Pliocene and Quaternary surface uplift of western Turkey revealed by long-term river terrace sequences

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Western Turkey forms the eastern part of the Aegean extensional province. In the 1980s it was accepted that vertical crustal motions in this region are caused solely by this active normal faulting, with footwall localities uplifting and hanging-walls subsiding. The presence of marine sediments, interpreted as Pliocene, at altitudes in excess of 400 m in some hanging-wall localities provided – in the late 1980s – the first clear evidence of Pliocene–Quaternary *regional* surface uplift. However, it has since been argued that the incision of river gorges in this region has been caused instead by *localized* uplift in normal-fault footwalls. We review the available geomorphological and sedimentary evidence from the Denizli area, within the drainage catchment of the Büyük Menderes river, in support of ~400 m of Plio-Quaternary regional surface uplift. We also examine the gorge reach of the Gediz river near Uşak, where a staircase of four high terraces, formed of cemented fluvial gravel at ~360, ~330, ~255, and ~225 m above river level, is identified. Farther downstream, a similar terrace, ~200 m above this river and so tentatively correlated with the ~225 m terrace upstream, was also identified within the Quaternary volcanic field around Kula. Nearby, a slightly lower (~190 m) terrace gravel is capped by basalt, K–Ar dated to ~1.2 Ma; below this, other similar terraces form a lower-level staircase. We interpret this evidence as indicating uplift rates of ~0.1 mm a^{-1} or more in the latest Pliocene, when the staircase of cemented high terraces appears to have formed, relative stability for much of the Early Pleistocene, but renewed uplift at rates approaching ~0.2 mm a^{-1} in the Middle and Late Pleistocene. The resulting uplift history resembles what is observed in other regions, and has been modelled as the isostatic response to changing rates of surface processes linked to global environmental change, with no direct relationship to the crustal extension occurring in western Turkey. Our results thus suggest that the present, often deeply-incised, landscape of western Turkey has largely developed from the Middle Pleistocene onwards, for reasons not directly related to the active normal faulting, the local isostatic consequences of which are superimposed onto this 'background' of regional surface uplift.

WESTERN Turkey forms the eastern part of the Aegean extensional province (Figure 1). Forces exerted on this region by subduction of the African plate beneath its southern margin cause extension of its continental crust and pull southwestward the small Turkish plate, the motion of which relative to Eurasia requires right-lateral slip on the North Anatolian Fault Zone (NAFZ)[1]. The NAFZ became active in the Early Pliocene (~ 5 Ma)[2–6]. Its geometry near its eastern end later (~ 3 Ma) experienced a significant adjustment[7,8], but no effects of this are documented in western Turkey.

The timing of the extension of western Turkey has been controversial. In the 1980s it was accepted that it 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[9,10]. Subsequently, as better evidence emerged, the initiation of the NAFZ was placed *later* (~ 5 Ma)[2,4]. Around the same time the start of extension was adjusted *earlier*, to ~ 18 Ma (early Middle Miocene), following reports of apparently extension-related sediments with biostratigraphic and isotopic dates of this age[11,12]. However, it was subsequently realized[13–15] that the presence of Middle Miocene sediments beneath the younger fill at some localities within actively-extending grabens is fortuitous; it simply indicates that when exten-

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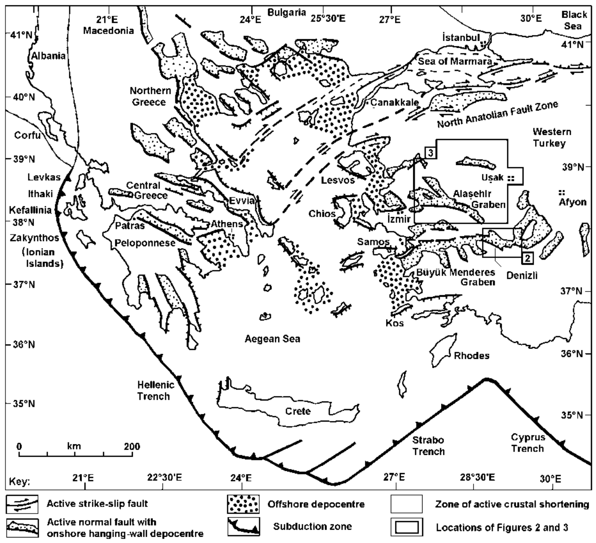

Figure 1. Map of the Aegean region, showing active faulting and related sedimentation, adapted from figure 1 of refs 19 and 39. Thick line with chevron ornament marks the surface trace of the subduction zone to the south and west of the region. Other thick lines mark significant active normal faults, with hanging-wall ticks. Fine dot ornament indicates hanging-wall sedimentary fill, which is now eroding in some localities. Coarser dots mark offshore depocentres. The right-lateral North Anatolian fault zone enters the study region from the northeast, its strands terminating against northeast-dipping normal fault zones which bound the northeast coasts of Evvia and adjacent islands.

sion began, some normal faults cut through pre-existing depocentres. Elsewhere, sediments once thought to be Miocene are now known from mammal faunas[16] to be Pleistocene. In other localities, Miocene sediment was inferred[17] to be extension-related in the absence of any structural evidence; normal faults thought to have accommodated this extension were simply interpreted along hillsides at the edges of sediment outcrops (see below). One instance of this, around Eynehan, is discussed below.

Some recent studies have suggested that this modern phase of extension was preceded by an earlier phase in the Early-Middle Miocene[13,14], these phases being separated by an interval of crustal shortening. In the localities investigated for this study (see below), Late Miocene sediments are invariably subhorizontally-bedded (except where they drape across older land surfaces or are obviously tilted by young normal faulting), providing no evidence that any significant crustal deformation was occurring then. However, we have observed abundant evidence of folded and tilted Early to early Middle Miocene sediments in localities that are distant from the present set of active normal faults (see below).

A second controversy concerns rates of crustal deformation in western Turkey. In the 1980s, many claims were made that these were very high. For instance, work using seismic moment summation[18] claimed that the slip rate on the NAFZ was ~ 40 mm a^{-1} and maybe as high as ~ 80 mm a^{-1}, and the extension rate across the Aegean was probably > 60 mm a^{-1} and maybe > 110 mm a^{-1}. Later studies[4,19], which combined careful use of this technique with structural analysis, have proposed much lower rates, such as a ~ 17 mm a^{-1} slip rate on the NAFZ and a maximum extension rate across western Turkey of ~ 3–4 mm a^{-1} with no more than a few tenths of 1 mm a^{-1} on any individual normal fault. Geodetic studies[20–22] indicate similar rates of deformation. A related controversy (which is addressed in this paper) concerns whether the vertical crustal motions occurring in western Turkey are[18,23,24] or are not[19,25–27] simply predictable as the isostatic consequences of active normal faulting.

We shall use fluvial evidence to investigate these vertical crustal motions. Rivers aggrade when they cannot transport all their sediment load, i.e. when the ratio of sediment transport to discharge is high. In Europe, they typically aggrade at times when the climate is sufficiently cold to cause reduced vegetation cover, but there is enough rainfall or seasonal meltwater for significant movement of sediment to occur[28]. In Europe, these conditions are expected during transitions to and/or from glacial maxima[28]. During interglacials, vegetation inhibits sediment transport, whereas during glacial maxima, precipitation is so low that little movement of sediment can occur. In Turkey, the climate is wetter during glacial maxima than at present[29,30]. However, much of this region is located at altitudes not far below the snow line during glacial maxima[31–33] and so will not support much vegetation at these times. It is possible, therefore, that glacial maxima, rather than climate transitions, are the main times of terrace aggradation in Turkey, unlike farther north in Europe. River terrace staircases will only form where the land surface is uplifting, as uplift will separate the gravels that aggrade during successive climate cycles[34,35]. Assuming a river develops an equivalent quasi-equilibrium profile during each phase of gravel aggradation, the vertical separation of the gravels will indicate the uplift that has occurred on the same time scale[36].

This paper examines the areas of western Turkey adjoining the headwaters of the Büyük Menderes and Gediz rivers (Figures 2 and 3). The geomorphology and disposition of fluvial and lacustrine sediments in these regions indicate hundreds of metres (most likely ~ 400 m) of regional uplift during the Plio-Quaternary, which has nothing to do with active normal faulting. This uplift is explained instead as a result of forcing of lower-crustal flow by the processes of non-steady-state erosion and cyclic loading of the crust by ice sheets and/or sea-level fluctuations[37].

The Büyük Menderes

From a source near Afyon (see Figure 3, inset), the Büyük Menderes river flows southwestward in its upper

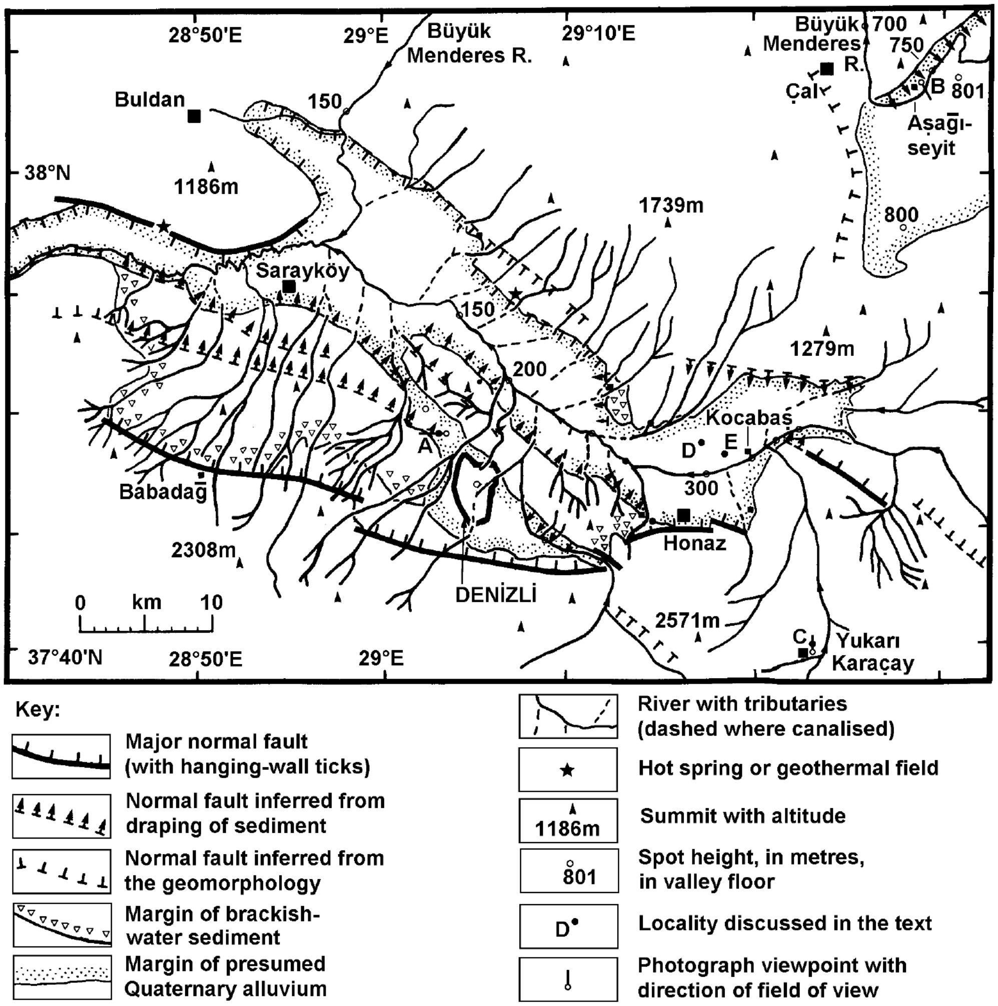

Figure 2. Map of the upper reaches of the Büyük Menderes river in the vicinity of Denizli, adapted from figure 2 of ref. 25. A, B, and C mark the locations of Figure 4 *a–c*. D marks another exposure of cemented marine beach rock, similar to A, capping Gözlek Tepe hill (at [QB 0150 8830], ~ 460 m altitude). E marks a nearby exposure of lacustrine marl with abundant bulrush fossils that was noted in ref. 25.

reaches, forming the axial drainage along the Baklan Graben. Near Çal (Figure 2), it leaves this graben, entering a ~ 70 km long gorge to the Denizli Basin where it joins the Çürük River, which forms the axial drainage of this basin. After another short gorge reach it then drains axially along the ~ 150 km long Büyük Menderes graben, westward to the Aegean coast.

Of particular interest to this study, the Denizli area provided the first indications that western Turkey is experiencing regional uplift[25], not just local elevation changes related to active normal faulting. First, fossiliferous marine beach deposits are evident within the Denizli Basin interior, most notably at a locality ~ 5 km west of Denizli city centre at ~ 420 m altitude (Figure 4 *a*). These deposits are overlain by brackish-water marl and conglomerate, which have all been tilted by post-depositional normal faulting, the sequence being capped by flat-lying river terrace deposits[25]. The presence of syn-extensional marine sediments more than 400 m above sea-level in the interior of a graben is clear evidence of regional uplift; in the absence of regional uplift, local subsidence would be expected within a graben. Second, the widespread gorge incision in this region has no clear relation to the active normal faulting, suggesting that vertical slip rates on normal faults are locally small compared with the regional uplift rate. For instance, around Aşağıseyit (Figure 2), the upper Büyük Menderes has incised a ~ 50 m deep gorge, through the base of the sedimentary fill and into metamorphic basement, in the interior of the Baklan Graben[25] (Figure 4 *b*). In addition, southeast of the Denizli Basin around Yukarı Karaçay, thick, flat-lying sediments of a Miocene lake basin are now uplifted to up to ~ 1500 m altitude and heavily dissected by river incision[38] (Figure 4 *c*). At the start of

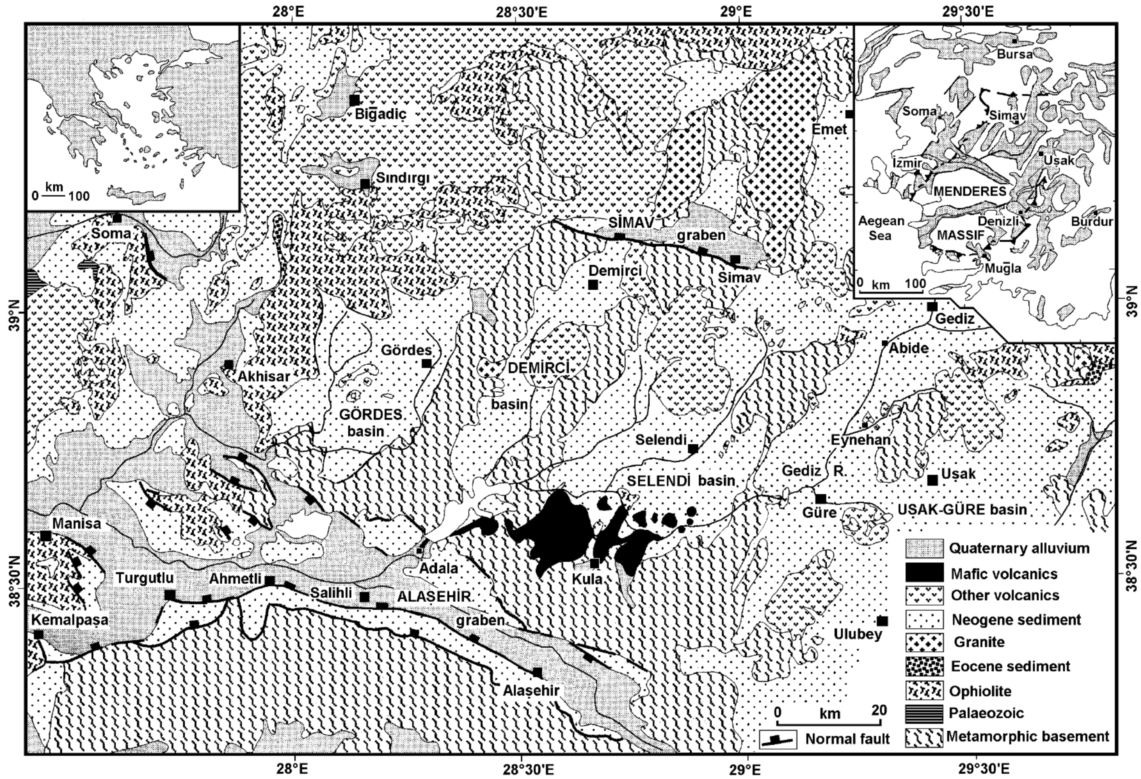

Figure 3. Map of the upper reaches of the Gediz river in the vicinity of Uşak and Kula, adapted from Figure 1 of ref. 17. Most of the outcrop labelled as 'Neogene sediment' is calcareous, and assigned to the İnay Group. In the SW Uşak-Güre Basin and southern Selendi Basin, the widespread subhorizontally-bedded outcrop is of the Balçıklıdere Member of the Ahmetler Formation[50]. This has yielded an abundant mammal fauna (ref. 50 has a full species list), which is Turolian (late Late Miocene; ~ 9.0–5.3 Ma), corresponding to mammal biozones MN11-MN13 (e.g. refs 51, 63). Among many others, this species list contains the hipparions (three-toed ancestral horses) *Cremohipparion mediterraneum* and *Cremohipparion matthewi*, which are known from mammal stages MN11–MN13, and MN12–MN13, respectively[64]. It also includes a specimen of the hyaena genus *Hyaenictis*, which is only known outside Africa from stage MN12 (e.g. ref. 65), and the bovid *Protoryx carolinae* that is only known anywhere from stage MN12 (e.g. ref. 66). A correlation with stage MN12 thus appears likely.

this erosion, as the lake sediments began to be dissected, basalt flows became trapped in these gorges[38]. These basalts are dated[38] to ~ 6 Ma; this uplift thus pre-dates the reported start of the present phase of local extension, which is Early Pliocene[13–15], and so requires a separate explanation. In 1993 it was thought[25] that the marine deposits in the Denizli basin were Late Miocene (Tortonian), the subsequent transition to brackish and fluvial environments possibly marking the Messinian regression. The revised Early Pliocene timing of the start of extension suggests that these marine conditions represent the Early Pliocene flooding of the Mediterranean basin after the Messinian event[27]. The subsequent time-averaged uplift rate of this area has thus been at least ~ 0.08 mm a^{-1} (~ 420 m/~ 5 Ma), but no younger local evidence currently exists to constrain any variations in this rate.

The Gediz

From a source north of Uşak (see Figure 1 and Figure 3, inset), the Gediz river flows southwestward along gorges incised into metamorphic basement or the floors of Neogene sedimentary basins (the Uşak-Güre and Selendi basins). Part of this reach lies within the Kula Quaternary volcanic field, where basalt flows have interacted with the river gorge and in places cap its terraces[24]. West of Kula, the Gediz enters the Alaşehir graben, along which it flows, axially, westward for ~ 80 km.

It was recently suggested[24] that the gorge incision (Figure 5 *a*) and terrace sequence along the Gediz around Kula have resulted from the river maintaining an equilibrium profile in response to footwall uplift behind the active normal fault bounding the Alaşehir graben (Figure 6). The ages of terraces, estimated from inferred field relationships to Ar–Ar dated basalt flows, were thus used[24] to determine variations in the slip rate on this fault. However, the Gediz is locally ~ 20–30 km from this fault (Figure 6). Given observational evidence from elsewhere in western Turkey[19,25] and general knowledge about the expected order-of-magnitude of the flexural rigidity (and thus, the flexural wavelength) of the upper continental crust[34,37,39], one would not expect vertical crustal motions due to faulting to persist so far. We thus suggest that this gorge incision reflects regional uplift instead.

To tie down this incision history we have obtained 6 new K–Ar dates for basalt samples, some from flows that directly overlie river terrace gravels (Table 1). Samples were crushed to a ~ 1 μm size to loosen the phenocrysts and allow the crushed groundmass to be magnetically separated. Analysis of this groundmass used the unspiked (or Cassignol) K–Ar technique[40–44], with a low blank, double vacuum furnace and an MAP-215 mass spectrometer equipped with a Faraday collector. The previous Ar–Ar analysis[45] instead dated phenocrysts, not groundmass, which could possibly cause the dates to be older than the eruption ages – as phenocrysts may have formed closed isotopic systems in a magma chamber or volcanic neck, long before eruption. In addition, several of these dated samples were from volcanic necks, up to ~ 10 km from the Gediz river. Mapping[45] indicated where each sampled flow reached the river gorge. However, other mapping[46] shows some of these flows differently, raising the possibility that dated necks[45] do not match the

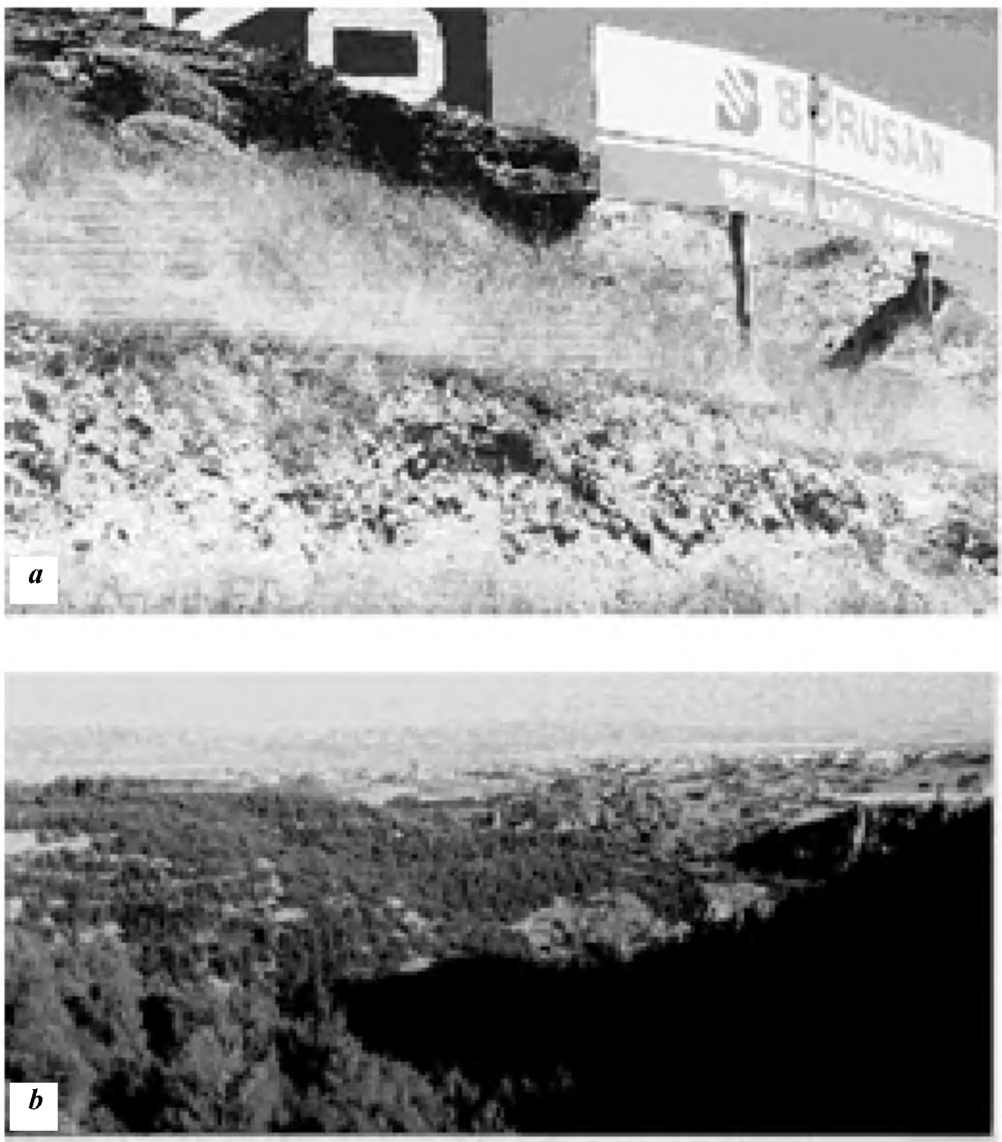

Figure 4. Field photographs from the vicinity of Denizli. ***a***, The exposure of calcareously-cemented marine beach rock between Denizli city and Sarayköy reported[25] at A in Figure 2 (at [PB 8065 8847], ~ 420 m altitude). The exposure consists of cemented sand and pebbles, containing abundant marine molluscs (predominantly of the genus *Cerastoderma*) and shell fragments. It dips at ~ 15° towards S10°W and is overlain by marl; ***b***, The view SSE from the vicinity of Aşağıseyit near the northern margin of the Baklan Graben (Figure 2), showing thin, subhorizontally-bedded lacustrine marl underlain by Menderes Schist basement, and incised to a depth of ~ 50 m by the Büyük Menderes river gorge.

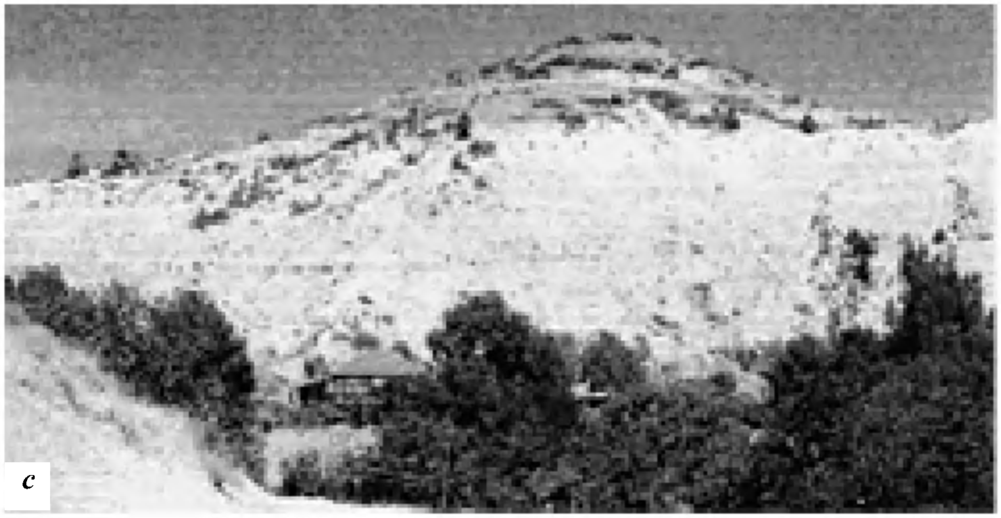

Figure 4 *c*. Field photographs from the vicinity of Denizli. View of the southern end of a N-S-trending ridge, in subhorizontally-bedded marl and lacustrine limestone, capped by ~6 Ma lamproite (basalt with phlogopite mica phenocrysts) to the north of Yukarı Karaçay (or Kocapınar). Summit is at [QB 0954 7021], at ~1350 m; viewpoint is at [QB 0932 6964], ~120 m lower.

Table 1. K–Ar dating results

Sample number	Locality	K_2O (wt%)	Mass (g)	$^{40}Ar^*$ (pmol/g)	$^{40}Ar^*$ (%)	Age (ka)	Mean age (ka)
00YM15G	Kula Bridge	3.720	0.400	0.1365	2.1	25 ± 7	
	river level		0.598	0.05703	0.9	11 ± 5	16 ± 4
00YM11G	Demirköpru	3.530	0.398	0.2528	2.4	50 ± 9	50 ± 9
00YM17G	Kula Bridge	3.890	0.401	0.4157	2.0	74 ± 15	
	youngest flow		0.600	0.2849	1.8	51 ± 11	60 ± 9
00YM12G	Adala	3.600	0.404	0.5099	2.6	98 ± 15	
			0.597	0.3458	2.2	67 ± 12	79 ± 10
00YM30G	Palankaya	2.720	0.611	0.5684	3.6	145 ± 16	
			0.202	1.172	7.5	299 ± 20	205 ± 13
00YM23G	Çakırca	2.670	0.124	4.628	55.2	1203 ± 27	
			0.603	4.914	59.0	1278 ± 13	1264 ± 15

Samples were analysed at the Argon Isotope Facility, Scottish Universities' Environmental Research Centre. Age calculations use standard decay and isotopic abundance constants[67]. Errors are estimates of analytical precision at a 68% confidence level (± 1 σ). Means are weighted by the inverses of the variances.

flows along the Gediz gorge to which ages were attributed. Apart from isotopic dating, fossil human footprints at a site just north of locality J in Figure 6 were TL dated[47], giving ages of 65 ± 7 ka for tuff below a footprint, 49 ± 9 ka for crystals of orthoclase and hornblende scraped from the footprint itself, and 26 ± 5 ka from scoria overlying the footprint. This 26 ± 5 ka TL date has been incorrectly attributed[24] to the young basalt flow unit that continues north from locality J to the Gediz river at locality I (Figure 6).

Several isotopic dates are critical for constraining this incision history. At Burgaz (G, Figure 6), there is[45] an Ar–Ar date of 1245 ± 130 ka for flat-lying basalt capping a subhorizontal land surface, with its base at ~560 m altitude, ~160 m above the river (Figure 5 *b*). A K–Ar date of ~1.2 Ma was also previously reported from here[48]. At Çakırca (sample 23, Figure 6), we dated to 1264 ± 15 ka basalt capping uncemented polygenetic fluvial gravel at 565 m, ~190 m above the river (Figure 5 *b*). At Palankaya (N, Figures 5 *c*, 6) we dated to 205 ± 13 ka basalt capping fluvial sand at 320 m, ~40 m above the river. A concordant 190 ± 50 ka date Ar–Ar date was obtained[45] for basalt at locality Q, in a flow unit that was traced to Palankaya. A 130 ± 90 ka Ar–Ar date was obtained[45] for basalt from neck 59 (K, Figure 6), in a flow unit that was traced into the contemporaneous floor of the Gediz gorge near our site 19 (Figure 6), where its base is at ~380 m, ~25 m above the river. Finally, near

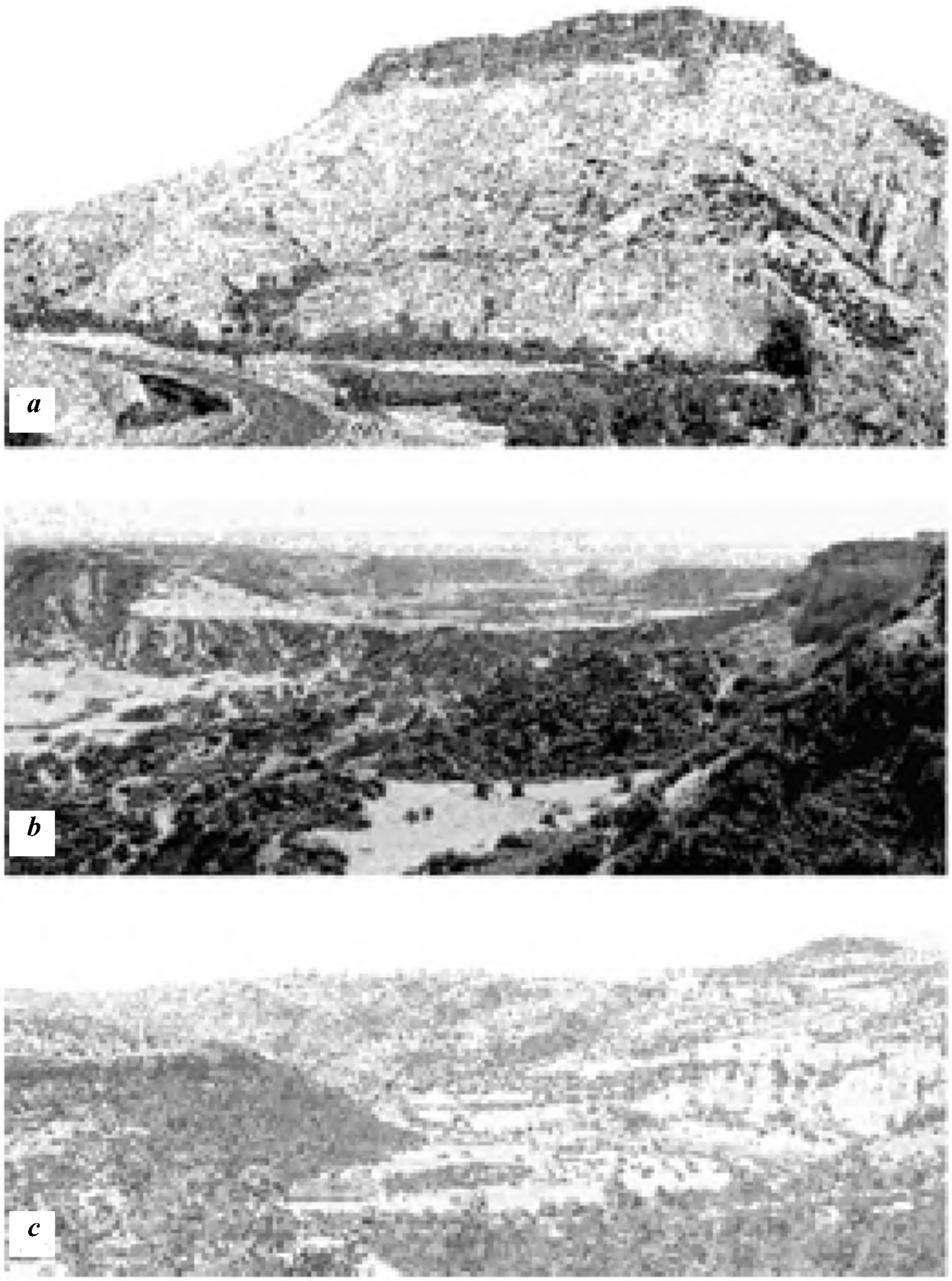

Figure 5. ***a***, View west from [PC 5820 7435], showing the basalt-capped hill Kale Tepe (H in Figure 6; summit: 593 m), and the underlying badland landscape in the subhorizontally-bedded upper part of the İnay Group (here, mainly lacustrine sand and silt). This is a well-known landmark on the main road between Ankara and İzmir, which crosses the Gediz river at the ~405 m level in the foreground; ***b***, View ESE from Toytepe (E, Figure 6), showing land surfaces capped by basalt, ~200 m above the Gediz at up to ~600 m, and a variety of lower-level sub-horizontal benches separated by abrupt scarps and areas of badland erosion of the İnay Group, which we interpret as marking a succession of levels of the Gediz during its progressive Pleistocene incision of this area. The ~200 m terrace (at ~580 m altitude) is exposed near the top of the cliff in the right foreground; ***c***, Montaged photograph looking west along the Gediz river at Palankaya (N, Figure 6): see text for discussion.

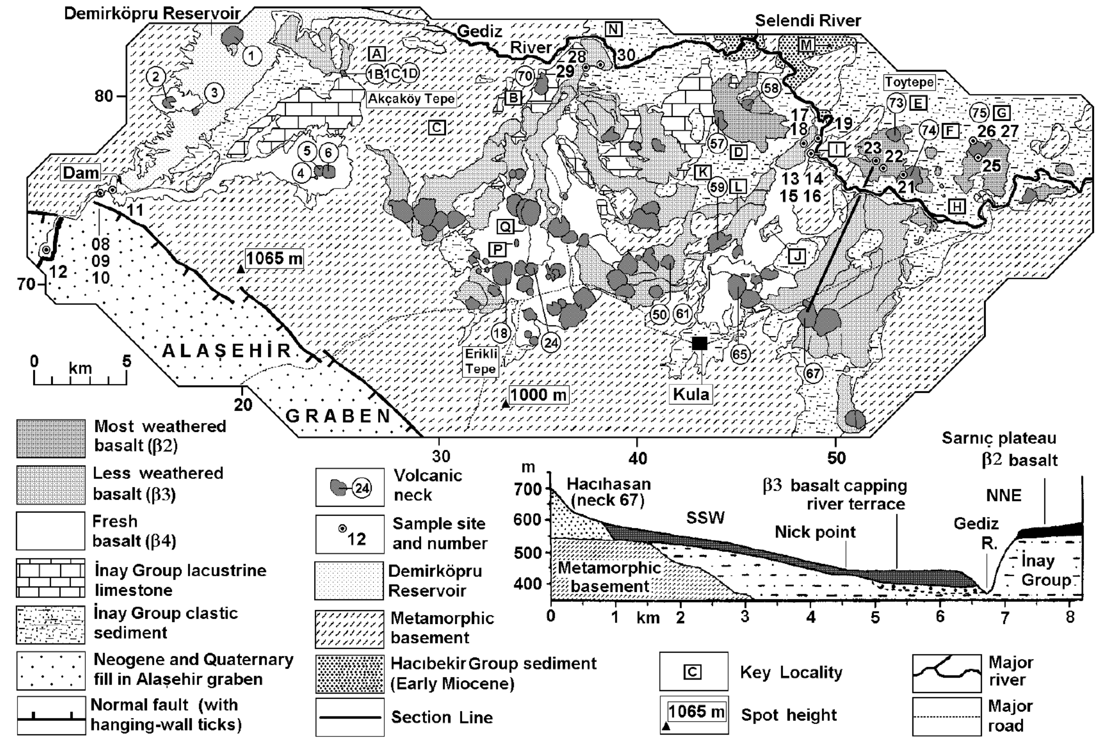

Figure 6. Map of the Gediz river gorge in the Kula area, redrawn from figure 2 of ref. 45 with additional information[17]. Inset shows transverse profile (adapted from figure 7 of ref. 46) across this gorge at Kalınharman illustrating typical field relationships between basalt flows, the Gediz gorge, and its river terraces, in this area.

Kula Bridge (sample 17 at I, Figure 6) we obtained a 60 ± 9 ka date for the youngest basalt flow unit (which covers fluvial sand at nearby site 19), its base at ~ 10 m above present river level (~ 370 m against ~ 360 m). This is the flow for which a 26 ± 5 ka age from TL dating has previously been quoted[24,45] (see above). These data are plotted in Figure 7.

The concordance between our K–Ar dates at Çakırca and Palankaya and the corresponding Ar–Ar dates[45] provides reassurance that both techniques are consistent; this dating of phenocrysts seems not to have produced any overestimation of age. We thus presume that the Ar–Ar dating of phenocrysts in the Denizli area[38] is also reliable.

Our sample 11 came from the basalt flow unit in the Demirköpru tributary valley, which reaches the Gediz just below Demirköpru Dam. This flow unit continues down the narrow Gediz gorge to Adala, where it spreads and thins on entering the Alaşehir Graben (Figure 6). Just below the dam it backfilled the Gediz gorge but has since been re-incised by ~ 50 m. Sample 11 from the top of this basalt yielded a 50 ± 9 ka date, whereas sample 12 from the top of the ~ 10 m basalt thickness exposed at Adala yielded a 79 ± 10 ka date. Combining these dates gives a weighted mean age for this flow unit of 63 ± 13 ka.

The ~ 130 ka flow unit backfilled the Gediz gorge around locality I to at least ~ 50 m above its present level. Our sample 15 was taken from the flow in the floor of the Gediz gorge at locality I (which supports the foundations of the river bridge and stratigraphically underlies the ~ 60 ka flow). However, our 16 ± 4 ka date indicates that it has not behaved as a closed isotopic system since eruption, presumably due to alteration resulting from prolonged contact with water. This is a common effect in Quaternary basalts elsewhere in Turkey[49]. We presume that this basalt was part of the ~ 130 ka flow unit nearby. The Gediz gorge was thus locally incised to roughly its present depth before 130 ka, then plugged by basalt to a thickness of at least ~ 50 m, then re-incised to within ~ 10 m of its present level before ~ 60 ka, then backfilled again by the ~ 60 ka flow unit, before being re-incised to its present depth.

These instances of backfilling and re-incision of the valley indicate that, overall, the Gediz has not maintained an equilibrium profile. A clearer instance of this effect is provided at Palankaya (Figure 5 *c*), where the ~ 60 m thick ~ 200 ka basalt appears to have filled the valley to ~ 100 m above its present level (~ 60 m above the previous terrace). Upstream of this erstwhile natural dam, this gorge is flanked by sediments (Figure 5 *c*) that appear at first sight to be typical sandy marls of the İnay Group (Figure 6). However, on closer inspection they are observed to be very large thicknesses of Quaternary sediment, incorporating large slumped masses of the Miocene bedrock; their Quaternary origin being testified by their incorporation of clasts of Kula basalt. We suggest that these sediments were emplaced when the pre-existing badlands landscape was flooded while the Gediz was impounded. The river has subsequently incised through each basalt dam, but has probably been prevented from maintaining an equilibrium profile during and immediately after major eruptions. By making use of terraces that pre-date eruptions, we hope that the mismatch between incision and uplift will be minimized.

The earlier part of the uplift history can be inferred from field evidence around Eynehan and Karabeyli (Figures 8 and 9), where we have identified four high terraces, formed of polygenetic cemented gravel, at levels of ~225, ~255, ~330, and ~370 m above the Gediz (Figure 10) The gravel in these localities had been mapped previously[17] but was regarded as forming a 'layer cake' stratigraphy (not a terrace staircase) designated as the 'Asartepe Formation'. We have also ob-

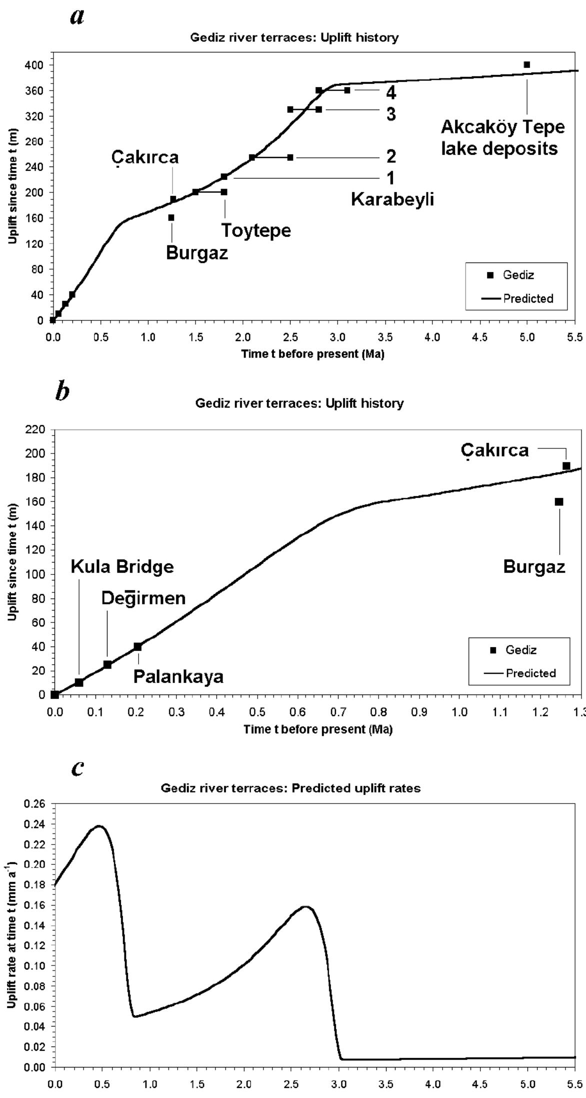

Figure 7. Uplift histories for the reach of the Gediz between Eynehan and Kula. Calculations follow the method of refs. 34 and 57, and use the following parameter values (defined in these references): z_b 15 km; z_i 25 km; c 20°C km^{-1}; κ 1.2×10^{-6} m^2 s^{-1}; $t_{o,1}$ 18 Ma, $\Delta T_{e,1}$ –20°C; $t_{o,2}$ 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 text for discussion.

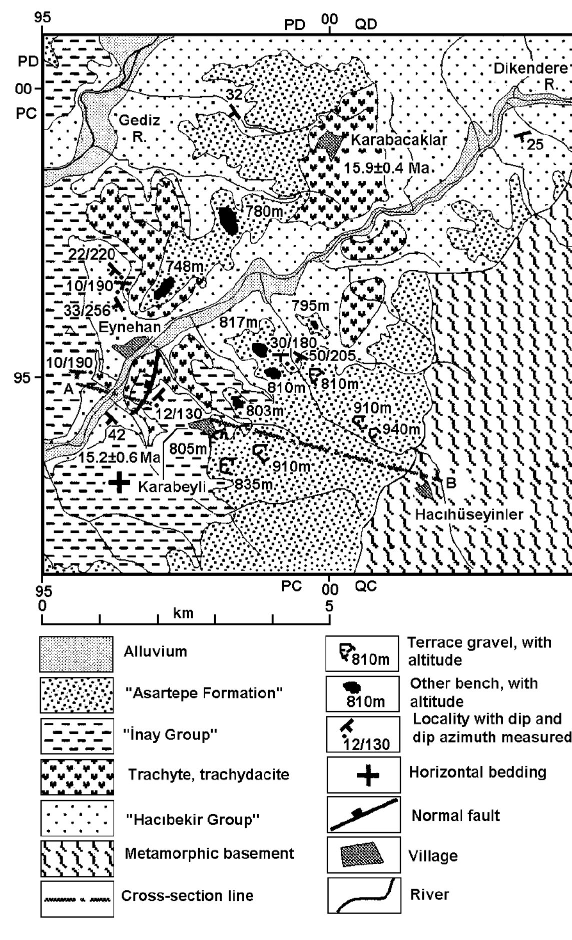

Figure 8. Map of the Gediz river and its Dikendere tributary in the Eynehan area, modified from figure 5 of ref. 17. The tilted İnay Group marl near river level is evidently Middle Miocene, as it is interbedded with lava of that age and lies near the base of the Group, just above its unconformity against the Early-Middle Miocene Hacıbekir Group: it is assigned to the Ahmetler Formation[50]. Seyitoğlu[17] mapped large areas as the 'Asartepe Formation', interpreted as thick deposits of stratified sand and gravel, deposited in the hanging-wall of an 'active normal fault' at Hacıhüseyinler. The disposition of these sediments suggests instead a sequence of four distinct high terraces of the Gediz (Figures 9 and 10).

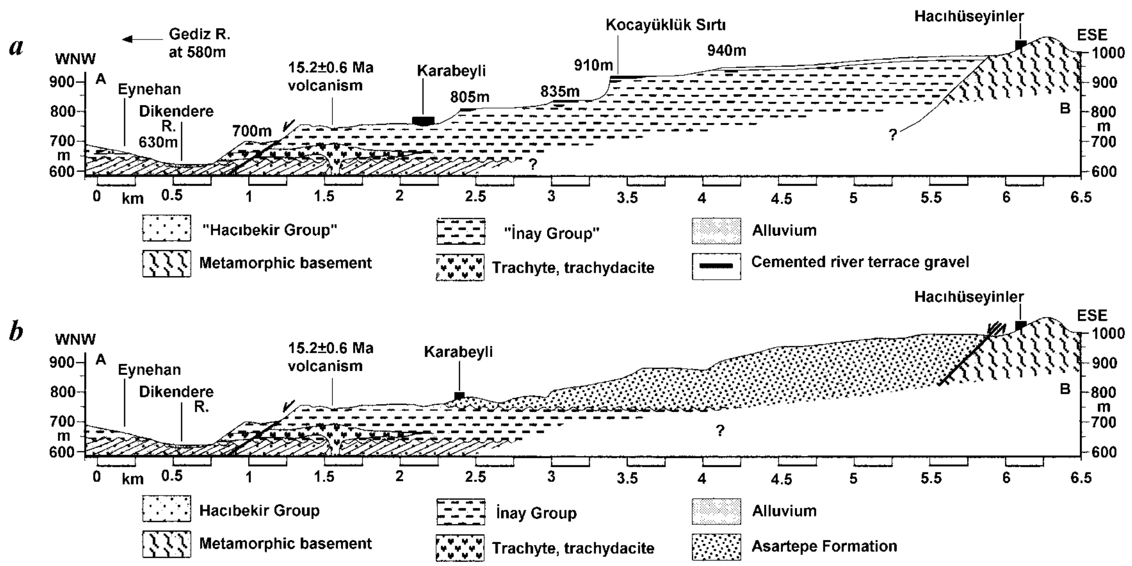

Figure 9. ***a***, Transverse profile across the terrace staircase of the Gediz through Eynehan and Karabeyli, redrawn from Figure 10 of ref. 17. ***b***, Previous interpretation of the same section[17], with the river terrace gravels and the upper part of the İnay Group lumped together as the 'Asartepe Formation', and with the contact between these sediments and basement interpreted as a normal fault that was active during deposition of these sediments. We found no field evidence that this hillside is a normal fault, let alone that it was active during this deposition. See Figure 8 for location.

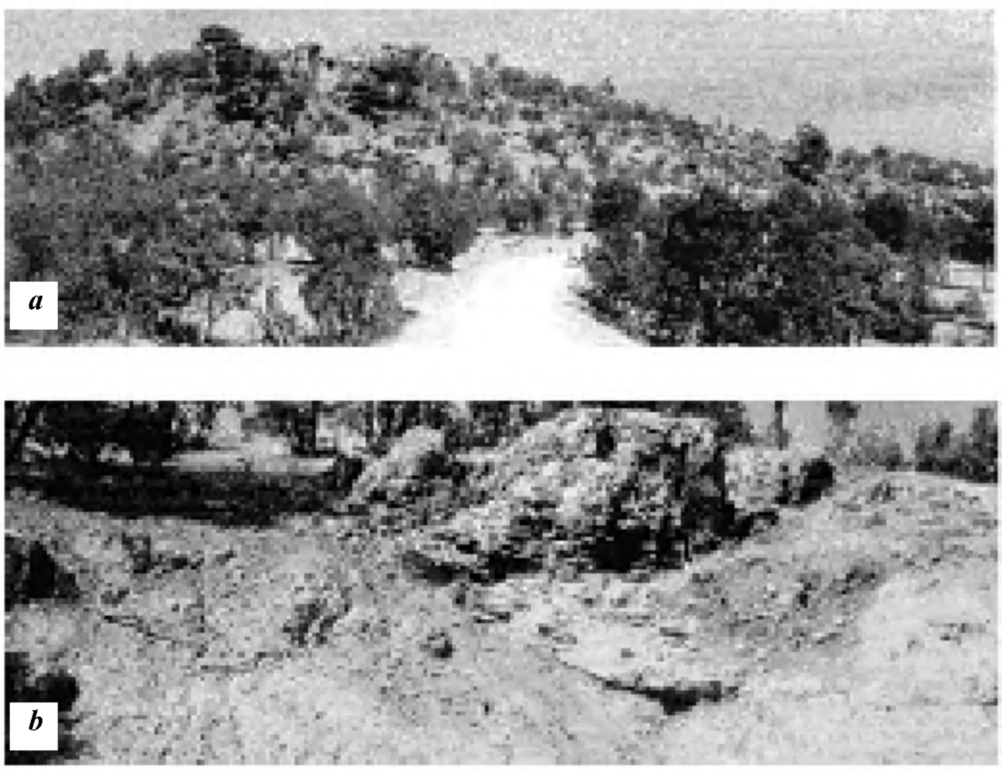

Figure 10. ***a***, Photograph of the Kocayüklük Sırtı (~ 330 m) terrace (at ~ 910 m altitude), the third youngest high terrace, showing the former channel profile with ~ 6–10 m thickness of calcreted gravel capping lacustrine deposits. View is to the NE from the terrace below at [PC 9856 9345]. ***b***, Section, at ~ 832 m altitude in the second youngest high terrace at [PC 9835 9337], through a small fluvial channel, filled with calcreted gravel consisting of clasts of quartzite, schist, and chert, cut into an alternation of white lacustrine marly limestone beds and red palaeosols.

served a similar terrace capped by basalt at Toytepe near Kula (locality E in Figure 6), ~ 200 m above the river (~ 580 m against ~ 380 m). From its altitude this may be equivalent to, or slightly younger than, the youngest terrace at Karabeyli, but it is older than the Çakırca terrace, which is closer to the river level.

In 1978, it was estimated[50] that İnay Group deposition spanned much of the Miocene and the Early Pliocene. The 'continental' Early Pliocene then included ~ 11–5 Ma, which is now regarded as Late Miocene. However, in 1997 it was argued instead[17] that deposition of this Group occupied just part of the Middle Miocene. The tilted İnay Group marl near Eynehan is evidently Middle Miocene (see Figure 8 caption). However, in the SW Uşak-Güre Basin and southern Selendi Basin (Figure 3), the widespread subhorizontally-bedded outcrop is instead of the much younger Balçıklıdere Member of the Ahmetler Formation[50] with an abundant mammal fauna (summarized in Figure 3 caption) that fits mammal biozone MN12, indicating[51] an age within 8.2–7.1 Ma. The switch from deposition to incision in the Uşak-Güre and Selendi basins thus occurred some time later, possibly around 6 Ma (as in the Denizli area; see above) or in the earliest Pliocene. The stratigraphy and chronology of the İnay Group have recently developed some significance for constraining the structural evolution of western Turkey[52,53]. It is evident that this group spans both the Middle and Late Miocene, not just part of the Middle Miocene.

The total incision by the Gediz can be estimated as ~ 400 m using the relative altitude of its highest terrace at Eynehan (Figures 8 and 9). Furthermore, near Akçaköy Tepe (A, Figure 6), lacustrine limestone, regarded as the youngest sedimentary unit in the layer-cake stratigraphy of the İnay Group[24], is observed at ~ 620 m altitude, ~ 400 m above the ~ 220 m natural level of the Gediz river. The time-averaged incision rate between times of terrace aggradation in the late Middle and Late Pleistocene can be estimated as ~ 0.2 mm a^{-1} (from the ~ 40 m incision at Palankaya since ~ 200 ka). As already noted, we have concentrated on terraces that pre-date major eruptions, when we can assume that the Gediz had developed roughly equivalent quasi-equilibrium states, such that this incision rate corresponds as closely as possible to the contemporaneous regional uplift rate. Extrapolation of this rate would place the Gediz at the ~+ 160–200 m relative level as late as ~ 1000–800 ka. However, the dating instead places it at this level at ~ 1200 ka, indicating a much lower uplift rate in the late Early Pleistocene than in the Middle Pleistocene. 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 recognized long ago[32,54–56]. We suggest that this regionally-important event is reflected in the uplift history along the Gediz, and caused the bulk of the incision between the ~+ 400 and ~+ 200 m levels to be concentrated in the Late Pliocene and early Early Pleistocene. We thus deduce that the high terraces at Eynehan aggraded at this time. Figure 7 indicates the resulting uplift history.

Discussion

Our predicted uplift history along the Gediz (Figure 7) is very different from that deduced previously[24], partly because our new evidence provides additional constraints. In particular, the acceleration in uplift in the latest Pleistocene, reported previously[24], is not part of our revised interpretation. This apparent increase in uplift rates was an artefact of using the 26 ka TL date[47] for the youngest basalt flow near Kula Bridge (I, Figure 6). As we noted earlier, the basalt itself was not originally dated, but now our K–Ar dating suggests that it is significantly older: 60 ± 9 ka. In contrast, the data now available indicates a dramatic increase in uplift rates in the early Middle Pleistocene.

Our estimate of ~ 400 m of uplift in western Turkey since the Early-Middle Pliocene is similar to the ~ 300 m observed in much of Europe[34,37,57–59], and matches the ~ 400 m in Syria on the same time scale[60]. The resulting uplift history can readily be explained as a consequence of flow in the lower continental crust induced by the combined effects of increased rates of erosion and cyclic surface loading by ice sheets and sea-level fluctuations[37,39,58], with no direct relation to the plate motions occurring in Turkey. However, the ~ 0.2 mm a^{-1} uplift rate estimated for western Turkey during the Middle–Late Pleistocene (Figure 7 *c*) is higher than in most other regions. This confirms that the uplift rate of western Turkey may be determined by the component of lower-crustal flow forced by the erosion of onshore Neogene sedimentary basins and the re-deposition of this sediment offshore[19]. The corresponding Quaternary deepening of the Aegean Sea due to outflow of lower crust to beneath the surrounding land areas is also well documented[61]. In central Greece, this mechanism seems to be even more vigorous, due to the closer proximity and thus stronger coupling between sediment sources and depocentres, and can account for up to ~ 700 m of surface uplift on this time scale[39,62].

Conclusions

Regions of western Turkey unaffected by active normal faulting have uplifted by ~ 400 m since the Early-Middle Pliocene. Like elsewhere, this uplift is regarded as a consequence of lower-crustal flow induced by cyclic surface loading and by non-steady-state erosion. Almost half this uplift has occurred during the Middle–Late Pleistocene, at ~ 0.2 mm a^{-1}. This rate is regarded as a consequence of the relatively high rate of erosion of this region, due to the dissection of its uplifting Neogene sedimentary basins.

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