı Karaçay area and elsewhere has very regular bedding; the minor intra-formational unconformities observed elsewhere, typically bounding conglomerate interbeds, presumably reflect localised channel incision and aggradation, possibly at times of reduced lake level caused maybe by reduced rainfall. The minor normal faulting currently observed to offset and tilt the basin sediments is evidently post-depositional. We can thus see no basis for accepting that this Neogene lacustrine sequence was deposited during extension.

Numerous evidence indicators reveal that this volcanism occurred after the end of lacustrine sedimentation and after the start of the subsequent incision. These include the evidence that the geometries of the basalt flows responded to palaeo-relief on the already slightly eroded sediment surface, and the absence of basalt clasts in the underlying lake sediment. Evidence that the volcanism occurred after the start of extension consists of the presence of dykes, which indicate an extensional stress field, and the locations of the necks adjacent to normal faults, which can be presumed to have acted as magma conduits. It is indeed doubtful that without such easy paths, such small-scale volcanism could have reached the Earth's surface: if impeded, it would more likely have frozen at depth. The fact that in each locality the basalt is typically tilted by about the same angle as, or slightly less than, the underlying sediment (Fig. 6) suggests that the basalt eruption occurred only shortly after the extension began.

To summarise, this volcanism occurred abruptly around 6.7 ± 0.2 Ma. This was after lacustrine sedimentation ceased, and shortly after the start of extension and incision of this lake sediment. The end of deposition of this lake sediment, which we suggest coincided with the start of extension, is thus estimated to have occurred around ~ 7 Ma, at the start of the Messinian stage of the Late Miocene.

3. The Baklan Graben

Subparallel to the adjacent Burdur and Acıgöl grabens, the Baklan graben is the northernmost of three SW-NE-trending grabens near the eastern margin of the Aegean extensional province, each bounded to the SE by a major normal fault dipping NW (Fig. 1). As already noted, this structure is also sometimes known as the Çivril Graben (e.g., [48]). The sparse kinematic indicators documented in this region suggest that these faults accommodate a component of E-W extension, in contrast to the N-S extension observed on the major E-W-trending normal faults farther west, thus indicating that the extension of western Turkey is biaxial [25, 54]. The Burdur and Acıgöl graben interiors at present drain internally into lakes. Many studies (e.g., [55]) have noted that at stages of the Pleistocene with wetter climate these lakes have overflowed into adjacent river systems (cf. Fig. 2). However, being isolated from these river systems for most of the time, these localities provide no simple record of regional surface uplift. In contrast, the Baklan graben is drained by the upper reaches of the Büyük Menderes River (Fig. 2).

As previously noted ([33]; Fig. 4b), the Büyük Menderes is incising into the floor of the Baklan graben interior. In the vicinity of Aşağıseyit near where this river leaves this graben (Fig. 2), it has indeed incised by ~ 50 m, cutting through the base of the Neogene sedimentary sequence and into the underlying Menderes Schist basement (Fig. 4b). As has also been previously noted [33], this is not what is expected from conventional arguments relating to the isostatic response to active normal faulting, and provides a clear indication that local vertical motions in the vicinity of active normal faults are being superimposed onto regional uplift.

The badland landscape that has developed in the unlithified sediments flanking the Büyük Menderes gorge exposes part of the Kızılören Member of the Çameli Formation (e.g., [56]). This comprises mainly red and yellowish-brown fluvial sand and silt, with some interbedded gravel, sometimes passing laterally into lacustrine marl. Southwest of Baklan town, in the vicinity of Baklankuyucak (Fig. 2), is a separate dissected exposure of the same sediment, uplifted ~ 200 m (to ~ 1,000 m above sea-level, against ~ 800 m) above the typical level of the graben interior (F in Fig. 2; Q in Fig. 3). This may be the result of footwall uplift on a minor normal fault that splays WSW from the main Baklan Fault (Figs. 2, 3) near Baklan town. As Fig. 2 indicates, the Baklan (or Çivril) Graben is bounded on both margins by normal faults. The NW-dipping Baklan Fault bounds its SE margin, and the SE-dipping Çivril Fault bounds its NW margin.

This Kızılören Member of the Çameli Formation also crops out over much of the back-tilted block forming the footwall of the Baklan Fault, north of the Acıgöl lake basin; NE of Bozkurt and east of Baklan town (Fig. 2) (e.g., [56]). However, unlike in the hanging-wall of the Baklan Fault, in these footwall localities this member is overlain by two younger members—the so-called "mudstone-marl member" and "limestone member". The obvious interpretation is that a throughgoing fluvial system in this area (represented by the Kızılören member) was disrupted by the start of slip on the Baklan fault. The area NW of this fault continued to be drained by what is now the Büyük Menderes River, whereas the area SW of it developed into an isolated lake basin, where lacustrine marl, mud and later limestone accumulated. It follows that the end of deposition of the Kızılören Member

marks the start of extension in this region, thus potentially allowing this event to be dated.

Previous mapping [56] has indicated that (at Elmalıyurt near its Çameli stratotype; Fig. 1) the basal part of the Kızılören member has yielded remains of an undetermined species of *Hipparion*. *Hipparion*, the three-toed ancestral horse, first entered SE Europe and SW Asia in the early Late Miocene, in mammal biozone MN9, starting at ~ 11 Ma (e.g., [57]). The base of the Kızılören member can thus be no older than this. Also, at Mahmutgazi at the NW margin of the Baklan graben (Fig. 2), in the uppermost part of this member that is exposed in the badland landscape incised by minor left-bank tributaries of the Büyük Menderes, a diverse mammal fauna has been reported (e.g., [39]). This comprises both small and medium-sized *Hipparions*, and representatives of the genera *Paleoryx* and *Gazella* (antelopes), *Diceros* (i.e., *Dicerorhinus*) and *Chilotherium* (rhinoceroses), *Samotherium* (giraffe), *Oioceros* (bovid), *Microstonyx* (pig), and *Ictitherium* (hyaena). This fauna has previously [39] been thought to belong to the later part of the "Pannonian" (i.e., Meotian) stage of geological time. It represents the Turolian mammal stage and, in modern terminology, is most likely equivalent to mammal biozone MN11 or MN12. It is probably in the latter, for instance because (by analogy with other sites in the region; see, e.g., [58]) the reported small *Hipparion* may well be the species now known as *Cremohipparion matthewi* (see Fig. 8 caption), which is known only from MN12-13 (e.g., [57]). It follows that deposition of the Kızılören Member ended around the end of mammal biozone MN12, which was at 7.1 Ma according to the numerical time scale that we are using [41].

The reason for not reporting this fauna from Mahmutgazi in more detail here is that another site with a very similar fauna is now much better documented, at Kemiklitepe near the SW end of the NE-SW-trending Uşak-Güre Basin, ~ 50 km farther NNW (Fig. 1). As reported by Ercan *et al.* [59] and summarised elsewhere [3, 4], the stratigraphy of this basin consists mainly of the İnay Group. This is subdivided into the Ahmetler and Ulubey formations, the former being further subdivided into the Balçıklıdere and Gedikler members. The Balçıklıdere member is mainly fluvial sand; the Gedikler member is silt- and clay-dominated; and the Ulubey Formation is mainly lacustrine limestone. This sequence of units thus bears a strong resemblance to the three members of the Çameli formation, with the Balçıklıdere member corresponding to the Kızılören member. The main biostratigraphic horizon at Kemiklitepe, designated Kemiklitepe AB, near the top of the section (Fig. 7), is magnetostratigraphically dated to ~ 7.1 Ma (Fig. 7) and biostratigraphically dated to mammal biozone MN12 (Fig. 8). An older horizon, Kemiklitepe D, seems from the magnetostratigraphy to lie around ~ 8 Ma and may represent early MN12 or late MN11 (Figs. 7, 8) (see also, e.g., [60, 61]). The implication is that here, also, a stable fluvial depositional system that may have lasted for millions of years was disrupted around 7 Ma and replaced by a lake basin.

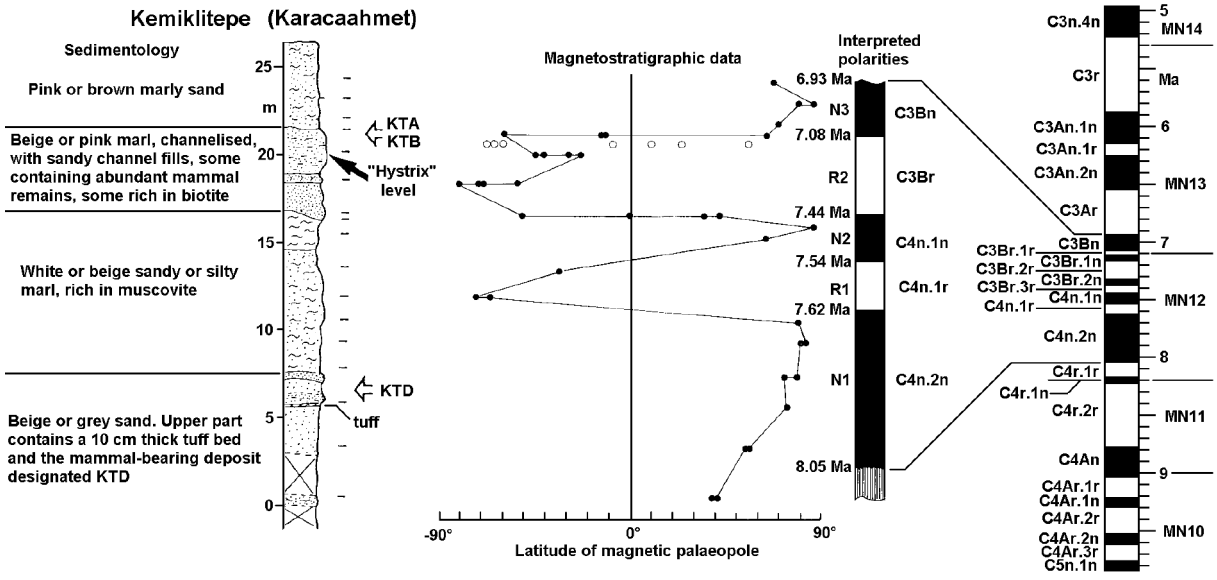


Fig. 7 Sedimentary and magnetostratigraphic records from the mammal site at Kemiklitepe [PC 893 540], located ~ 2 km S of Karacaahmet at the western margin of the southern part of the Uşak-Güre Basin. Adapted from Fig. 5 of [60], modified to incorporate the geomagnetic polarity time scale and mammal biozones from [41], the latter being calibrated largely using data from the Eastern Mediterranean region. Using data from the Iberian Peninsula, Agusti *et al.* [128] suggested modifications to the boundaries of MN zones: their MN10-MN11 boundary was slightly later, at the end of chron C4An (~ 8.7 Ma) rather than early in this chron (~ 9.0 Ma); their MN11-MN12 boundary was considered uncertain, but placed most likely within chron C4n.1n (~ 7.5 Ma) rather than chron C4r.1n (~ 8.2 Ma); their MN12-MN13 boundary was also slightly later, within chron C3Ar (~ 6.8 Ma) rather than chron C3br.1r (~ 7.1 Ma); and their MN13-MN14 boundary was also later, at the end of chron C3n.4n (~ 4.9 Ma) rather than near the end of the preceding chron C3r (~ 5.3 Ma). These modifications relate primarily to value judgements regarding faunal turnover in different mammalian datasets. We prefer to use the Steininger *et al.* [41] solution that was calibrated for the present study region, as faunal turnover in the Iberian Peninsula may have been diachronous. This decision makes no difference to the interpretation of the KTA and KTB sites, which fit MN12 in both schemes, but the older KTD site adjusts from late MN11 in the Steininger *et al.* [41] scheme to early MN12 in the Agusti *et al.* [128] scheme.

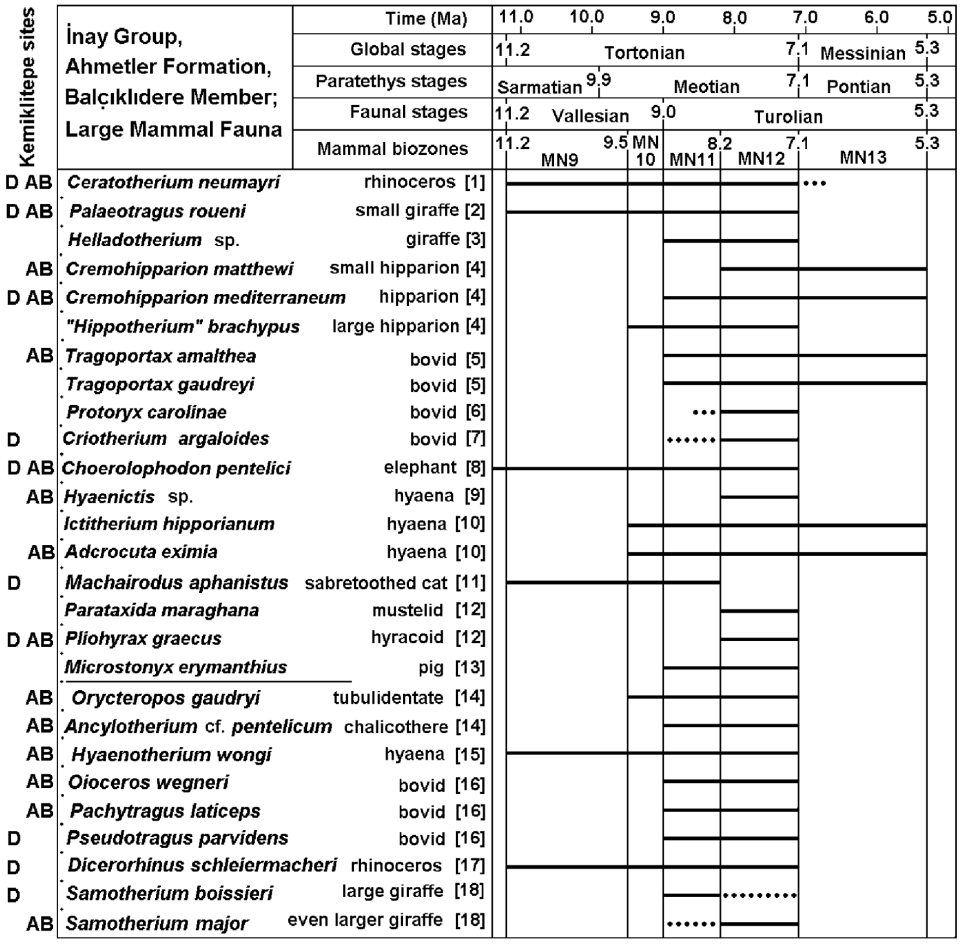


Fig. 8 Age constraints on mammal taxa from the Balçıklıdere Member of the Ahmetler Formation, the uppermost clastic part of the İnay Group in the SW Uşak-Güre Basin, again using the time scale from [41]. Fossil taxa, from [59], were listed (without noting precise provenances) at sites in the SW Uşak-Güre Basin: at Akçaköy [PC 840 700] and Fakılı [PC 790 750], and at another site called "Balçıklıdere". The Kemiklitepe site, located around [PC 893 536] in the uppermost Balçıklıdere Member on the eastern flank of the Balçıklıdere river gorge, ~ 2 km SSW of Karacaahmet village (Fig. 2), is thought to be this "Balçıklıdere" site (e.g., [60]). The horizontal dividing line separates species listed by Ercan *et al.* [59] from those additionally listed by Şen *et al.* [60]. The first column indicates whether each species is present in the Kemiklitepe D or AB levels. Notes are: **[1]** Listed as synonym *Diceros neumayri*; age range data are from [129]. Data for Kemiklitepe are from [60] and [130]. **[2]** Age range data from [71]. Data for Kemiklitepe are from [60] and [131]. **[3]** Likely species: *Helladotherium duvernoyi*; age range of this species is from [71]. **[4]** Age range data from [57]. As this reference discusses, the historical "species" *Hipparion gracile* may well be a synonym for "*Hippotherium*" *brachypus*, and possibly also for *Hippotherium primigenium*. The quoted age range is for *H. brachypus*. Data for Kemiklitepe are from [60] and [132]. **[5]** Listed as synonyms *Tragoceros amaltheus* and *Gazella gaudreyi*. Age range data are from [71]; data for Kemiklitepe are from [60] and [133]. **[6]** Restricted to MN12 according to [71], but appears late in MN11 according to [58]. This discrepancy relates to different correlations between the stratigraphy of Samos (where this species is found) and the age spans of the MN units. **[7]** Restricted to MN12 according to [58]. Present also in MN11 according to [71]. Identified as "cf." by Ercan *et al.* [59], but considered definitely present at Kemiklitepe D by Şen *et al.* [60] and Bouvrain [133], who noted that it is the only species known from its genus. **[8]** Age range data are from [58]. Data for Kemiklitepe are from Şen *et al.* [60] and Tassy [134], who reported that site D yielded a more primitive subspecies than AB. **[9]** Likely species is *Hyaenictis graeca*. Age range data for this species are from [135] and [136]. Data for Kemiklitepe are from [60] and [137], who reported *Lycyaena* sp.—a synonym for *Hyaenictis* sp. **[10]** Age range data are from [136]. *Adcrocuta eximia* is listed as synonym *Hyaena eximia*. Data for Kemiklitepe are from [60] and [137]. **[11]** Identified to genus level by Ercan *et al.* [59]. The species *Machairodus aphanistus* was identified at Kemiklitepe D by Şen *et al.* [60] and de Bonis [137]; age range data for it are from [135]. Other occurrences of this genus reported by Ercan *et al.* [59] may represent the younger species *M. giganteus*, which lived during MN 11-13 according to [135]. **[12]** Age range data are from [58]. Data for Kemiklitepe are from Şen *et al.* [60] and Baudry [138], who reported identification to species level at AB but only to genus level at D. **[13]** Age range data from [139]. **[14]** Reported at Kemiklitepe AB by Şen *et al.* [60] and Şen [140]. Age range is from [58], presence of each species in MN12 being based on their occurrence in the Main Bone Beds of Samos. **[15]** Reported at Kemiklitepe AB by Şen *et al.* [60] and de Bonis [137]. Age range is from [58], as for **[14]**. **[16]** Species identifications are from Şen *et al.* [60] and Bouvrain [133], who reported synonyms *Protoryx laticeps* and *P. parvidens*; age range is from [71]. **[17]** Reported as synonym *D. pikermiensis* by Şen *et al.* [60] and Geraads [130]; age range is from [58], as for **[14]**. **[18]** Şen *et al.* [60] and Geraads [131] reported *Samotherium major* from Kemiklitepe AB and a slightly different form, which they listed as *Samotherium* ? sp. but considered to be an intermediate form between *S. boissieri* and its descendent *S major*, from Kemiklitepe D. Gentry and Heizmann [71] considered that in combination these taxa spanned MN11-12, so this species (or subspecies) change probably occurred sometime around the MN11-12 boundary.

4. The Kale-Tavas Basin

The Kale-Tavas basin is another NE-SW-trending depocentre located to the south and SW of Denizli. It is ~ 60 km long, from near Tavas (Fig. 3) in the northeast to SW of Kale (Fig. 1) in the southwest. Many studies (e.g., [20, 43, 62, 63, 64]) have reported that local Oligocene to Early Miocene shallow marine sedimentation is superseded by a late Middle Miocene to Pliocene terrestrial sequence. Akgün and Sözbilir [43] classified the initial marine sequence, their Akçay Group, as "molasse", but believed the overlying terrestrial sequence (which they called the Muğla Group) to be extension-related (cf. [64]). However, no evidence of active normal faulting bounding this depocentre was presented; the deduction that this deposition was extensional was based on reasoning to the effect that it fell within the time-scale during which extension was thought by the authors to have taken place.

Deposits of the initial marine "molasse" sequence are now found at up to ~ 1,200 m above sea-level around both Tavas and Kale. As the global sea-level at that time was probably less than ~ 100 m higher then than at present, these altitudes provide rough indications of net amounts of subsequent surface uplift in these localities. The youngest deposits at the top of the Muğla Group (unit N3 B5 of [62, 63]; the Yarkındağ Formation of [20]; Fig. 9), crop out in the vicinity of Göktepe (Fig. 1) at up to ~ 900 m above sea-level.

In order to make rough estimates of uplift rates in this region, the chronology of this terrestrial deposition needs to be established. Decades ago, four sedimentary units—the Turgut, Sekköy, Yatağan, and Milet Formations were recognised (Fig. 9) within what is now known as the Muğla Group. These correspond, respectively, with Gökçen's [62, 63] units N3B1, N3B2, N3B3 and N3B4 together, and N3B5, as illustrated in Fig. 9. Gökçen [62, 63] subdivided this sequence into three ostracod biozones, as also illustrated in Fig. 9, thus permitting some measure of correlation with other terrestrial sequences in the Paratethys region. The mammal biostratigraphy identified at sites in these sedimentary units has been summarised by Sickenberg and Tobien [39]. The Turgut Formation has yielded characteristic Middle Miocene taxa, notably of the genera *Gompotherium* and *Anchitherium* (Fig. 9). Although these forms survived later, into the Early Vallesian mammal stage (biozone MN9; to ~ 9.5 Ma; [41]) (e.g., [65]), it thus seems reasonable to place this Formation in the preceding Astaracian mammal stage (biozones MN6 to 8; ~ 16 Ma to ~ 11.2 Ma); a conclusion also reached by others [43]. Atalay [66] also reported Astaracian faunal elements in the Sekköy Formation. Becker-Platen *et al.* (1977) obtained K-Ar dates for volcanics interbedded with it of 13.2 ± 0.35 Ma and 11.1 ± 0.2 Ma. These numerical ages should be increased by 2.5% to be consistent with the modern decay constants from Steiger and Jäger [67], making them 13.53 ± 0.36 Ma and 11.38 ± 0.21 Ma (cf. [68]). Sickenberg and Tobien [39] also reported instances of *Hipparion* remains and other Turolian faunal elements in localities in the Yatağan Formation in this region

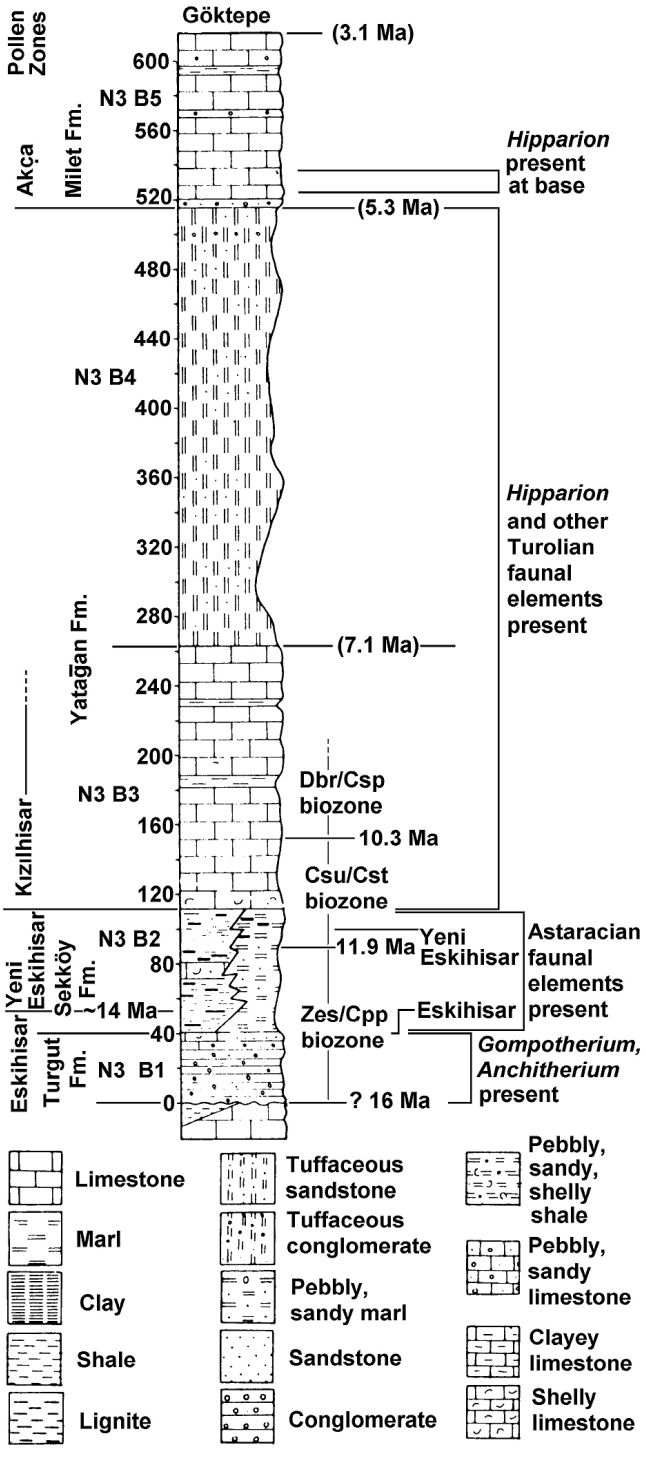


Fig. 9 Stratigraphic column for the upper terrestrial sequence in the Kale-Tavas basin (which we call the Muğla Group, after [43]) at Göktepe (Fig. 1), adapted from part of Fig. 2 of [63]. The numerical ages for Gökçen's [62, 63] ostracod biozones are derived in this study by applying modern numerical age constraints (from [41] to the correlations presented in her original papers. The biozones are identified by abbreviations of their index species: Zes/Cpp means *Zonocypris eskihisarensis*/*Candona parallela pannonica*; Csu/Cst means *Candona suevica*/*Candona steinheimensis*; Dbr/Csp means *Darwinula brevis*/*Cypria* sp.

and their presumed correlatives elsewhere in Turkey. Atalay [66] reported both Vallesian and Turolian faunal elements at different levels, and Becker-Platen *et al.* [69] reported K-Ar dates of 10.2 ± 0.15 Ma and 9.25 ± 0.20 Ma (which adjust to 10.46 ± 0.15 Ma and 9.48 ± 0.21 Ma) for volcanics interbedded within this unit. These four K-Ar dates are all for andesitic tuff [68]; they evidently just pre-date the start of alkali basaltic volcanism in this part of western Turkey (Table 3). The 11.38 ± 0.21 Ma date is for tuff directly below the Yeni Eskihisar 2 Astaracian mammal site (with *Aceratherium* aff. *tetradactylum* and other species [69]) near the top of the Sekköy Formation, whereas the 13.53 ± 0.36 Ma date is for tuff directly above the Yeni Eskihisar 1 mammal site, which is also the stratotype for the Yeni Eskihisar pollen biozone ([69]; see also below). Tuna [70] recently reported a new mammal site within a clayey unit of this Yatağan Formation at Yukarısazak near Kale (Fig. 1), which yielded four characteristically Turolian species: the giraffe *Helladotherium duvernoyi* and the bovids *Tragoportax amalthea*, *Protoryx carolinae*, and *Gazella deperdita*. According to Bernor *et al.* [57] all four of these species spanned both biozones MN11 and MN12, although Gentry and Heizmann [71] considered the last two restricted to MN12. Sickenberg and Tobien [39] also reported instances of *Hipparion* in the basal part of the overlying Milet Formation, which implies that the Miocene-Pliocene boundary occurs in this part of the sequence. We thus tentatively estimate that the limestone-dominated unit N3B5 (the

Table 3
Dated Late Miocene alkali basaltic volcanism in western Turkey bordering the Aegean Sea.

Name	Location	Source	Method	Age (± 1σ)
Karatepe, Çorlu	~ 25 km NE of Tekirdağ	[141]	K-Ar, on ???	8.95 ± 0.2 Ma
Karakaya	~ 10 km W of Tekirdağ	[141]	K-Ar on groundmass	8.92 ± 0.10 Ma
		[52]	Ar-Ar on glass	6.47 ± 0.43 Ma
Mahmutköy	~ 15 km S of Keşan	[142]	Magnetostratigraphy	6.74 ± 0.19 Ma
		[143]	K-Ar, on ???	6.7 ± 0.7 Ma
Tavşan Islands	~ 40 km SW of Çanakkale	[144]	K-Ar, on ???	10.1 ± 0.2 Ma
		[144]	K-Ar, on ???	9.5 ± 0.3 Ma
Taştepe	~ 20 km S of Çanakkale	[144]	K-Ar, on ???	11.0 ± 0.4 Ma
		[52]	Ar-Ar on glass	10.51 ± 0.98 Ma
Ezine	~ 40 km S of Çanakkale	[144]	K-Ar, on ???	8.4 ± 0.3 Ma
		[52]	Ar-Ar on amphibole	3.86 ± 2.25 Ma
		[145]	K-Ar, whole rock	9.7 ± ??? Ma
Ayvacık	~ 60 km S of Çanakkale	[86]	K-Ar, whole rock	8.32 ± 1.09 Ma
		[144]	K-Ar, on ???	9.9 ± 0.6 Ma
Eğrigöl	~ 10 km N of Dikili	[20]	Biostratigraphy	Late Miocene
Karaburun	~ 65 km WNW of İzmir	[52]	Ar-Ar on amphibole	8.3 ± 3.9 Ma
Urla	~ 40 km W of İzmir	[145]	K-Ar on biotite	11.9 ± ??? Ma
		[145]	K-Ar, whole rock	11.3 ± ??? Ma
Söke	~ 5 km W of Söke	[146]	K-Ar, on ???	6.99 ± 0.22 Ma
Bodrum	~ 10 km W of Bodrum	[52]	Ar-Ar on biotite	9.66 ± 1.02 Ma
		[93]	K-Ar, on ???	7.90 ± 0.25 Ma
		[93]	K-Ar, on ???	7.75 ± 0.25 Ma

Notes: Paton [52] appears to be the source of various Ar-Ar dates for these volcanics, reported by [20], [141], and others, which are cited as "unpublished" or as "personal communications", or are attributed to secondary sources such as [147]. At Karakaya, Tekirdağ, the 8.92 ± 0.10 Ma date reported by [141] was determined by one of us (H.G.), using the unspiked K-Ar method. The 8.90 and 8.96 Ma dates also reported in the same study were for the two splits of the same sample, whose weighted mean was 8.92 ± 0.10 Ma. For the Mahmutköy flow unit, the reversed geomagnetic polarity [142] suggests chron C3AR given the existing numerical age. The age bounds to this chron, from [41], are quoted. At Ezine, Paton [52] dated one split of one sample, but the value above and 7.11 ± 2.25 Ma were both stated for its numerical age. Reworking through the age calculation suggests that this older alternative value is a typo. The Eğrigöl basalt has been dated using the biostratigraphy of lacustrine sediment with which it interbeds.
It is evident that overall this data set is unsatisfactory: some dates were obtained using techniques (notably, whole-rock K-Ar, or Ar-Ar on amphibole phenocrysts) that are prone to systematic error due to inherited argon; some dates have very large margins of analytic uncertainty or no stated error bounds. In addition, no field descriptions have been published that are detailed enough to allow anyone to ever find precisely where many of these basalt samples were collected (and some descriptions are so limited that even deducing the volcanic field responsible can be a matter of educated guesswork). Nonetheless, this dating evidence supports the view that this phase of alkali basaltic volcanism began in Aegean coastal parts of western Turkey around the start of the Late Miocene.

Milet Formation; the Yarkındağ Formation of Yılmaz *et al.* [20]) is Early-Middle Pliocene, and the underlying clastic unit N3B4 (the upper Yatağan Formation; the Karagöl Formation of Yılmaz *et al.* [20]) is of Messinian age (Fig. 9). These proposed age assignments are thus essentially the same as those suggested by Yılmaz *et al.* [20] for this terrestrial sedimentary sequence, although they did not explain how the ages were determined. This Pliocene age assignment for sediments containing the Akça pollen assemblage is consistent with the results of Besang *et al.* [68] and Becker-Platen *et al.* [69], who reported a site SE of Konya in central Turkey where a lacustrine bed yielding this assemblage was overlain by dacite with a K-Ar date on biotite of 3.72 ± 0.07 Ma (corrected to 3.81 ± 0.07 Ma).

It follows that there has been ~ 1,200 m of net surface uplift in this region since the Early Miocene, and something of the order of ~ 900 m since the Middle Pliocene (when deposition ceased), depending on the palaeo-altitude of the depositional surface at the time (this point is discussed further below). These amounts of uplift clearly have nothing to do with normal faulting, as no major Late Cenozoic normal faults are known in this region (Fig. 1). They thus evidently reflect regional uplift instead. Supporting evidence for such large amounts of regional uplift in this part of SW Turkey comes from the terraces of the River Dalaman, which flows SW from the vicinity of Acıpayam (Fig. 2) to the Mediterranean coastline. At Akköprü (Fig. 1) this terrace staircase has been mapped by Doğu [72], and interpreted in terms of oxygen isotope stages (OIS) by Demir *et al.* [73] (Fig. 10). About 300 m of surface uplift since the late Early Pleistocene (OIS 22; 870 ka) is thus indicated. The resulting estimated time-averaged rate of ~ 0.3 mm a^{-1} is the same as can be estimated for ~ 900 m of uplift since the Middle Pliocene (~ 3 Ma).

5. The Denizli Basin

The Denizli Basin is at present an asymmetric graben, ~ 50 km long and ~ 20 km wide, with WNW-ESE trend (Fig. 2), adjoining the eastern end of the Büyük Menderes Graben (Fig. 1). The most detailed study of it to date is that by Westaway [33], although this work does contain some mistakes—not least because it relied on the old Seyitoğlu and Scott [11] chronology. This basin is bounded by escarpments at both margins, which have been interpreted as the footwalls of segmented normal faults, and accommodates N-S extension (e.g., [33]). In the eastern part of this basin, both these fault sets seem to be active, as seismological studies [33] suggest that the 19 August 1976 earthquake (M_L 5.0) occurred on the normal fault bounding its southern margin where it plunges north beneath Denizli city, whereas the 13 June 1965 earthquake (M_L 5.3) occurred on the normal fault bounding the northern margin of the basin, near its eastern end. The geomorphology, notably the much clearer footwall escarpment that reaches much higher altitudes (Fig. 2), and the typical southward tilt of the basin sediments [33], both suggest that the normal faulting bounding the southern margin of this basin is the more important of the two sets. Honaz Dağ, which rises to 2,571 m above sea level near the eastern end of this footwall escarpment (Fig. 2), indeed has the highest summit anywhere in the interior of the Aegean extensional province in western Turkey.

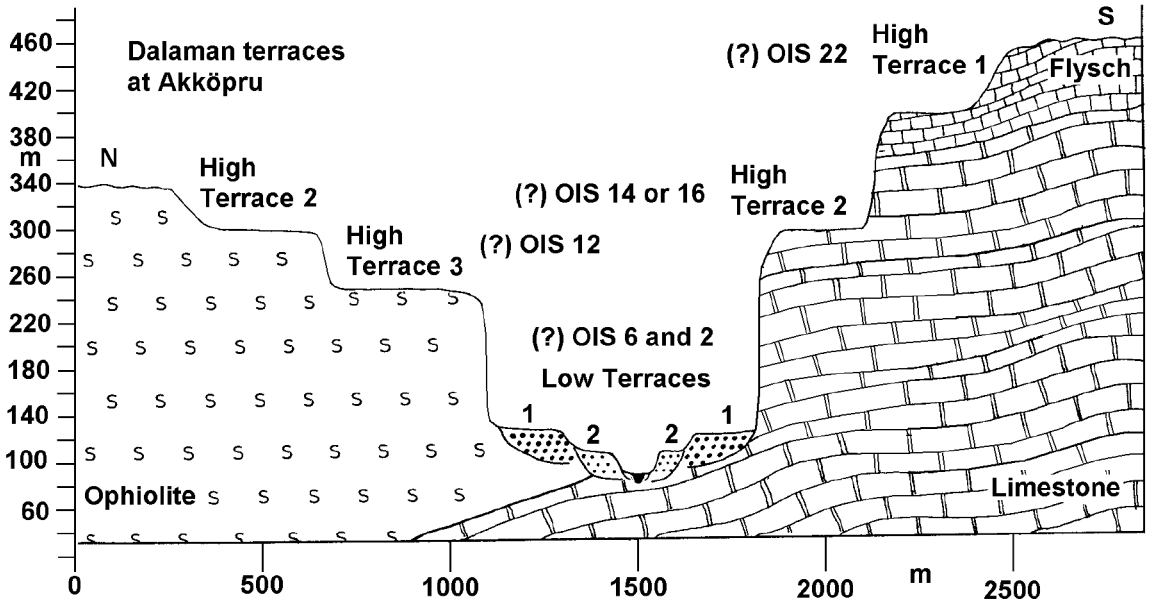


Fig. 10 Transverse profile through the terraces of the Dalaman River around Akköprü (Fig. 1). Based on Fig. 11b of [73], adapted from Doğu [72], showing tentative OIS assignments.

Sediment known as the "Denizli Molasse" (e.g., [43, 44]) crops out outside the northern margin of this basin, in and around the Sağdere river valley (J in Fig. 2). Now formally designated as the Acıdere Group, this sequence comprises conglomerate overlain by deltaic and marine sandstone, mudstone, and limestone of Late Oligocene - Early Miocene (Chattian-Aquitanian) age [43]. An equivalent sequence, known as the Acıgöl Group, crops out along the northern flank of the Acıgöl Graben and in the footwall of the Baklan Fault, notably between Baklan and Bozkurt (K in Fig. 2; [56]). These sediments are also considered to be lateral equivalents of the Akçay Group or "Kale-Tavas Molasse" in the Kale-Tavas Basin farther southwest [43]: collectively, these outcrops delineate a zone ~ 160 km long and ~ 20 km wide bordering the SE margin of the metamorphic basement of the Menderes Massif. It is accepted that these sediments pre-date the present phase of crustal extension in western Turkey (e.g., [43]).

The next sedimentary unit, the Belevi Formation of Akgün and Sözbilir [43], represents the bulk of the sedimentary fill in the Denizli Basin. What appear to be its stratigraphically lowest units crop out outside the NE margin of this basin, overlying the Acıdere Group, being exposed along the main road between Denizli and Uşak as it crosses the upland between the Denizli Basin and the Baklan Graben (G-H in Fig. 2). As described by Westaway [33] the exposure along this section consists typically of red-weathered conglomerate and marl. This outcrop is contiguous with that of the Kızılören Member of the Çameli Formation described by Şenel [56] nearby, around Baklankuyucak (F in Fig. 2); these two names evidently represent different parts of the same deposit.

Similar red-weathered clastic sequences, of diverse composition comprising fluvial sediment in some localities and lignite-bearing lacustrine sediment elsewhere, have also been reported farther west within the Büyük Menderes graben (e.g., [11, 14, 16, 45, 74]). These sediments have been assigned to the Eskihisar pollen biozone (e.g., [11, 74]), with nominal age 20-14 Ma (e.g., [13]). Several studies (e.g., [11, 45, 74]) have argued that these sediments mark the start of the present phase of crustal extension in this graben. However, the criteria used to justify this, such as lateral variations in thickness [45], are hardly compelling. Bozkurt [16] indeed noted that there is no convincing structural evidence for extension during deposition of these sediments in what is now the Büyük Menderes graben; he thus regarded them as indicating a depocentre that was unrelated to this extension, which was later fortuitously cut through by normal faulting. The disposition of these red clastics outside the NE margin of the Denizli Basin bears no relationship to normal faulting (Fig. 2), like that of the Kızılören Member of the Çameli Formation (already noted) around the Baklan Graben. We thus conclude, like Bozkurt [16], that these red clastics predate the start of extension in the Denizli Basin. As already noted, we regard them as older than ~ 7 Ma, based on the mammal biostratigraphy at Mahmutgazi (Fig. 2).

The bulk of the sediment within the Denizli Basin is carbonate-dominated: lacustrine limestone and marl (e.g., [33]), also assigned to the Belevi Formation of Akgün and Sözbilir [43]. This carbonate sequence appears to stratigraphically overlie the older clastic sequence at some points around the Denizli Basin margin. This relationship is confirmed by evidence from the western part of the basin between Sarayköy and Kızıldere (Fig. 2), where a geothermal borehole has penetrated ~ 1,100 m of Neogene sediment with the red clastics at the base [75]. However, elsewhere Westaway [33] noted this Neogene carbonate sequence in direct contact with crystalline Mesozoic limestone: notably near the SE margin of the basin at L and in the western part of the basin interior at M (Fig. 2). The lacustrine limestone, which often contains bulrush fossils (e.g., at E in Fig. 2), is typically overlain by brackish-water marl; in some places (e.g., near L in Fig. 2) an angular unconformity is evident between the two [33].

Pamir and Erentöz [76] estimated from the ostracod biostratigraphy that this marl was deposited in a brackish-water environment of "Pannonian" age, which in modern usage (e.g., [41]) would mean ~ 10-7 Ma. However, Pamir and Erentöz [76] appear to have not recognised the subsequent Pontian stage of the Paratethys as distinct from their "Pannonian". As a result, their age assignment could mean anywhere between ~ 10 Ma and ~ 5 Ma. The fact that this marl is typically tilted less than the underlying limestone ([33]; as evidenced by the angular unconformity near L) suggests that marl deposition post-dates the start of local extension. It is thus possible that the switch from lacustrine limestone to brackish-water marl deposition marks the increase in aridity expected during the Messinian stage of the Late Miocene.

Near its top, the lacustrine carbonate sequence gives way to fluvial conglomerate, for instance around locality P on the Sarayköy-Babadağ road at [PB 6645 9280], revealed by upward coarsening of the deposits and by the regular bedding in the lacustrine sequence being superseded by fluvial cross-bedding. Elsewhere (e.g., N in Fig. 2; [33]), repetitive alternations of superimposed conglomerate sheets and marl beds occur in the uppermost stratigraphic levels in this sequence that are exposed. In some localities these deposits can be observed unconformably overlain by, or inset by, younger unlithified fluvial sands and gravels, for instance around N [33], at ~ 510 m above sea-level; ~ 270 m above the Çürük River, and near P (at [PB 6662 9358]), at ~ 420 m above sea-level, ~ 280 m above the Çürük—Büyük Menderes confluence. Both these distinct types of fluvial deposits (i.e., superimposed and cemented, and inset and unlithified) have previously been lumped together as the "Asartepe Formation" (e.g., [43]). However, it is evident that the first type is part of the stacked sedimentary sequence, whereas the second type represents high river terraces deposited during the subsequent incision. A similar distinction has been observed (e.g., [3, 4]) on the upper reaches of the Gediz River in the NW Uşak-Güre Basin near Uşak (Fig. 1); another locality where both types of fluvial deposit had also previously been

lumped together as the "Asartepe Formation". Westaway [33] tentatively suggested that this switch from stacked deposition to fluvial incision in the Denizli Basin occurred in the Late Pliocene, a view that remains entirely reasonable (see below).

The characteristic badland landscape developing in the eroding sediments of the Neogene lacustrine sequence in the south of the western part of the Denizli Basin interior (south of Sarayköy; Fig. 2) suggests that active normal faulting has migrated from the fault at the southern basin margin around Babadağ to other faults farther north in the vicinity of Sarayköy (Fig. 2). This is a very widely-observed morphology in active normal fault zones in the Aegean region (e.g., [77]). Jones and Westaway [78] first tentatively suggested, from structural observations in the Büyük Menderes graben, that the development of these characteristic "uplifted Neogene basins" that flank modern graben interiors, in the footwalls of a new set of active normal faults, began at ~ 1 Ma. This view was subsequently confirmed by mammalian biostratigraphy ([40, 79, 80]; see also [14] and [16]), which shows that the uppermost sediments in these "uplifted Neogene basins" typically contain small mammals attributable to the Early Biharian mammal biozone that spans the Early Pleistocene. Westaway [34] suggested that this characteristic change, from extension on moderately steeply dipping normal faults to extension on steeper normal faults that reach the Earth's surface in the hanging wall of the earlier set, occurred as a result of climate change around the end of the Early Pleistocene. Many studies (e.g., [81]) have noted that the first really large northern hemisphere glaciation, comparable to the largest Middle and Late Pleistocene glaciations, occurred in the Late Early Pleistocene in OIS 22 (~ 870 ka). One would expect the resulting periglacial conditions that developed in western Turkey at this time to lead to reduced vegetation cover, causing increased rates of erosion and sediment transport into normal fault zones from their flanks. The components of the stress tensor associated with footwall unloading due to erosion and hanging-wall sediment loading will cause the maximum principal stress at depth to rotate so it plunges towards the footwall [34]. This change can thus inhibit slip on existing normal faults that are now misaligned relative to the stress tensor, so new, steeper, normal faults are required to form for extension to continue [34]. Abundant evidence now exists for increases in erosion rates in and around the Aegean region around this time (e.g., [4, 29, 82, 83]).

The ~ 300 m difference in altitude between the river terrace deposits mentioned earlier (at N and P) and present river level has of course been caused (at least in part) by slip on active normal faults along the line Honaz—Üzerlik—Sarayköy (Fig. 2). However, the footwall of this fault zone is back-tilted, at up to ~ 20° [33], so at least part of its local vertical motion will die out before these localities are reached. At this stage we can see no way of dating these high river terrace deposits precisely, but given the earlier discussion we anticipate that they formed either before or shortly after the end of the Early Pleistocene.

Fossiliferous marine beach rock was first noted by one of us (R.W.) in 1989 (at [PB 8065 8847], ~ 420 m altitude), in the Denizli Basin interior ~ 5 km west of Denizli city centre (A in Fig. 2; Fig. 4c), and was reported by Westaway in 1993 [33]. This deposit is both underlain and overlain by marl, and tilted by post-depositional normal faulting. D in Fig. 2 (at [QB 0150 8830], ~ 460 m altitude) marks another similar exposure, capping Gözlek Tepe hill in the eastern part of the Denizli Basin. The deposit at A provided the first clear evidence that western Turkey has experienced regional surface uplift during the Late Cenozoic, not just local elevation changes caused by active normal faulting: in the absence of regional uplift, local subsidence would be expected within a graben. Given the view at the time [11, 12] that extension began in western Turkey in the Early-Middle Miocene, Westaway [33] suggested that these marine deposits were Late Miocene (Tortonian), the subsequent transition to brackish and fluvial environments possibly marking the Messinian regression. However, as Yılmaz [35] and Westaway *et al.* [3] have noted, it now seems more likely that this marine incursion marked the resumption of marine conditions in the Mediterranean basin after its Messinian regression. This revised Early Pliocene timing is consistent with the estimated timing of extension and other deductions in our present study. We thus presume that the early part of this phase of extension, in the Messinian, created the overall normal-fault-bounded form of the Denizli Basin, and lowered its interior relative to its external surroundings. As a result, the sea was able to flood eastward into this basin in the Early Pliocene, possibly only for a brief time before this marine connection was eliminated by the subsequent regional uplift. This interpretation thus requires a narrow but ~ 200 km long marine embayment along the hanging-wall of the Büyük Menderes Graben (Fig. 1), creating a geometry similar to that at present on some active normal fault zones in central Greece (such as the Gulf of Corinth east of Patras, or the Northern Gulf of Evvia between Evvia island and the mainland of Central Greece; Fig. 1). However, we are not aware of any decisively age-diagnostic fossils in these Denizli marine deposits (see Fig. 4c caption). No marine deposits have been reported in the Büyük Menderes Graben, but their expected stratigraphic position—buried beneath many hundreds of metres of younger sediment near the base of its Neogene and Quaternary sequence—can explain this. Assuming this Early Pliocene age, the subsequent time-averaged uplift rate of this area has thus been at least ~ 0.08 mm a^{-1} (~ 420 m/~ 5 Ma). However, given the difficulties already noted concerning resolving local effects of active normal faulting, no younger local evidence currently exists to constrain any variations in this rate.

6. Discussion

We have suggested that the present phase of crustal extension in the Denizli region of SW Turkey began at ~ 7 Ma.

The best evidence for this timing comes from our new K-Ar dating (Table 2) of the Karaçay volcanism to ~ 6.7 Ma, given the geochemical evidence that this volcanism was extension-related and the field evidence that it immediately post-dates the local start of extension. The start of extension in the adjacent Baklan Graben seems to be marked by the end of deposition of the Kızılören Member of the Çameli Formation, which is also dated to ~ 7 Ma from its mammal biostratigraphy. Of course, we are not suggesting that the earliest extension anywhere in the Aegean region began at ~ 7 Ma. For instance, there is clear evidence from biostratigraphic dating of hanging-wall sediment that extension (on fault sets that continue to be active) began in SW Bulgaria during the Meotian stage of the Late Miocene (e.g., [84]), between ~ 10 and ~ 7 Ma, or possibly late in the preceding Sarmatian stage. In localities near the Aegean coastline of Turkey, the start of basaltic volcanism, which in some localities is highly alkaline, and thus generally considered to be extension-related, is dated to ~ 12-11 Ma (Table 3). A decade ago, Seyitoğlu *et al.* [13, 85] noted this change in character, from the widespread, much less mafic, volcanism in the Early-Middle Miocene, but did not draw what would appear to be the obvious conclusion—that this evidence indicates that extension began in these coastal regions around 12 Ma. On the contrary, in 2000 Aldanmaz *et al.* [86] suggested that this basaltic magmatism, which they noted began in the Aegean coastal part of western Turkey around the start of the Late Miocene, has been extension related, and regarded the very widespread, but less mafic, older calc-alkaline volcanism of western Turkey as subduction-related in the absence of contemporaneous extension. It is thus possible that extension began around ~ 12-10 Ma in these more centrally-located parts of the Aegean region, then spread eastward to the present study region at ~ 7 Ma. A related issue concerns the extent to which this phase of Aegean extension has been syn-kinematic with slip on the NAFZ. Westaway [1, 2] has argued that the NAFZ became active at ~ 7 Ma, not ~ 5 Ma as most other recent studies have suggested. If so, it would appear that the earliest evidence of the present phase of extension (during ~ 12-7 Ma) pre-dated slip on the NAFZ, then when the NAFZ became active at ~ 7 Ma it was accompanied by an increase in the dimensions of the region affected by extension, and presumably also a related increase in the maximum rate of this extension. The occurrence of extension-related volcanism at Isparta (Gölcük) (Fig. 1), K-Ar dated to 4.70 ± 0.50 and 4.07 ± 0.20 Ma [87], fits this suggested pattern by indicating a small amount of additional eastward expansion of the extensional province during the Pliocene. The ultrapotassic volcanism at nearby Afyon (Fig. 1) is significantly older, being K-Ar dated to between 14.75 ± 0.3 Ma and 8.6 ± 0.2 Ma [68] (which corrects to 15.12 ± 0.31 Ma to 8.82 ± 0.21 Ma for the revised decay constants [67]), but is not thought to be exension-related (e.g., [88]). The main significance of this volcanism for the present study is that its oldest date, 15.12 ± 0.31 Ma, is from a trachytic lava flow overlying lacustrine beds that have yielded pollen of the Eskihisar assemblage, thus providing a significant age constraint for the span of time that includes this biozone (see below). According to Cihan *et al.* [89], extension did not begin in the Sandikli graben near Afyon until the Late Pliocene. Westaway [1, 2] suggested that the ~ 7 Ma initiation of the NAFZ (cf. [90]) was caused by the change in the regional state of stress that accompanied the fall in the Mediterranean sea-level at the start of the Messinian salinity crisis. This effect was shown to be feasible if the brittle upper crust in this region is not bonded to the mantle lithosphere, but instead separated by a low-effective-viscosity lower-crustal layer. The presence of such a layer with the required low viscosity has also been suggested by geodetic studies of postseismic deformation following the 1999 İzmit earthquake on the NAFZ (e.g., [91, 92]), and numerical modelling of the coupled isostatic responses to erosion and sedimentation in the Aegean region (e.g., [4, 29]).

Another evident pattern concerns the relative timing of the subduction-related and extension-related phases of volcanism. In NW and central-western Turkey, many isotopic dates (e.g., [13, 86]) indicate that the earlier phase ended around 15 Ma, ~ 3-4 million years before the earliest local evidence of extension-related volcanism (Table 3). However, in the SW, this earlier phase seems to have ended later, with some evidence of a briefer interval before the extension-related volcanism. For instance, Ercan *et al.* [93] noted that at Bodrum (Fig. 1), andesitic lavas K-Ar dated to 9.7 ± 1.0 and 9.3 ± 1.0 Ma were followed by alkali basalts at ~ 8 Ma (Table 3). Farther north, at Söke (Fig. 1), Paton [52] obtained an Ar-Ar date on amphibole of 11.29 ± 0.70 Ma for the earlier phase, compared with a local timing of ~ 7 Ma for the extension-related basalt. As Aldanmaz *et al.* [86] discussed, such changes in timing may well relate to changes in the geometry of the subducting slab beneath the Aegean region. At present, subduction-related volcanism is only observed in the far south, in and around the volcanic centres of Nisiros, Thira, Milos, and Methana (Fig. 1).

It is noteworthy that it was relatively straightforward to deduce the timing of the start of extension for the Baklan Graben, where not much extension has occurred, but this was more difficult for the Denizli Basin, where more extension has occurred. Indeed, if the Baklan Graben were not present adjacent to the Denizli Basin, and greater footwall uplift had led to erosion of the (?) Middle Miocene red clastics outside the eastern end of the northern margin of this basin (around F, G and H in Fig. 2), it would be extremely difficult to establish these sediments as unrelated to the present phase of extension. The difficulties apparent over interpreting the depositional environment of the equivalent sediments in the Büyük Menderes Graben, where much more extension has since occurred, are thus entirely understandable. Counterparts to these Middle Miocene red clastics are also known within the Alaşehir Graben (e.g., [14, 20, 21, 24, 36, 45, 94]). However, collectively, these studies present a diverse range of hypotheses concerning the relationship

between these sediments and crustal extension. Reviewing all this work would be a major task, beyond the scope of the present study; and made complicated by the different stratigraphic nomenclatures used by the different authors and by the fact that some of the published biostratigraphic dating for this sequence is wrong, because pollen samples are reworked (e.g., [40]; see below). Nonetheless, it has been generally accepted (e.g., [14, 15, 45]) that eruption of the Toygar andesite on the northern flank of this graben coincided with the early part of this sedimentary sequence. This andesitic volcanism has been K-Ar dated to 14.8 ± 0.4 Ma [95]. Application of the previously suggested general relationship between volcanism and extension for western Turkey would thus indicate that this early part of this sedimentary sequence pre-dates the present phase of crustal extension.

Figure 11 summarises our age assignments for the sedimentary sequences that we have discussed. As already noted, many studies have assumed that sedimentation in western Turkey is an indicator of crustal extension: that significant bodies of sediment can only accumulate in the presence of active normal faulting to create the necessary accommodation space. However, it is now clear that, with the lower crust highly mobile, sediment loading can dynamically create accommodation space for more sediment (e.g.,

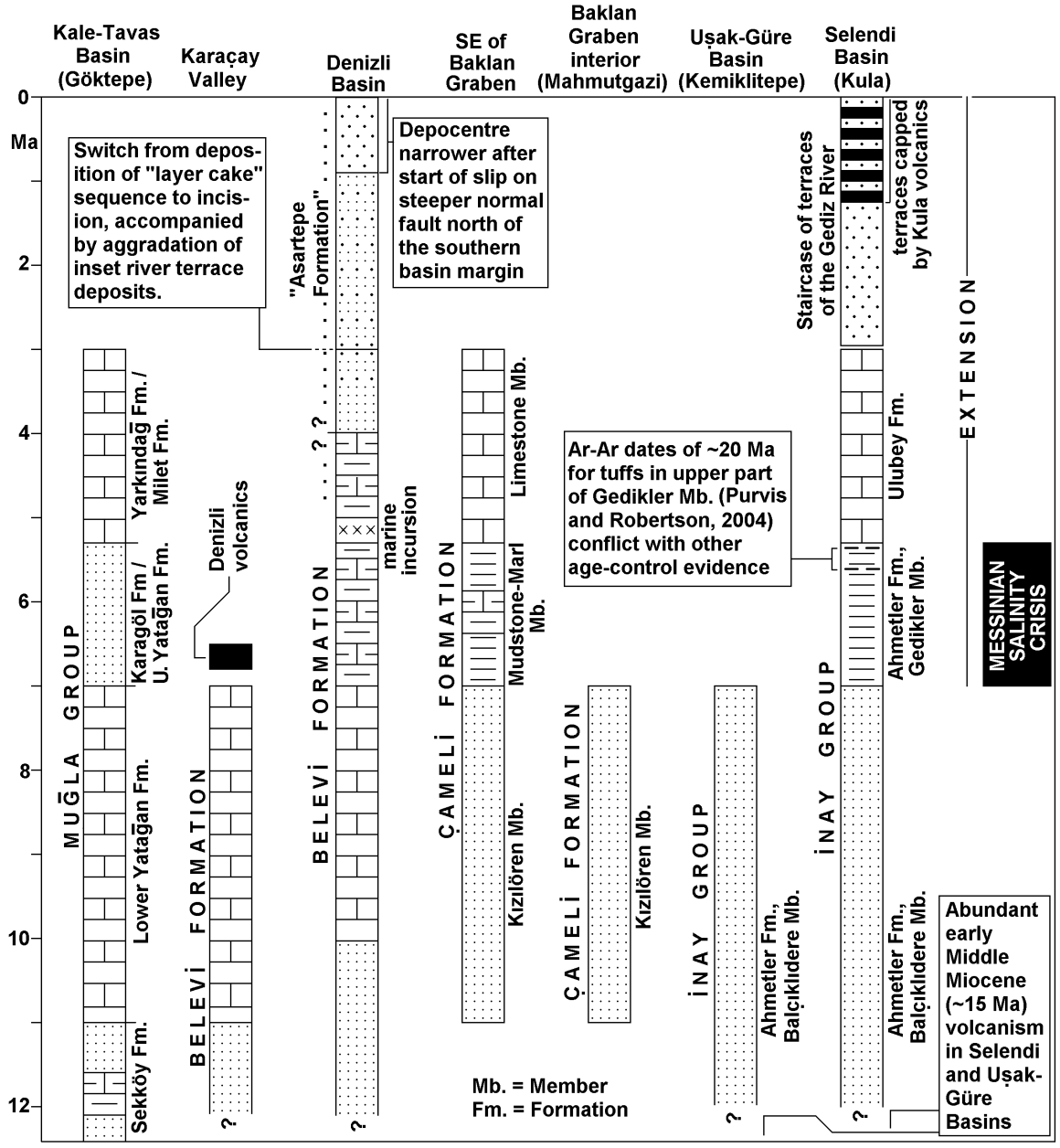


Fig. 11 Schematic stratigraphic columns summarising the sedimentary sequences examined in this study and our proposed chronological interpretation. Other nomenclatures also exist in the literature. See text for discussion.

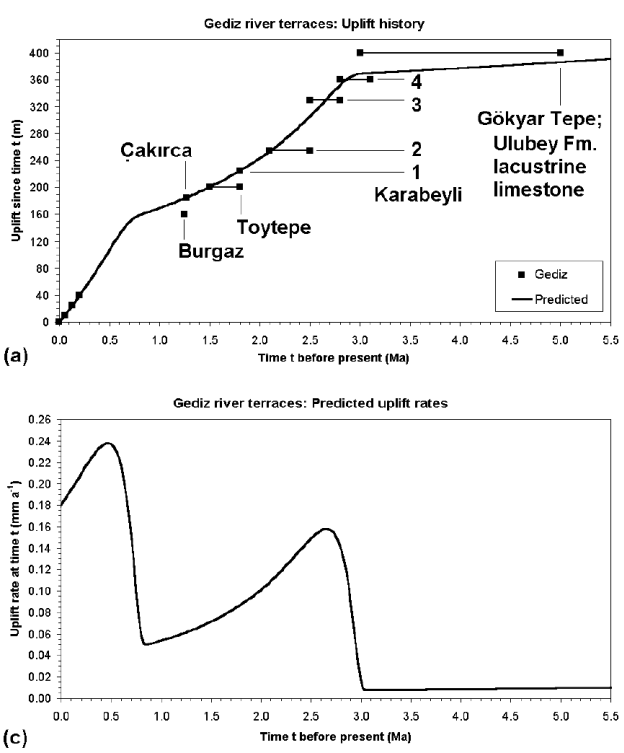
[25, 26, 27, 28, 29, 82]). As a result, there need be no requirement for sedimentation and extension to accompany each other in this region. In 2000, Yılmaz *et al.* [20] reported many similarities in the stratigraphy of Neogene basins throughout western Turkey (see their Fig. 5). In particular, they noted the typical presence of what they called the "middle unit" of sediments, deposited during the Late Miocene to Middle Pliocene. This corresponds to our sequences spanning from the Late Miocene to ~ 3 Ma, illustrated in Fig. 11. However, our deductions regarding the relationship between the deposition of this unit and the timing of extension differ from their interpretation.

Figure 11 indeed suggests that the relationship between sedimentation and crustal extension in western Turkey is much more complex than is usually assumed. In only one locality, the interior of the Denizli Basin, has sedimentation been more or less continuous from the Miocene to the present day. This locality can thus be interpreted as an instance where a graben developed in the Late Miocene, occupying part of the area of a pre-existing depocentre from before this phase of extension began. The main subsequent change is that - like in other grabens in the Aegean region - extension appears to have shifted from the segmented normal fault at the margin of this graben to a steeper normal fault in its hanging wall at a relatively late stage (Fig. 2), apparently around the start of the Middle Pleistocene, possibly in response to a change in the form of the stress field linked to the loading and unloading effects of the increased rates of sediment transport caused by climate change around this time, as suggested previously (e.g., [34]). Deposition is thus now largely confined to a relatively narrow alluvial plain along the northern margin of the earlier depocentre (Fig. 2).

Elsewhere, notably at Göktepe, SE of the Baklan Graben, and in the Kula area of the Selendi Basin (Fig. 11), sedimentation seems to have continued throughout the Late Miocene and into the Pliocene, apparently ending sometime in the Late Pliocene (a nominal age of ~ 3 Ma has been assumed). The main pattern in these sequences seems to be a switch from clastic sedimentation in the Messinian to carbonate-dominated sedimentation in the Early-Middle Pliocene. One can argue that the Pliocene resumption of a relatively moist climate after the aridity expected during the Messinian regression of the Mediterranean Sea led to regrowth of vegetation, which stabilised hill slopes causing reduced clastic input into rivers and lakes: hence this switch to carbonate deposition. Elsewhere in Turkey, a characteristic transition from continuous sedimentation in lacustrine basins to river gorge incision, tentatively dated to the "Villafranchian" or Late Pliocene to Early Pleistocene, was recognised long ago (e.g., [96, 97, 98, 99]). Like Westaway *et al.* [3, 4], we suggest that this change was caused by an increase in regional uplift rates, possibly as the isostatic response to increased rates of erosion caused by climate change in the Late Pliocene.

The final pattern, evident in the Karaçay Valley, the interior of the Baklan Graben, and at Kemiklitepe in the Uşak-Güre Basin (Fig. 11), is for sedimentation to have ceased around 7 Ma, at the start of regional extension. This is of course the opposite of what has previously been expected to happen. In each case we suspect that a unique combination of local circumstances is responsible. For instance, the Karaçay Valley is in the hanging-wall of normal faults to the east, west, and south, but also in the footwall of the more important Dereköy normal fault to the north (Fig. 3). The southward back-tilting of the sediment in this area (Figs. 3, 6) suggests that footwall uplift behind this Dereköy Fault led initially to local uplift and initiated the dissection of the sedimentary sequence in the Karaçay Valley. Additional uplift can be expected to have then been induced as the isostatic response to this erosion. In the Baklan Graben, as already noted, the main effect seems to have been incision by the throughgoing Büyük Menderes River. As suggested by Westaway *et al.* [4], it is possible that incipient slip at ~ 7 Ma on the normal fault zone bounding the northern margin of the eastern end of the Alaşehir Graben (Fig, 1) may have caused very gentle northward back-tilting of the land surface in the adjacent southern part of the Uşak-Güre Basin around Kemiklitepe, causing deposition to migrate elsewhere. It is thought that the local relief at this time was much less than at present, so very gentle tilting could have had a significant effect on drainage.

Concerning amounts of surface uplift since the Miocene, Demir *et al.* [73] noted that uplift rates in Turkey seem in general to be fastest in regions of high rainfall, such as along the eastern part of the Black Sea coastline. The obvious interpretation is that this uplift is the isostatic response to relatively high rates of erosion under non-steady-state conditions (see, e.g., [27, 28, 29, 100]). Detailed analysis of the terrace staircase of the Gediz River [3, 4] indicates that ~ 400 m of surface uplift has occurred in the vicinity of Kula and Uşak (Fig. 1) since the Middle Pliocene (Fig. 12). The ~ 400 m present-day altitude of the marine deposits within the Denizli Basin, thought to relate to an Early Pliocene marine incursion (Fig. 4c), may underestimate the regional uplift on this time-scale (as would be revealed by localities outside the Denizli Basin) due to the effect of any net component of hanging-wall subsidence resulting from slip on the normal faults bounding the margins of this basin (Fig. 2). These parts of Turkey have relatively low annual rainfall, typically ~ 400-600 mm. Much higher annual rainfall, in the range ~ 1000-2000 mm, occurs in the mountainous Lycian Taurus region to the south and southeast of Denizli, bordering the Mediterranean Sea (Fig. 1). The data that was discussed earlier (Fig. 13) indicates that this region has uplifted farther, and faster, than has occurred farther north, presumably as a consequence of higher rates of erosion resulting from the climate. Under non-steady-state conditions, such erosion is expected to reduce the local lithostatic pressure at the base of the brittle upper crust, causing a net influx of mobile lower-crustal material that leads to net crustal thickening and thus surface uplift (e.g., [27, 28, 29]). The most likely source of this material is from the lower-crust



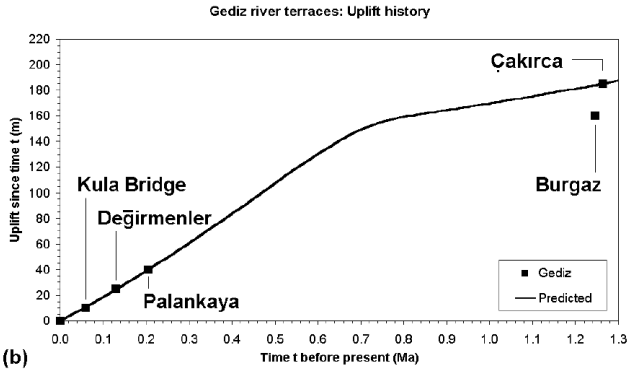
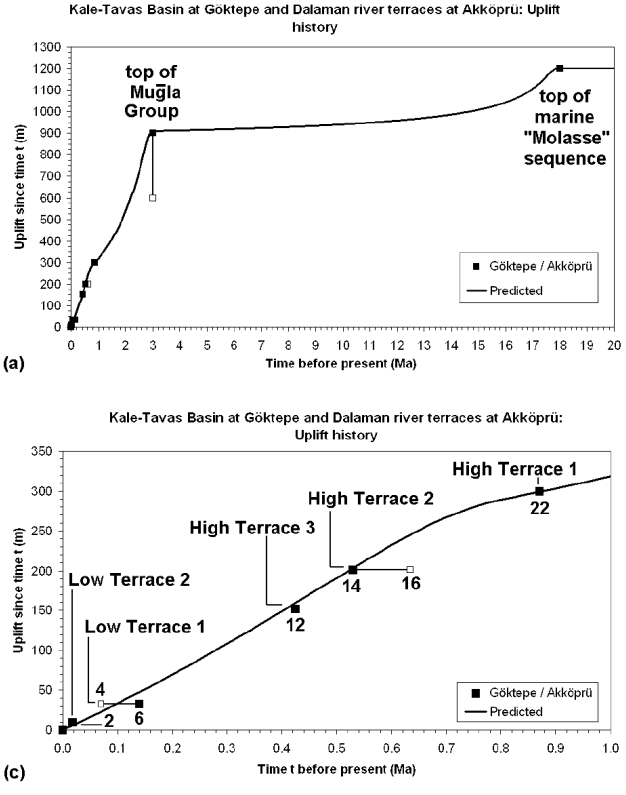


Fig. 12 Uplift history for the reach of the Gediz River around Kula (Fig. 1), adapted from Fig. 21 of [4]. Calculations follow the method described in [109] and [110], and are based on the following parameter values (defined in these references): z_b 15 km; z_i 25 km; c 20 °C km^{-1}; κ 1.2 mm^2 s^{-1}; $t_{o,1}$ 18 Ma, $\Delta T_{e,1}$ – 20 °C; $t_{o,2}$ 3.1 Ma, $\Delta T_{e,2}$ – 8.5 °C; $t_{o,3}$ 2.5 Ma, $\Delta T_{e,3}$ 0 °C; $t_{o,4}$ 1.2 Ma, $\Delta T_{e,4}$ 0 °C; and $t_{o,5}$ 0.9 Ma, $\Delta T_{e,5}$ – 11 °C. **(a)** Predicted uplift history and supporting data for the Pliocene and Quaternary; **(b)** Enlargement of (a) showing the late Early Pleistocene onwards; **(c)** Predicted variation in uplift rates for the same time-scale as (a). See Westaway *et al.* [3, 4] for descriptions of the named field localities and explanations of the nature of the available dating evidence from each. The older river terraces, which are not independently dated, have been assigned ages to bracket the predicted uplift history. See text for discussion.



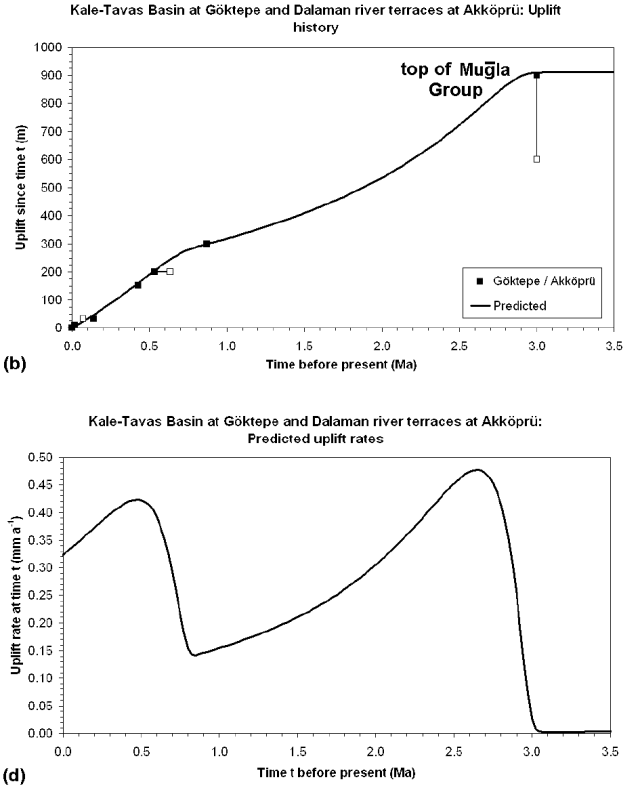


Fig. 13 Uplift history for the reach of the Dalaman River around Akköprü and for the Kale-Tavas Basin around Göktepe (Fig. 1). Calculations follow the same method using the same parameter values as Fig. 12, except: $\Delta T_{e,1}$ – 7.5 °C, $\Delta T_{e,2}$ – 26 °C, and $\Delta T_{e,5}$ – 17 °C. **(a)** Predicted uplift history and supporting data for the Middle Miocene to Quaternary; **(b)** Enlargement of (a) showing the Middle Pliocene onwards; **(c)** Enlargement of (b) showing the late Early Pleistocene onwards; (c) Predicted variation in uplift rates for the same time-scale as (b). See text for discussion.

beneath the submerged Anaximander Mountains to the south, where marine geophysical studies [101] indicate that the sea floor over the submerged continental crust has become much deeper since the Miocene, requiring net outflow of lower crust, consistent with this hypothesis.

The modelling depicted in Fig. 13 is extremely crude, and is presented solely to provide a rough indication of what appears to have happened overall in the Lycian Taurus region since the Early Miocene, to illustrate the point that there is no direct connection between the timings of phases of rapid uplift and of extension. First, it is a composite solution; the measurements of net uplift since the Early Miocene, Pliocene, and Early Pleistocene have been merged from different localities. The model solution is fitted to 900 m of uplift since the Middle Pliocene, which is consistent with the assumption that during the deposition of the Muğla Group the altitude of the sediment surface remained constant around sea-level, which would occur if this sediment load was entirely compensated by outflow of an equivalent mass of lower crust. Of course, if this deposition was accompanied by an increase in the altitude of the sediment surface, then less uplift will have occurred since this deposition ceased, which means—since the total is fixed—more must have occurred beforehand. This point is illustrated in Figs. 13a, b by showing an alternative Middle Pliocene datum, consistent with ~ 600 m of surface uplift since then (i.e., assuming that the upper surface of the Muğla Group stood ~ 300 m above sea-level in the Middle Pliocene). Likewise, no attempt is made to consider any component of subsidence of the crust beneath the Muğla Group depocentre during its deposition, as a result of sediment loading. This is justifiable in this first-order solution, because the measurement of post-Early Miocene net uplift, at Kale, is in a locality where only limited deposition of the Muğla Group occurred (see, e.g., Fig. 11 of [20]). It is also evident that the surface uplift in this region has nothing directly to do with crustal thickening associated with emplacement of the Lycian Nappes, as field relationships (e.g., [43]) indicate that in this region this emplacement had ended before the sediments discussed in this study were deposited. Localised reverse faulting associated with the final stages of Lycian Nappe emplacement did continue in localities > 100 km farther east, near Antalya (Fig. 1), until the Late Miocene (Tortonian) (e.g., [102, 103, 104]), but had evidently ceased in the vicinity of Kale long before this. The general form of uplift history depicted in Figs. 12 and 13 is also observed elsewhere in Turkey [105], and also in many other regions (e.g., [27, 106, 107, 108, 109, 110]). It is independent of any horizontal component of crustal deformation occurring in response to plate motions, and is indeed observed in many regions (such as north-western Europe; e.g., [27]) that are unaffected by plate motions.

Finally, we return to the topic that prompted this investigation in the first place: the difficulties that have arisen due to past misuse of dating evidence in interpretations of the Neogene of western Turkey. It is clear that, in the absence of isotopic dates, we have relied mainly on mammalian (and ostracod) biostratigraphy, and have not made any use of pollen. Giving priority to mammalian evidence over pollen evidence is in accordance with established biostratigraphic procedure (e.g., [111]). However, Fig. 9 also provides rough indications of where the established pollen biozone boundaries (and the designated stratotypes for some of these zones) fit into our scheme. According to Sickenberg and Tobien [39], the site at Eskihisar, the stratotype for the Eskihisar pollen biozone, is located at the boundary between the Turgut and Sekköy formations (Fig. 9). The age span of this biozone, as originally defined, was thus taken as continuing into the lower part of the Sekköy Formation. Assuming the age-control tie points indicated in Fig. 9 are correct, the age of the Eskihisar site can be estimated as ~ 12.5-13 Ma, making the young age bound of the associated pollen biozone ~ 12.5-12 Ma. This would place both these stratigraphic levels within the Astaracian mammal stage, consistent with the fauna present (Fig. 9): with the Eskihisar stratotype probably late in mammal biozone MN6 and the young age bound to the pollen biozone at the end to MN6 or in MN7. Likewise, the stratotype of the Yeni Eskihisar pollen biozone, in the upper part of the Sekköy Formation, was considered by Sickenberg and Tobien [39] to have a mammal fauna contemporaneous with or just pre-dating the Astaracian-Vallesian faunal turnover (which in the 1970s defined the boundary between the "continental" Miocene and Pliocene). This boundary between mammal biozones MN8 and MN9 is dated at 11.2 Ma by Steininger *et al.* [41]. Pending future refinements, it thus seems reasonable to place at this age the boundary between the Sekköy and Yatağan Formations, and that between the Yeni Eskihisar and Kızılhisar pollen biozones, as tentatively indicated in Fig. 9.

It has previously been suggested (e.g., [13, 37]) that the Eskihisar pollen biozone spans 20-14 Ma. This age span was determined by isotopic dating of sites away from the biozone stratotype, and was not checked for consistency with this stratotype. Taking account of the revised decay constants (from [67]), the young age bound to this biozone adjusts, given the original definition (cf. [69]), to somewhere between the K-Ar dates of 15.12 ± 0.31 Ma and 13.53 ± 0.36 Ma that have already been mentioned, and is thus consistent with Seyitoğlu's [13, 37] 14 Ma estimate. This particular tie point in the stratigraphy thus appears to be reliably constrained, and uncertainty in it is thus not a significant factor in resolving disputes between Seyitoğlu's [37] interpretations and the work of others (cf. [4]). For instance, Seyitoğlu [37] has claimed that the deposition of the İnay Group in the Uşak-Güre and Selendi Basins is confined to this pollen biozone, and thus had ended by 14 Ma. This differs somewhat from the timing suggested in Fig. 11, which places much of this Group in the Late Miocene (consistent with the mammalian biostratigraphy and magnetostratigraphy in Figs 7 and 8) and its upper part in the Pliocene. Westaway *et al.* [3, 4] have noted that some of the İnay Group pollen sites reported by Seyitoğlu [37] as indicating the Eskihisar biozone lie near its

base, and could thus have been deposited as late as ~ 14 Ma given the young age bound to this biozone, which is also consistent with K-Ar and Ar-Ar dating of local volcanism (e.g., [4]; Fig. 11). However, other pollen evidence was from much higher up the sequence, leading Westaway *et al.* [4] to suggest that it was reworked. Previous studies (e.g., [20, 40]) have indeed noted that pollen used by Seyitoğlu and Scott [11, 36] to determine biostratigraphic ages for the sediment in the Alaşehir Graben (Fig. 1) has been obtained from reworked clasts of lignite—not *in situ* beds—and thus overestimates the age of this sediment.

In addition to the investigations by Westaway *et al.* [3, 4], the stratigraphy of the Selendi Basin has recently also been studied by Purvis and Robertson [112]. Like Westaway *et al.* [3, 4], Purvis and Robertson [112] concluded that the Miocene sediment in this basin is not a simple syn-extensional sequence. However, Purvis and Robertson [112] disagree with all other chronostratigraphies that have been suggested for this basin (e.g., [4, 37, 59]). Notably, using the Ar-Ar technique on feldspar and biotite they dated a succession of tuffs, interbedded with the silt of the Gedikler Member of the Ahmetler Formation (Fig. 11), to the Early Miocene (~ 20 Ma). As is evident from Fig. 11, this chronology is fundamentally inconsistent with the dating evidence used by Westaway *et al.* [4], some of which is presented in Fig. 7, although most of this evidence was from the Uşak-Güre Basin and not the Selendi Basin. Two possibilities thus suggest themselves. Either the dating evidence from Purvis and Robertson [112] is in error, or the sequences in the Uşak-Güre and Selendi Basins cannot be simply correlated. A possible explanation for systematic error in the Purvis and Robertson [112] dating results is that volcanism in the (?) latest Miocene or Early Pliocene may have entrained mineral grains that back in the Early Miocene had cooled below their closure temperature for argon retention, without heating them sufficiently to reset their ages. Inherited argon is a major problem affecting dating of the Quaternary volcanism in the Selendi Basin [4], and it is thus possible that this region's (?) latest Miocene or Pliocene volcanism might be similarly affected. This problem was noted previously in western Turkey by Besang *et al.* [68], who tried to date a tuff in the Middle Miocene Turgut Formation (Fig. 9) near Muğla. Biotite from this sample yielded a Late Miocene K-Ar date of ~ 9 Ma, whereas muscovite, which has a higher closure temperature for argon retention, yielded ~ 52 Ma. Although neither mineral in this case yielded a reliable age, this example does illustrate significant retention of radiogenic argon from before the actual eruption age. The alternative possibility, that the sequences in the Selendi Basin and its neighbours have been miscorrelated with each other in previous studies such as Ercan *et al.* [59] and Seyitoğlu [37], seems unlikely, because these sequences are so similar, but remains possible until fieldwork designed to test this possibility is undertaken. In the meantime, the main arguments in the present study are unaffected by the dating by Purvis and Robertson [112], because they relate to the sequence in the Uşak-Güre Basin, whose chronology is secure (cf. Fig. 7), and not the sequence in the Selendi Basin.

If it is accepted as a result of this analysis that the present phase of extension began in western Turkey in the Late Miocene, then it is reasonable to consider what was occurring beforehand. Many studies (e.g., [23, 112, 113, 114, 115, 116]) have proposed that western Turkey was affected by an earlier phase of extension on low-angle normal faults in the Late Oligocene/Early Miocene. However, some of the low-angle normal faults inferred to have been active at this time are hypothetical (cf. [112]). Furthermore, as has been pointed out by Bozkurt [16] and others, other presently low-angle normal faults can be deduced, from their relationships to hanging-wall sediments, to have formed at steep dips and to have thus become back-tilted as a result of their own slip and that on successive generations of faults. For instance, many studies have noted that the southern margin of the Alaşehir Graben is bounded by a major presently low-angle normal fault. Using the data from Purvis and Robertson [112], this structure has a typical present-day dip of ~ 16°, and the oldest sediments in its hanging-wall (which may predate the start of extension on it; cf. e.g., [16]) dip at up to ~ 60°. The initial dip can thus be estimated using the standard technique of Westaway and Kusznir [117], and is ~ 64° (cf. [118]). Furthermore, the cooling history of the footwall of this fault ([115]; Fig. 14a) shows no evidence for any significant slip on it before the latest Miocene, although a much earlier phase of cooling (pre-dating all the direct evidence) cannot be excluded. A second problem is evident from the thermochronologic data in Fig. 14: the crust of western Turkey was clearly hot before the supposed phase of Late Oligocene/Early Miocene low-angle normal faulting occurred. However, theory for low-angle normal faulting [119] requires the crust to have been cold beforehand, otherwise normal faults with steep initial dips will form instead. Nonetheless, the strongest arguments in favour of low-angle normal faulting in western Turkey have been based on thermochronologic datasets such as that in Fig. 14b. As long ago as 1996, Westaway [120] suggested that the evidence commonly attributed to low-angle normal faulting in the Menderes Massif, including the observed cooling histories, may instead result simply from erosion, and Erdoğan and Güngör [121] have recently repeated this suggestion, but it is now clear that erosion on its own is not a tenable explanation for the observed cooling histories (such as in Fig. 14b) and that something important was evidently happening in this region in the Late Oligocene/Early Miocene.

It is well-known that, by effectively introducing a second layer of lithosphere between the Earth's surface and the asthenosphere, flat subduction can reduce the geothermal gradient and thus cool the crust. Examples of this effect are known from the Laramide orogeny of North America (e.g., [122]), and the Late Cenozoic evolution of parts of western South America (e.g., [123]). In such geometries, shear tractions applied across the subhorizontal plate interface will

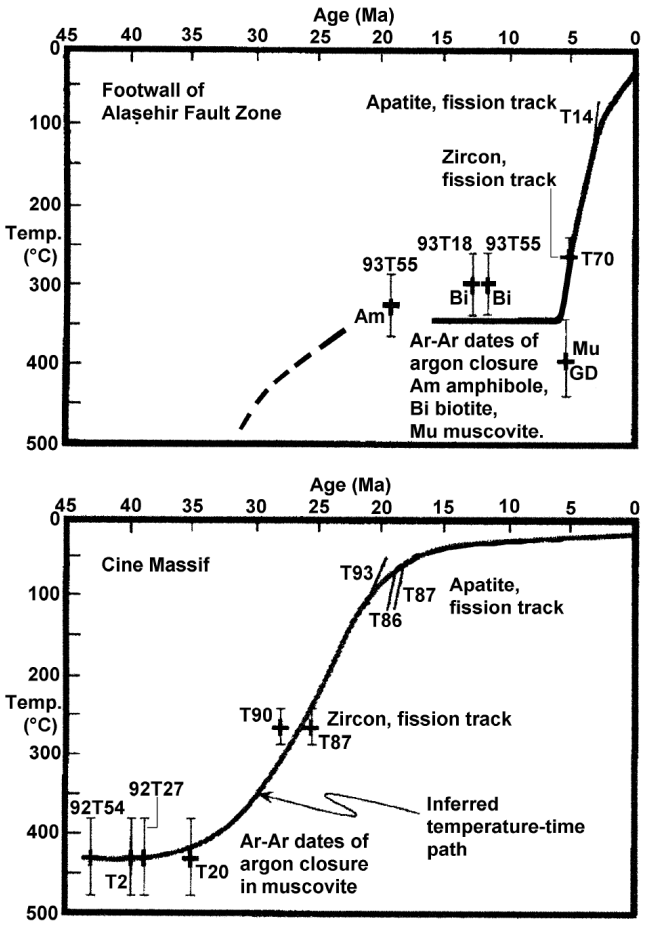


Fig. 14 Cooling histories for the crust in two parts of western Turkey, adapted from Figs. 7a and 7c of [115]. **(a)** The footwall south of the Alaşehir graben. **(b)** The Cine Massif (between Söke and Bodrum; Fig. 1), a region where no significant extension has occurred during the regions present extension phase. In (a) the closure temperatures for argon have been adjusted to ~ 400 °C in muscovite and ~ 300 °C in biotite, in lieu of the values used by Ring *et al.* [115]. For apatite fission track data only the initial parts of modelled cooling histories are shown. Codes denote sample numbers used in [115] and other studies. Thick solid lines indicate estimates by the original authors of cooling histories that fit the data. These are simple sketches, not based on any calculation. Dashed line in (a) indicates the cooling history inferred in this study before ~ 20 Ma, for which no direct evidence exists. See text for discussion.

act to cause shortening in the upper layer of lithosphere. During the Eocene, convergence between the African and Eurasian plates was accommodated by closure of Neotethyan ocean basins in what is now western Turkey (e.g., [124]). If it is presumed that around the end of the Eocene, plate convergence switched to being accommodated by flat northward subduction of the eastern Mediterranean basin beneath western Turkey, then subsequent cooling of the crust can be expected. Such a hypothetical phase of flat subduction can also explain the widespread reverse-faulting known to have occurred in western Turkey in the Oligocene (e.g., [23, 103, 125]), as a result of shear tractions exerted across the subhorizontal plate interface. Figure 15a provides an indication of the resulting cooling effect at mid-crustal depths. It indicates that if this subduction began at 40 Ma, and lasted ~ 12 Ma (after which, presumably, subduction along the Hellenic Trench adjusted to a steep plunge, as at present), then the maximum cooling effect at mid-crustal depths would be felt around 20 Ma, after which the upper crust would begin to heat up during its return to its initial temperature profile. Steepening of this subduction zone around this time has previously been considered in regional models (e.g., [113]), but not its thermal consequences. As Fig. 15b indicates, superimposition of such effects onto the cooling effect of steady erosion can result in the overall form of cooling history shown in Fig. 14b, without any requirement for low-angle normal faulting in the Late Oligocene/Early Miocene. This general explanation for this region's Cenozoic thermochronologic datasets, which has not been suggested before, will be addressed in more detail elsewhere [126].

7. Conclusions

We report evidence that the present phase of extension of the continental crust began in the Denizli region of western Turkey at ~ 7 Ma, around the start of the Messinian stage of the Late Miocene. This timing matches the estimated start of right-lateral slip on the North Anatolian Fault Zone, and corresponds to a substantial increase in the dimensions of the Aegean extensional province to roughly its present size: beforehand, between ~ 12 Ma and ~ 7 Ma, extension seems to have only occurred in the central part of this modern extensional province. In some localities, terrestrial sedimentation that began before this start of extension continued into this extensional phase, both within and outside normal fault zones. However, in other localities located within the hanging-walls of normal faults, the start of this phase of extension marked the end of sedimentation. Relationships between sedimentation and crustal extension in this region are therefore not simple, and the deposition of sediment thus cannot be used as an indicator of extension. During the time-scale of extension, the Denizli region has also experienced major vertical crustal motions that are unrelated to this extension. The northern part of this region, in the relatively arid interior of western Turkey, has uplifted by ~ 400 m since the Middle Pliocene, whereas its southern part, closer to the Mediterranean Sea and with a much wetter climate, has uplifted by ~ 1,200 m since the Early Miocene, by up to ~ 900 m since the Middle Pliocene, and by an estimated ~ 300 m since the Early Pleistocene. This regional uplift, superimposed on the local effects of active normal faulting, is interpreted as a consequence of lateral variations in rates of erosion. A reliable chronology for this region's extension in relation to changes in the geometry of adjoining plate motions, and in relation to Late Cenozoic environmental change, is now in place.

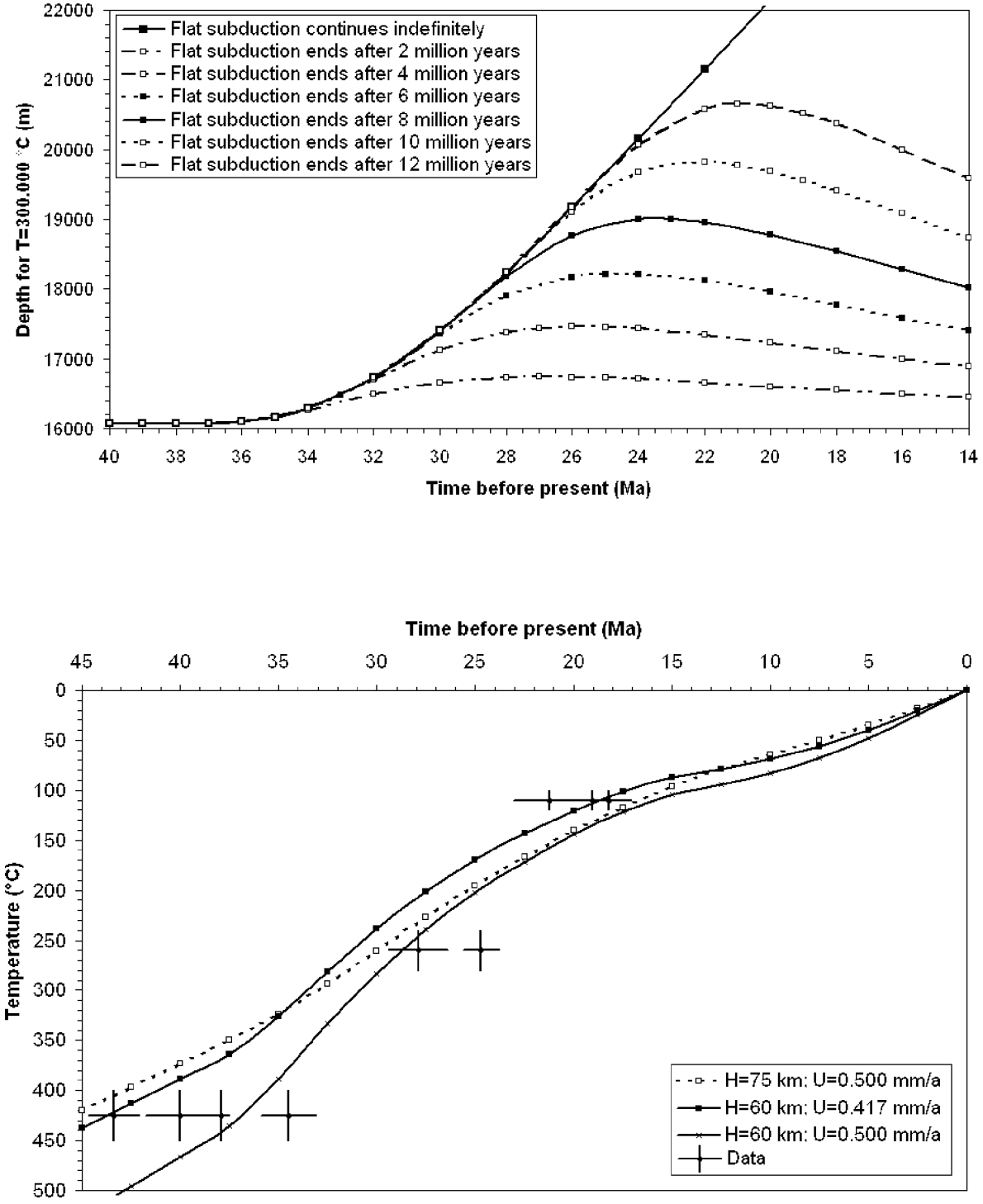


Fig. 15 Calculations are based on equations derived by Westaway [126]. The asthenosphere temperature is assumed to be 1400 °C, with a thermal diffusivity of 1.2 $mm^2\ s^{-1}$ assumed throughout the lithosphere. **(a)** Changes to the predicted thermal state of the crust of western Turkey, given the assumption that flat subduction began at 40 Ma and persisted for varying amounts of time. A layer of lithosphere of thickness H = 75 km is assumed to become underlain, as a result of flat subduction, by a second layer of the same thickness. Graph shows predicted variation in the depth of the base of the brittle upper crust, estimated as the depth at which the instantaneous temperature is precisely 300 °C. Note the increasing temperature perturbation (i.e., cooling) at mid-crustal levels the longer flat subduction persists. **(b)** A family of predicted cooling histories, assuming uniform rates of erosion, in a steady-state situation (cf. [26, 28, 29]) where each layer of material eroded is balanced by inflow of an equivalent layer of lower crust to maintain a constant crustal thickness. Flat subduction is assumed to persist from 40 Ma to 20 Ma. U is assumed erosion rate. Calculations do not incorporate perturbations to the geotherm caused by erosion (cf. [29]), only those due to flat subduction, so they are not exact solutions, but they do illustrate how flat subduction can convert what would otherwise be a uniform cooling history into one with a convex-upward profile, as is observed (Fig. 14b). Data are for the Cine Massif, as in Fig. 14b, from [115] and [148]. See text for discussion.

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