Timing of extension on the Büyük Menderes Graben, western Turkey, and its tectonic implications

ERDİN BOZKURT

Middle East Technical University, Department of Geological Engineering, Tectonic Research Unit, TR-06531 Ankara, Turkey (e-mail: erdin@metu.edu.tr)

Abstract: The Büyük Menderes Graben is one of the most prominent structures of western Anatolia (Turkey) and borders the Aegean. New structural and stratigraphic evidence demonstrates that the (?)Miocene fluvio-lacustrine, coal-bearing red clastic sediments exposed along the northern margin of the graben are northward back-tilted, locally folded and overlain unconformably by horizontal terraced Pliocene–Pleistocene sediments. Also, there is no evidence that these red clastics at the base of the Neogene sequence were deposited during neotectonic extension. It is suggested here that these sediments cannot be regarded as passive neotectonic graben-fill deposits.

This new evidence further indicates that the age of the modern Büyük Menderes Graben is Pliocene, younger than previously considered (Early–Middle Miocene) and that initiation of north–south neotectonic extensional tectonics in the graben, and thus in western Anatolia, is unlikely to have resulted from orogenic collapse. The Pliocene estimate of the start of extension is in close agreement with the start of slip on the North Anatolian Fault Zone. The north–south extensional tectonics, and associated east–west faulting and basin formation, commenced during the Pliocene due to the effect of westward tectonic escape of the Anatolian block along the North and East Anatolian Faults. New mammal evidence also constrains the start of slip on the younger faults which bound the present-day graben floor to *c.* 1 Ma.

The Büyük Menderes Graben has experienced a two-stage extension. An initial extension (latest Oligocene–Early Miocene) along initially moderately, steeply dipping normal faults was superseded by movement on steeper normal faults during the (?)Pliocene. The two phases of deformation appear to reflect significant changes in the tectonic setting of western Anatolia and are attributed to orogenic collapse followed by tectonic escape.

Western Anatolia (Turkey) is a region presently dominated by approximately north–south directed continental extension. It is part of a zone of distributed extensional deformation affecting a large area (the Aegean extensional province) that includes the Aegean Sea, Greece, Macedonia, Bulgaria and Albania, and is bound by the Hellenic Trench in the south (Fig. 1). Regional Global Positioning System (GPS) data show that the central Aegean is currently moving southwestwards, relative to Eurasia, at a rate of *c.* 30–40 mm a^{-1} (Le Pichon *et al.* 1995; Barka & Reilinger 1997; Reilinger *et al.* 1997 and refs cited therein), whilst Anatolia, which is undergoing counterclockwise rotation, is escaping westwards from eastern Anatolia at a rate of *c.* 30 mm a^{-1} and is being expelled onto the African oceanic Plate along the Hellenic Trench. This all results from the collision of the Eurasian and Arabian Plates (Barka & Reilinger 1997; Reilinger *et al.* 1997 and refs cited therein).

In western Anatolia, east–west and west-northwest–east-southeast grabens (e.g. the Gökova, Büyük Menderes, Gediz, Bakırçay, Simav and Kütahya Grabens) and their related active normal faults are the most prominent neotectonic features (McKenzie 1978; Dewey & Şengör 1979; Şengör *et al.* 1985; Şengör 1987; Jackson & McKenzie 1988; Seyitoğlu & Scott 1991, 1992, 1996; Emre & Sözbilir 1995; Görür *et al.* 1995; Emre 1996; Koçyiğit *et al.* 1999) (Fig. 2). The activity of these structures is shown by the numerous historical earthquakes which have occurred along the faults (e.g. Ambraseys 1988; Ambraseys & Jackson 1998 and refs cited therein). In addition to these structures, north–south basins (e.g. the Gördes, Demirci, Selendi and Uşak-Güre basins), characterized by the widespread occurrence of Neogene sediments, are also important features (Fig. 2).

Apart from the high-angle, active graben-bounding normal faults and their role in the neotectonics of the region, evidence is available that initially moderately steeper, but presently low-angle, inactive normal faults also played an important role in exhuming the metamorphic rocks of the Menderes Massif, in controlling

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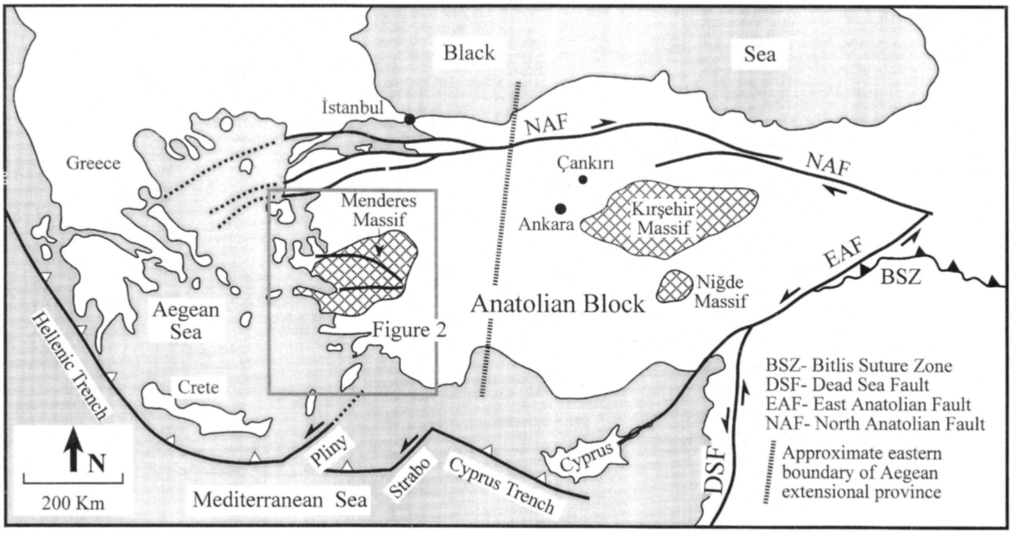

Fig. 1. Simplified tectonic map of the eastern Mediterranean region showing major tectonic elements [simplified from Barka & Reilinger (1997)].

sedimentation in the hanging-wall basins and in the consequent extension during latest Oligocene–Early Miocene phase of orogenic collapse (Bozkurt & Park 1994, 1997; Emre & Sözbilir 1995; Hetzel *et al.* 1995, 1998; Emre 1996).

The origin and age of crustal extension in the Aegean have been subjects of controversy for many years. Extension in this region has been explained by three different models: (1) the *tectonic escape model* – the westward escape of the Anatolian block along its boundary structures, the dextral North and sinistral East Anatolian Faults, since the Late Serravalian (12 Ma) following collision of the Arabian and Eurasian Plates across the Bitlis Suture Zone (Dewey & Şengör 1979; Şengör 1979, 1987; Şengör *et al.* 1985; Görür *et al.* 1995); (2) the *back-arc spreading model* – back-arc extension caused by the south-southwestward migration of the Hellenic Trench System [the mechanism of subduction roll-back; see McKenzie (1978), Le Pichon & Angelier (1979) and Meulenkamp *et al.* (1988)]. However, there is no common agreement among scientists on the inception date for the subduction roll-back process and proposed ages range between 60 and 5 Ma (McKenzie 1978; Le Pichon & Angelier 1979, 1981; Kissel & Laj 1988; Meulenkamp *et al.* 1988); (3) the *orogenic collapse model* – localized extension induced by late orogenic gravitational collapse of overthickened crust following the latest Palaeocene collision across Neotethys along the İzmir–Ankara–Erzincan Suture Zone during the Late Oligocene–Early Miocene (Seyitoğlu & Scott 1991, 1992).

More recently, Koçyiğit *et al.* (1999) proposed an '*episodic, two-stage graben model*', with an intervening phase of short-term compression for the evolution of the Gediz Graben: a Miocene–Early Pliocene first stage occurred as a consequence of orogenic collapse and a second phase of north–south extension originated from westward escape of the Anatolian block, triggered by the commencement of seafloor spreading along the Red Sea during the Early Pliocene. They consider that the intervening short-term compressional episode resulted from a change in the kinematics of the Eurasian and African Plates.

The Büyük Menderes Graben is bounded by one of the principal active normal fault zones in western Turkey. The main aspect of this paper is to propose that neotectonic extension in the Büyük Menderes Graben began in the Pliocene or later, rather than in the Early–Middle Miocene as others have claimed in recent literature (Seyitoğlu & Scott 1991, 1992). The previous age proposal was based on the age of sediments and volcanic rocks at (or near) the base of the sedimentary sequence. It is argued instead that these sediments, exhumed with respect to the present-day graben floor, have nothing to do with neotectonic extension prevailing in the region. The purpose of this paper, based on

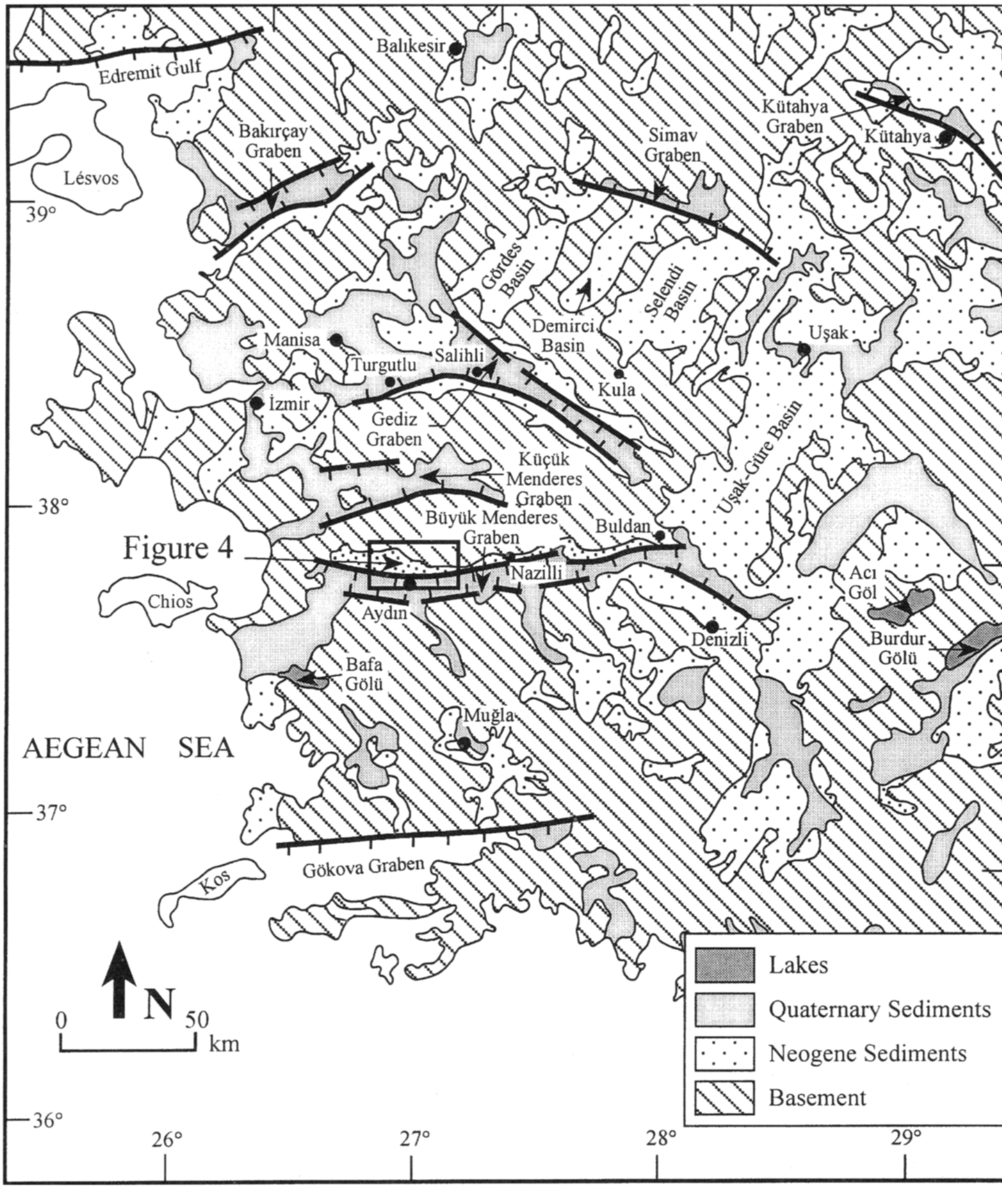

Fig. 2. Outline geological map of western Anatolia showing Neogene and Quaternary basins [simplified from Bingöl (1989)]. Note that the (?)Miocene and Pliocene sediments are not differentiated due to lack of data.

mapping, field observations and the reassessment of available literature, is therefore to present new structural and stratigraphic information from the area around Aydın (Figs 2 and 3) that bears influence on the age of the Büyük Menderes Graben, and to discuss its implications for the age and cause of neotectonic extension in western Anatolia.

Büyük Menderes Graben

Established knowledge

The Büyük Menderes Graben, one of the major east–west grabens in western Anatolia, is a structure *c.* 125 km long and 8–12 km wide. The plain in the interior of the graben consists of an

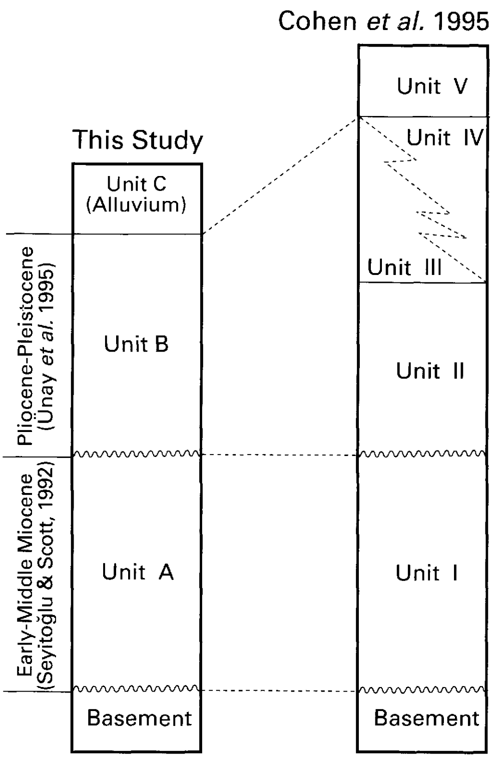

Fig. 3. Stratigraphy of the Büyük Menderes Graben around Aydın and its correlation with that of Cohen *et al.* (1995).

axial fluvial depocentre bounded to the north by a segmented, moderately steep, south dipping active normal fault. Some parts of this fault have slipped in recent times, recorded by instrumental records and historical earthquakes [e.g. the 1899 Nazilli–Denizli Earthquake, the 1956 Söke-Balat Earthquake and the 1965 Denizli Earthquake; see Ambraseys (1988), Westaway (1993) and Ambraseys & Jackson (1998 and refs cited therein). In other parts, the fluvial depocentre is bounded to the south by less important antithetic normal faults. For most of its length, the uplifting footwall of this active normal fault on the north side of the graben floor contains a narrow (*c.* 5–10 km) former depocentre which is now eroding. This depocentre is bounded to the north, at the southern edge of the outcrop of Menderes Massif metamorphic rocks, by a straight mountain front controlled by another segmented south dipping normal fault and its depositional substrate accumulated in the hanging wall when this fault was active. This contact is known to be a low-angle (at present) normal fault, as in places it is possible to measure its slip sense (e.g. Westaway 1990*a*, *b*; this study). However, there is no evidence (e.g. from seismicity) that it is active at present. It has long been suggested that neotectonic extension began on this more northerly fault zone (e.g. Jackson & McKenzie 1988; Seyitoğlu & Scott 1991, 1992)

In addition to the axial fluvial sedimentation, many small lateral rivers cut through the uplifted basin on the northern flank of the Büyük Menderes Graben. These have caused erosion of the Menderes Massif and uplifted Neogene basin, and deposition of alluvial fans on the valley floor where they are interbedded with the axial fluvial sediments. As the same pattern is evident within much of the sequence of eroding sediments of the uplifted western Anatolian Neogene basins (e.g. Roberts 1988; Paton 1992; Cohen *et al.* 1995), it is assumed that the same sedimentary and geomorphological environment existed at the time when these latter basins were infilled.

Sedimentary sequence

Fluvio–lacustrine sediments in and around the Büyük Menderes Graben are best exposed in a 2–5 km wide zone along its northern margin. These sediments are exhumed along the footwall of the south facing active normal faults with respect to the present-day graben floor. Three main lithological associations, based on their distinct structure, have been mapped in the Aydın area: (1) northwards tilted sediments (unit A); (2) almost flat-lying, terraced sediments (unit B); and (3) marginal alluvial fans and present-day graben-floor sediments (unit C; see Figs 3 and 4). Each unit contains vertical and lateral variations and displays various relationships of interfingering and intergradations.

Unit A. This unit consists mainly of northwards tilted continental clastic sediments located between the metamorphic rocks of the Menderes Massif in the north and the present-day graben-bounding faults in the south (Fig. 4). The basal lithology is a reddish, coarse-grained, well-cemented, poorly sorted, polygenetic conglomerate composed of clasts derived from the underlying metamorphics and minor but widespread interbedded lignites. Above the conglomerates, the unit is composed of siltstone, mudstone and shale alternations, together with conglomerates and pebbly sandstones. Lateral and vertical transitions from one lithology to another are very common throughout this sequence, which is also characterized by numerous scour-and-fill structures filled with channel

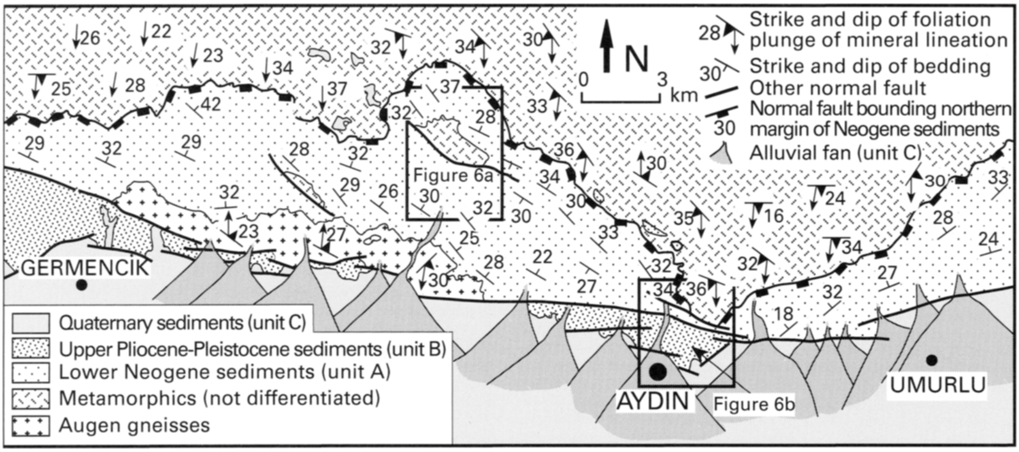

Fig. 4. Simplified geological map of the northern margin of the Büyük Menderes Graben in the area between Germencik and Umurlu.

conglomerates. This unit comprises a broadly coarsening-upwards sequence with a total thickness of *c.* 2 km (Cohen *et al.* 1995). Laminar to trough-like cross-bedding, pebble imbrication, graded bedding and normal-type growth faults are commonly observed synsedimentary structures in this unit. More details are given in Cohen *et al.* (1995).

Unit B. This unit comprises approximately horizontal, massive, cobble to boulder conglomerates with alternations of sandstone, siltstone, mudstone and claystone which crop out to the south of the tilted sediments of unit A. Unit B is bound by approximately east–west trending, high-angle normal faults along the contacts, both with the deformed sediments of unit A to the north and the younger basin-fill sediments (unit C) to the south (Fig. 4).

Unit C. These sediments, with the present-day configuration of the Büyük Menderes Graben, are juxtaposed with unit B sediments along high-angle graben-bounding normal faults. They are composed mainly of marginal alluvial fan and graben-floor sediments. The northern margin of the Büyük Menderes Graben is marked by many steep, well-developed, alluvial fans of diverse size, aligned in a narrow zone (Fig. 4). The source of the alluvial fan sediments is the metamorphic basement and exhumed unit A and B sediments. The alluvial fans grade into fine-grained basin-floor sediments along the Büyük Menderes River. In places, the alluvial fans coalesce and degrade and result in a fault-parallel alluvial fan apron (Fig. 4). The coarse-grained nature of the marginal sediments and the steepness of the alluvial fans indicate rapid uplift of the source mountains, accompanied by erosion and rapid sedimentation, attesting to the activity of these graben-bounding faults.

Structure

Three types of major structures occur along the northern margin of the Büyük Menderes graben: (1) an inactive, presently low-angle, normal fault; (2) west-northwest–east-southeast to northwest–southeast folds within the unit A sediments; and (3) approximately east–west high-angle, graben-bounding normal faults.

Table 1. *Measurements of slickensides and slickenlines on the presently low-angle normal fault*

Location	Dip direction (°N)	Dip amount (°)	Rake (°)	Sense
1	192	22	86	Normal
2	192	30	85	Normal
3	220	36	72	Normal
4	200	28	87	Normal
5	202	32	82	Normal
6	190	26	80	Normal
7	192	28	76	Normal
8	194	29	88	Normal
9	192	33	75	Normal
10	195	30	78	Normal
11	204	32	76	Normal
12	206	34	85	Normal
13	194	29	86	Normal

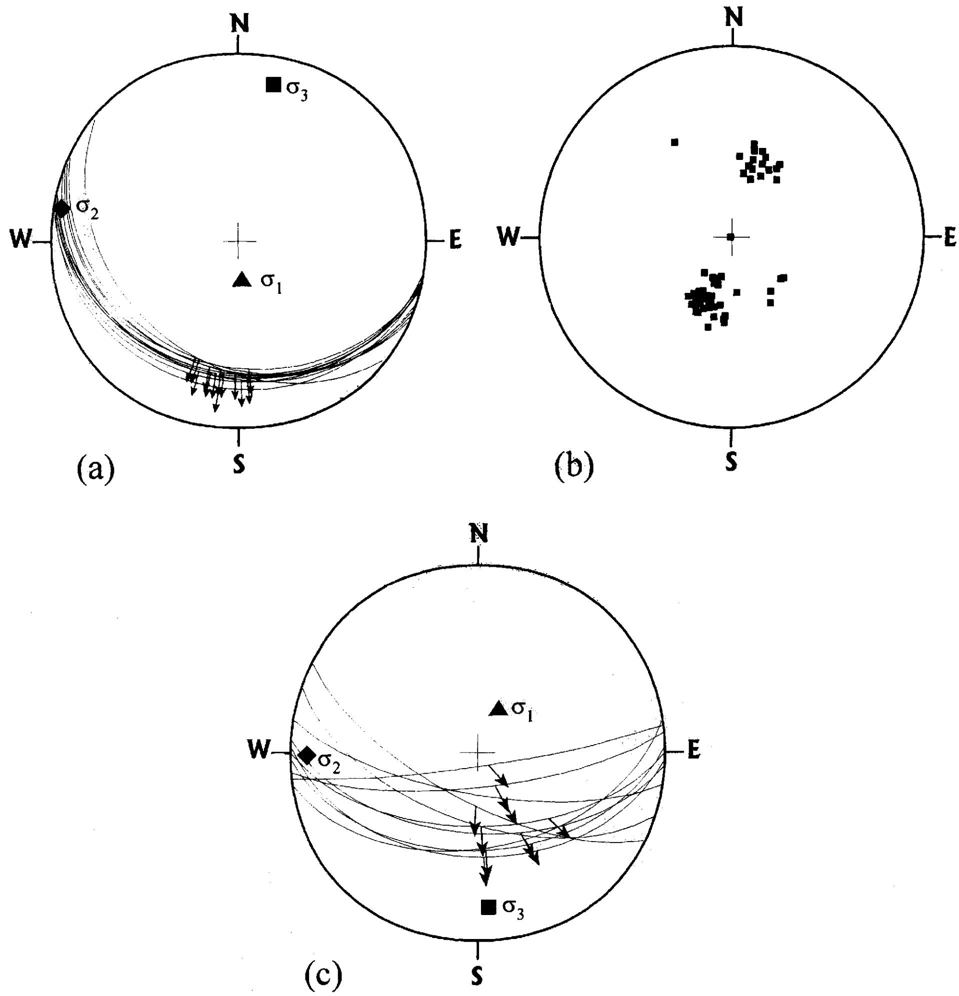

Fig. 5. Schmidt lower hemisphere equal-area projections of: (**a**) presently low-angle fault; (**b**), poles to bedding planes in Lower–Middle Miocene sediments; (**c**) graben-bounding, high-angle active normal faults. Great circles in (a) and (c) are fault surfaces, the arrows are striations (see Tables 1 and 2 for details).

The northern boundary of the unit A sediments is a major south facing, low-angle (22–34°; see Table 1; Fig. 5a), inactive, normal fault that separates them in the hanging wall from ductilely deformed metasediments and in the footwall from metagranite of the Menderes Massif to the north (Fig. 4). This fault has been cut by steeper graben-bounding active normal faults (as discussed below) and the metamorphics have been progressively uplifted, mylonitized and exhumed in the footwall. The deformed unit A sediments may thus be regarded as being deposited in a basin that was situated on the upper plate of this fault. Another, but circumstantial, piece of evidence of contemporaneous sedimentation and faulting is that the unit A sediments dip to the north, suggesting rotation of both the fault and the strata during the evolution of this fault. The calculated σ_1 trend, from stratum on this presently low-angle fault plane, is 163° and plunges steeply at 71°, whereas σ_2 and σ_2 axes plunge at 5 and 15°, respectively (Fig. 5a). These estimates of stress field orientations and others elsewhere in this study are calculated from observed slip vector orientations using the computer program of Caputo (1989).

Dips of beds within the unit A sediments vary throughout the basin. The available data (Fig. 6a) are interpreted as evidence of folding with

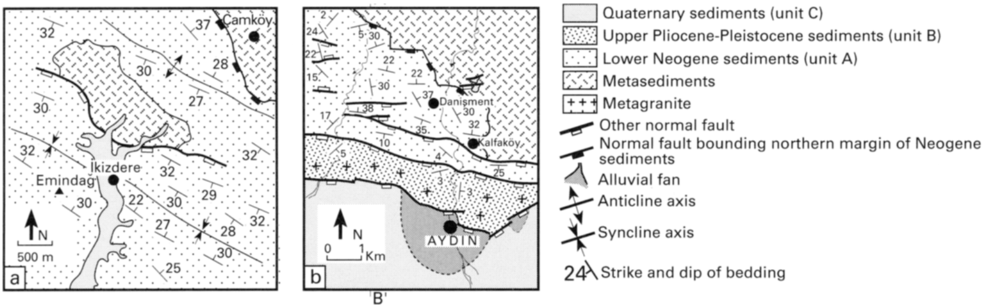

Fig. 6. Geological maps of (**a**) İkizdere and (**b**) Aydın areas [simplified and interpreted from Cohen *et al.* (1995)] showing the folds in unit A sediments and their boundary relationships with the Upper Pliocene–Pleistocene fluvial sediments (unit B).

Table 2. *Measurements of slickensides and slickenlines on the high-angle graben-bounding faults*

Location	Dip direction (°N)	Dip amount (°)	Rake (°)	Sense
1	208N	69	80	Normal
2	200N	60	78	Normal
3	180N	58	60	Normal
4	190N	72	76	Normal
5	172N	84	86	Normal
6	170N	48	77	Normal
7	174N	75	84	Normal
8	182N	44	86	Normal
9	178N	47	88	Normal
10	184N	55	70	Normal

west-northwest–east-southeast axes. This folding can be seen directly in the field at the locations covered in Fig. 6a, where dips of beds change direction systematically over scales of typically several hundred metres. This folding thus occurs on a much larger scale than the minor folding noted by Cohen *et al.* (1995) and causes lateral variations in dip on a scale of a few metres. These folds are open structures with vertical to inclined axial planes and gently plunging axes that run parallel to the graben-bounding normal faults (Figs 4, 5b and 6a). The structures are observed to fold the bedding planes of the unit A sediments. The dip of beds averages 30° but in areas close to the inactive normal fault this may increase to 35–40° (Fig. 5b).

Although the unit A sediments are northward tilted and locally folded, the unit B strata in the basin show a different evolution. They are deformed only by graben-bounding normal faults, which are the most conspicuous features of the northern margin of the Büyük Menderes Graben. These faults, which dip southwards (Fig. 5c; Table 2), form the boundary between the deformed unit A sediments and the approximately horizontal unit B and younger basin fill (unit C; see Figs 4 and 6). These faults dip southwards at angles of 44–84° (Table 2) and show normal faulting with minor components of left-lateral slip (Fig. 5c). Computed results of slip data measurements on these fault planes define an approximately vertical σ_1 trending 25° and plunging steeply at 75°, and σ_2 and σ_3 dipping gently at 10 and 15°, respectively (Fig. 5c). The unit A sediments have been exhumed along the footwall of these active structures and provide the source of both the terraced unit B and younger basin-fill sediments (unit C). As these faults control rapid changes in the morphology and the drainage pattern, and are marked by triangular facets, fault scarps and active and extensive development of steep alluvial fans, they may have a neotectonic origin. The activity of these structures is indicated by the recent earthquakes that have occurred along them (see Established knowledge).

Interpretation

Sedimentary unit A

The relatively steeply north dipping fluvial and fluvial fan sediments with red weathering, which are situated at the base of the young sedimentary sequence in the uplifted basin north of the Büyük Menderes Graben near Aydın, are called unit A. This unit can be correlated with unit I of Cohen *et al.* (1995; Fig. 3). Cohen *et al.* (1995) tentatively accepted these sediments as equivalent to the lignite-bearing sediments, also with red weathering, from near Nazilli (Fig. 2). Seyitoğlu & Scott (1992) assigned to these red

Table 3. *The results of mammal sites dated by Ünay* et al. *(1995) and Ünay & De Brujin (1998)*

Location	Name	Situation	Mammal age
1	Söke	BM valley floor	(a) Early Biharian (Early Pleistocene) (b) Toringian (Middle–Late Pleistocene)
2	Ortaklar	Uplifted basin outside BMFZ	Late Villanian (Late Pliocene)
3		BM valley floor	Late Pliocene–Early Pleistocene
4	Germencik	BM valley floor	Late Biharian–Early Toringian (Early–Middle Pleistocene)
5		BM valley floor	Late Pliocene–Pleistocene
6	Kurttepe	Uplifted basin N of BMFZ	Late Villanian–Early Biharian (Late Pliocene–Early Pleistocene)
7	Bozköy	Uplifted basin N of BMFZ	Late Villanian (Late Pliocene)
8	Nazilli–Şevketin Dağ	Uplifted basin N of BMFZ	Late Villanian–Early Biharian (Late Pliocene–Early Pleistocene)

sediments the Eskihisar sporomorph assemblage which is dated Early–Middle Miocene (20–14 Ma; Benda & Meulenkamp 1979). Because these are the oldest Neogene sediments in the Nazilli area, Seyitoğlu & Scott (1992) inferred that they mark the start of neotectonic extension in the Büyük Menderes Graben. Similarly, Lower–Middle Miocene coal-bearing sediments were reported from different parts of the Büyük Menderes Graben to the east of the present study area in some earlier studies (Karamanderesi 1972; Emre & Sözbilir 1995). There are, of course, red Neogene sediments in a lot of other places in western Turkey, including sites outside extensional basins (e.g. Becker-Platen 1971; Sickenberg & Tobien 1971; Kaya 1981; Gökçen 1982; Steininger & Rögl 1984). However, it is not obvious that all of these sediments are the same age. Most of the correlations in the previous works were based on red weathering. This suggests only that at some time since the youngest of these sediments were deposited the climate was subtropical for a while, and thus oxidized whatever sediments happened to be already exposed. It is known from the literature that climates favourable to hematite genesis in western Turkey persisted until the early part of the Late Miocene, or possibly even later (e.g. Steininger & Rögl 1984; Robertson *et al.* 1991). Thus, there is no convincing evidence that the red fluvial sediments near Aydın have the same age as the red lignite-bearing sediments near Nazilli. The argument related to Neogene climate change places a lower bound to the age of the sediments at Aydın.

More recently, a mammal site from the previously mapped Lower–Middle Miocene sediments (Seyitoğlu & Scott 1991) in the Nazilli area (Şevketin Dağ: Table 3; Fig. 7, location 8) yielded Late Pliocene–Early Pleistocene ages (Late Villanian–Early Biharian) from these (approximately horizontal) sediments (Ünay *et al.* 1995; Ünay & De Brujin 1998). This means that the Miocene ages for the sediments quoted by Seyitoğlu & Scott (1992) are no longer tenable and revision of their stratigraphy and its interpretation are urgently required. Moreover, another mammal site (Bozköy: Table 3; Fig. 7, location 7) from the northward tilted red clastics, designated as unit A sediments in the present study, yielded a Late Pliocene (Late Villanian) age (Ünay *et al.* 1995; Ünay & De Brujin 1998). However, Ünay *et al.* (1995) and Ünay & De Brujin (1998) reported neither the stratigraphy at these locations nor the positions of the dated sites in the succession. Pending the necessary revision of the Seyitoğlu & Scott (1991) stratigraphy and taking account of the other points already mentioned, the age of the unit A sediments will be quoted in this paper as Early Neogene (?Early-Middle Miocene or even as young as ?Pliocene).

As noted by Cohen *et al.* (1995), the presence of abundant small-scale normal faults are another characteristic feature in the unit A sediments. The cut-off angle with the bedding suggests that these structures formed as high-angle faults (70° to the vertical) but rotated to lower angles during northward back-tilting of the unit A sediments. My own observations strongly suggest a syndepositional origin for these structures, confirming the conclusion of Cohen *et al.* (1995). Evidence for syndepositional fault activity includes: (1) abrupt and rapid termination of fault displacements upwards in the stratigraphy; (2) thickness variations in the lithologies across the faults where

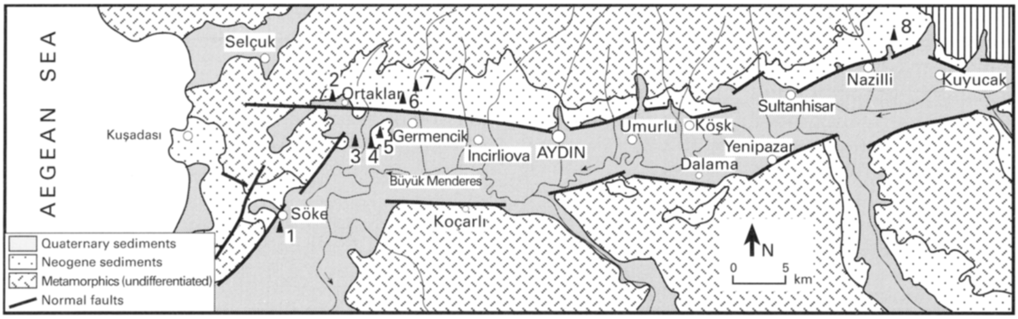

Fig. 7. Simplified map of the Büyük Menderes Graben showing mammal sites dated by Ünay *et al.* (1995) and Ünay & De Brujin (1998). See Table 3 for details.

the sediments (usually, relatively coarse grained) are thicker in their hanging walls than in their footwalls; and (3) wedging of these hanging-wall sediments, which thin away from the fault.

Sedimentary unit B

This unit can be correlated with units II–IV of Cohen *et al.* (1995) (Fig. 3). It is not weathered red and is thus younger than the time of any climate which allowed that style of weathering to happen. Evidently, an axial river existed when it was deposited, so it presumably post-dates the start of neotectonic extension within the Büyük Menderes Graben. On the other hand, it is not back-tilted nor does the bedding diverge as would be expected if it thickened towards a normal fault. This can be interpreted in two ways: (1) unit B was deposited over a sloping palaeoland surface and so it did not thicken towards any active fault and does not indicate the palaeohorizontal; (2) there was an initial phase of extension on the more northerly fault bordering the Menderes Massif metamorphics which tilted unit A, followed by a pause during which unit B was deposited; finally, extension resumed. The second interpretation is preferred here.

The clear structural difference between the unit B sediments and the older Neogene fluvial sediments of unit A (Cohen *et al.* 1995; this study) suggest a structural discontinuity between them. This, in turn, indicates either a substantial time gap while the older unit was eroded (see Discussion) or a sudden erosional event, in which case the most probable cause is a reduction in the base level of the river which drained this area at the time. One possibility is that this change relates to the start of cyclic drawdown in global sea level *c.* 2.5–2 Ma, caused by the first development of northern hemisphere ice sheets. Another possibility is that it reflects the Messinian drawdown in the level of the Mediterranean at the end of the Miocene (e.g. Robertson *et al.* 1991 and refs cited therein). Many of the existing palaeogeographic maps show a land bridge in the way, such that this region either drained internally or northwards into the Paratethys [i.e. Black Sea; see e.g. Robertson *et al.* (1991 and refs cited therein)]. If so, changes in its base level could have responded to climate-induced changes in the level of the Black Sea. The angular unconformity between Miocene and Pliocene, and/or younger sediments, has long been known and was previously reported from the area to the east of Aydın (Karamanderesi 1972).

Until recently, no diagnostic fossil evidence existed from the post-Miocene sediments of the Büyük Menderes Graben. Ünay *et al.* (1995) and Ünay & De Brujin (1998) presented such evidence from eight sites within the graben and assigned, on the basis of mammal faunas, a Late Pliocene–Pleistocene age to these fluvial sediments. The dated sites, except for location 7, lie outside the study area, but their results are summarized in Table 3 and the locations are given in Fig. 7. The important note to be added here is that the dated samples (except for location 7) are from horizontal sediments, which are designated as unit B in this study. The obvious interpretation is that sedimentation within the present fluvial depocentre has been continuous since at least the Late Pleistocene, but sedimentation in what is now the uplifted basin north of this modern depocentre had ceased by the late Early Pleistocene or Middle Pleistocene. In other words, this fossil evidence supports the view that the present set of normal faults bounding the modern depocentre became active around the end of the Early Pleistocene, i.e. *c.* 1 Ma.

This interpretation is consistent with that of

Jones & Westaway (1991) who made the first tentative estimate of *c*.1 Ma for the timing of transition to the modern set of faults. Subsequently, similar timings have been proposed for other Aegean normal faults, notably in the Gulf of Corinth where timing is constrained by well-dated marine sediments (Westaway 1996). Westaway (1994*a*, *b*) first suggested that this timing may be the same throughout the region. Later, Westaway (1996, 1998) proposed a possible physical mechanism.

Tilting and folding of Unit A

Unit A is tilted northwards which is thought to be the result of back-tilting beside a set of south dipping normal faults, as suggested by Cohen *et al.* (1995). This further means that extension was occurring on a fault system which approximates to the fault that bounds the oldest Neogene sediments (in its hanging wall, with the Menderes Massif metamorphics in its footwall) sometime after and/or during deposition of unit A but before unit B was deposited. This fault was later cut and locked up when slip began on the now active fault zone along the edge of the fluvial depocentre. This change occurred for some reason connected with the observation that slip on the initial fault zone back-tilted it and thus changed its orientation so that slip could no longer be maintained in the regional stress field (e.g. Jackson & McKenzie 1988). Similar abandoned young depocentres are evident in the footwalls of other active normal fault zones in western Turkey and central Greece (e.g. Roberts & Jackson 1991; Westaway 1998), suggesting that a systematic effect affected fault systems throughout this region. However, in some localities in the Büyük Menderes Graben, such as around Aydın, there are three generations of faults instead (Fig. 6b), indicating a more complex pattern.

The fluvial sediments of unit A are also observed to be folded (Figs 4 and 6). One particular area has been chosen in which to study these structures (Fig. 6a) and it is described here for the first time in the literature. It is a particularly good locality to study because access is relatively easy.

The age of this folding is uncertain. Because of this uncertainty in timing it is also not clear whether the folding occurred synchronously with extension during the deposition of unit A sediments or during a short time interval following the deposition of unit A. There are, of course, plenty of possible mechanisms for folding during extension due to: (1) differential compaction; (2) draping; (3) fault 'drag'; and (4) lateral variations in tilt caused by individual fault segments dying out along-strike. However, the scale of the folding and the lack of any clear relationship to the normal faults do not indicate a synsedimentary cause (see below).

Before going further, it is important to decide whether folds are local, i.e. unique to the Büyük Menderes Graben, or regional. Similar structures within the Neogene sediments have long been known. There are reports from many Aegean islands, particularly those located close to the coast of western Turkey (e.g. Kos, Samos, Chios, Paros, Naxos, Mykonos, Anafi and Milos) (Angelier 1976, 1978; Angelier & Tsoflias 1976; Mercier 1976, 1979, 1981; Mercier *et al.* 1976, 1979; Jackson *et al.* 1982; Boronkay & Doutsos 1994) and many of the western Anatolian grabens. In most of these studies it has been emphasized that extension in the Aegean was interrupted, at least in some places, by one or more shorter periods of compression involving folding and/or thrusting. In contrast, Jackson *et al.* (1982) propose that these shortening structures can be satisfactorily explained by uplift in the footwall blocks of normal faults and do not require regional compression. They also suggested that the compressional episodes are probably not regional in extent and may not be truly compressional in origin, but are more likely to be a consequence of considerable rotation due to internal deformation of blocks bounded by major normal faults. However, more recent studies, particularly that of Boronkay & Doutsos (1994), report evidence from the central Aegean region which suggests that crustal shortening occurred during the Miocene and that the resulting transpressive structures controlled the evolution of sedimentary basins.

Moreover, personal field observations near the eastern end of the Büyük Menderes Graben to the south of Buldan (Fig. 2), and in other east–west grabens, and the integration of available literature (e.g. Nebert 1960, 1978; Ercan *et al.* 1978; Dumont *et al.* 1979; Boray *et al.* 1985; Yalçın *et al.* 1985; İnci 1991; Yağmurlu 1991; Koçyiğit *et al.* 1995, 1999; Bozkuş 1996; Seyitoğlu 1997; Yılmaz 1997; Altunkaynak & Yılmaz 1998; Koçyiğit & Bozkurt 1998; Yılmaz *et al.* 2000) confirm that the Miocene deposits in many of the western Anatolian basins (regardless of their size and orientation) are deformed and folded, strongly suggesting a regional event.

This information favours the second possibility that folding occurred during a time interval after deposition of unit A ceased but before the deposition of unit B sediments began. This event is constrained between the age of deformed Lower Neogene sediments (unit A) and unconformable Pliocene–Pleistocene sediments (unit B). Given the available information (already discussed), on the timing of unit A deposition, this folding event can be dated sometime after the Middle Miocene but before the Late Pliocene. It is noteworthy that this time interval corresponds to a major break in sedimentation and magmatism, and a regional folding event, across many of the western Anatolian basins (see Discussion).

Nevertheless, whatever the cause of folding in unit A sediments, the important point is that unit B sediments do not bear any sign of deformation. The obvious interpretation is that the Lower Neogene sediments, exhumed on the shoulders of present-day Büyük Menderes Graben, have nothing to do with the age of initiation of neotectonic extension in the graben as was previously thought by Seyitoğlu & Scott (1992). Instead, it is the younger Pliocene–Pleistocene undeformed sediments which are coeval with formation of modern Büyük Menderes Graben.

This interpretation thus indicates two distinct phases of extension. The first phase of extension involved slip on the presently low-angle normal fault bounding the northern edge of the Lower Neogene sediments (Fig. 4). This extension appears to have accompanied the deposition of unit A sediments. It was followed by an interval during which unit A sediments were folded. Later still, extension resumed and led to the modern geometry of the Büyük Menderes Graben.

Initial dip of the low-angle normal fault

As already mentioned, the northern boundary of unit A sediments is a major normal fault, with a present-day dip of 22–34° (Fig. 5a). The present-day dips can be related to the dips of the steepest dipping Lower Neogene sediments, which reach 30–35°. The most appropriate way to restore such dips is [following Westaway & Kusznir (1993a)] to assume that the rocks deformed during extension by distributed vertical shear. If α and β are initial and present-day dips of the fault, respectively, and δ is the present-day dip of the oldest hanging-wall sediments, then:

$$\tan \alpha = \tan \beta + \tan \delta \qquad (1)$$

With β = 22—34° and δ = 30–35°, the initial dip of the fault plane (α) can be calculated as 44–54°, i.e. 49 ± 5°. Dips of this order are common for many faults in the Aegean region (e.g. Westaway 1993; Westaway & Kusznir 1993*a*, *b*) and are explained by conventional theory. Thus, this particular boundary fault is a normal fault with an expected initial dip which has been back-tilted as a result of substantial extension. This agrees very closely with the estimates by Cohen *et al.* (1995) for the Büyük Menderes Graben and those given by Westaway (1993) and Westaway & Kusznir (1993*a*, *b*) for the initial dip of Denizli Normal Fault (Fig. 2).

Age of the Büyük Menderes Graben

The above observations demonstrate that unit A sediments, since the earliest Neogene ones along the northern margin of the Büyük Menderes Graben are back-tilted northwards and folded, cannot correspond to the early graben fill as previously suggested by Seyitoğlu & Scott (1992). Moreover, the clear angular difference between the tilted beds of Lower Neogene and horizontal Pliocene–Pleistocene sediments implies the presence of a major regional unconformity. However, they were previously considered to form a single continuous megasequence (Seyitoğlu & Scott 1992). The early sediments of the neotectonic graben must therefore correspond to the horizontal terrace sediments exhumed in the footwall of the graben-bounding normal faults (unit B), thus indicating a Pliocene or younger age for the initiation and formation of the present-day graben.

The important erosional surface that developed on the Lower Neogene sedimentary rocks (unit A) is not unique to the Büyük Menderes Graben but also occurs in many of the east–west trending grabens in western Turkey, such as the Gediz Graben (Yağmurlu 1987; Cohen *et al.* 1995; Koçyiğit *et al.* 1999), the Gökova Graben (Görür *et al.* 1995) and the Kütahya Graben (Koçyiğit & Bozkurt 1998). In all of these cases, this surface is overlain unconformably by Pliocene fluvial conglomerates. There are, of course, many places in Turkey where sediments such as those at the base of the Büyük Menderes sequence are found in sag basins or ovas (plains). In principle, it seems reasonable to consider the possibility that, in the Büyük Menderes Graben, the situation may be one where young normal faults have cut an older sag basin and the early sedimentation was previously misinterpreted.

Another way of attempting to estimate the age of Büyük Menderes Graben is to divide the extension across it (i.e. the sum of heaves of the graben-bounding normal faults) by the extension rate. The total heave (= horizontal slip) across the graben-bounding normal faults are measured, as accurately as possible, on a structural cross-section based on fig. 11b of Cohen

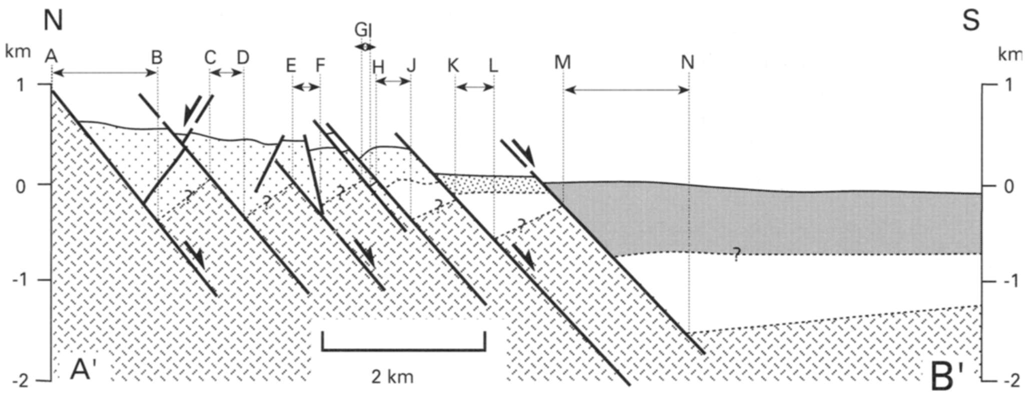

Fig. 8. Geological cross-section of the northern margin of the Büyük Menderes Graben (see Fig. 6b for location) based on fig. 11b of Cohen *et al.* (1995). This cross-section indicates a total of *c.* 5 km of extension. Assuming a uniform extension rate, the age of the fault zone is (*c.* 5 km/1 mm a^{-1}) 5 Ma. One could partition this with a possible *c.* 3 km of extension on the first set of faults, during the *c.* 5–2 Ma interval, then *c.* 1 km on the second fault set during the *c.* 2–1 Ma interval, then another *c.* 1 km on the present set since *c.* 1 Ma. A–B, 1.32 km; C–D, 0.4 km; E–F, 0.35 km; G–H, 0.3 km; I–J, 0.45 km; K–L, 0.5 km; M–N, 1.62 km.

et al. (1995) (Fig. 8). Westaway (1994*a*, *b*) argued that a reasonable present-day extension rate is *c.* 1 mm a^{-1}. This cross-section indicates a total of *c.* 5 km of extension. Assuming a uniform extension rate, the age of the fault zone is c. 5 km/1 mm a^{-1}, or 5 Ma (see Fig. 8 caption).

It was pointed out earlier that the deposition of unit A and other Lower–Middle Miocene sediments exposed along the northern margin of the Büyük Menderes Graben may have accompanied an early phase of extension along the presently low-angle normal fault at the northern margin of the depocentre. Furthermore, it is not clear how much of the tilting of these sediments, and the slip, occurred during such an earlier phase and how much occurred later. The same normal fault surface, active in the first phase of extension, may have been reactivated during the early part of the second phase of extension. This is quite logical since it is already known that new structures commonly follow pre-existing planes of weakness.

The Pliocene initiation age for the Büyük Menderes Graben is in close agreement with those suggested for the Gediz Graben (Early Pliocene: Koçyiğit *et al.* 1999), for the Gökova Graben (latest Miocene–Pliocene: Kurt *et al.* 1999) and for the whole of western Turkey (Pliocene: Yılmaz *et al.* 2000). Furthermore, Burchfiel *et al.* (2000) confirm that the initiation of east–west trending grabens (that mark the northern boundary of Aegean graben system) in central Bulgaria is no older than 9 Ma (perhaps no older than *c.* 6.5 Ma).

Discussion

Seyitoğlu & Scott (1991, 1992) proposed that the initiation of Büyük Menderes Graben, and therefore the neotectonic north–south extensional tectonics in western Anatolia, occurred in the Early Miocene. They thus suggested that this extension involved the spreading and thinning of crust which had previously thickened as a result of the Late Palaeogene continental collision following closure of the Neotethys Ocean. A Pliocene inception age for the graben, proposed here, clearly contradicts these previous conclusions. It is worth mentioning that Seyitoğlu & Scott (1992) provided no evidence that the lignite-bearing red clastics were deposited during extension. The only basis for their model is the assumption that Neogene sediments record the initiation age of the graben [following previous contentions by Şengör & Yılmaz (1981), Şengör *et al.* (1985) and Şengör (1987)] and the reassessment of the sediment age using the newly proposed age span of the Eskihisar sporomorph association (Benda & Meulenkamp 1979) contained within the lignite layers.

In contrast, other lines of evidence suggest that the age of the present-day extension prevailing in western Anatolia is Pliocene and is therefore unlikely to be the consequence of orogenic collapse.

- Evidence from the Niğde Massif in central Anatolia (Whitney & Dilek 1997, 1998) suggests that Late Oligocene–Early Miocene extensional collapse and core-complex formation is not unique to western Anatolia but is widespread and affects larger areas, including central Anatolia. Similarly, Early–Middle Miocene extension has also been postulated for the Çankırı Basin in central Turkey (Kaymakçı *et al.* 2000).

- The presence of Early Miocene normal faults and associated sedimentary basins in their hanging wall is not limited to Turkey. Similar zones are recognized in the Cycladic Massif (e.g. Lister *et al.* 1984; Urai *et al.* 1990; Faure *et al.* 1991; Lee & Lister 1992; Gautier *et al.* 1993; Gautier & Brun 1994; Vandenberg & Lister 1996) and in the Rhodope Massif (e.g. Dinter & Royden 1993; Tzankov *et al.* 1996; Dinter 1998; Burchfiel *et al.* 2000). Therefore, it can be proposed that extensional deformation driven by gravitational collapse was widespread, affecting larger areas including the central Aegean, western Turkey and as far east as central Anatolia. In contrast, north–south neotectonic extension in Turkey is limited to western Anatolia (Fig. 1).
- Using recent GPS measurements, Barka & Reilinger (1997) and Reilinger *et al.* (1997) further demonstrated that central Anatolia [the Ova Province of Şengör *et al.* (1985)], previously affected by orogenic collapse-accommodated extension, is now undergoing an approximately north–south or north-northeast—south-southwest shortening and anticlockwise rotation due to slip along the dextral North Anatolian Fault.
- Although Miocene continental sediments are widespread in western Anatolia, within both the north–south and east–west grabens around and within the Menderes Massif, the Late Miocene–Pliocene sediments are confined to the east–west grabens (e.g. Paton 1992; Görür *et al.* 1995; Yılmaz 1997; Yusufoğlu 1998; Koçyiğit *et al.* 1999; Yılmaz *et al.* 2000), supporting the view that the east–west grabens developed during the Pliocene.
- It has been concluded, based on palaeomagnetic data and extrapolation of modern strain rates, that most of the total extension in the Aegean has occurred since the Early Pliocene (Jackson & McKenzie 1988; Kissel & Laj 1988).
- Although the Miocene sediments in western Turkey are usually deformed (as mentioned above), the Pliocene and younger sediments show no sign of such a style of deformation.
- Lastly, movements dated to 6–7 Ma (from $^{40}Ar–^{39}Ar$ laser probe experiments on white mica; Lips 1998) have been documented along the southern margin of the Gediz Graben. This implies that a phase of Early Miocene extension (Hetzel *et al.* 1995) was followed by latest Miocene extension at this locality, both slips being accommodated on the same fault (as is proposed in this study for the Büyük Menderes Graben).

This evidence suggests that the neotectonic phase of north–south extension in western Anatolia commenced during the latest Miocene or Pliocene. This age is in agreement with the inception of dextral movement along the North Anatolian Fault Zone (sometime in the Pliocene; Tokay 1973; Barka 1984, 1997; Barka & Kadinsky-Cade 1988; Koçyiğit & Rojay 1988; Toprak 1988; Dirik 1991; Rojay 1993; Tatar 1993; Westaway 1994*a*; Bozkurt & Koçyiğit 1995, 1996; Westaway & Arger 1996, 1998; Ünay *et al.* 1998; Yürür *et al.* 1998). This phase of extension can thus be attributed to a 'tectonic escape' mechanism. It is either a direct consequence of slip on the North Anatolian Fault Zone or an indirect consequence caused by another change – e.g. a change in the geometry of subduction along the Hellenic Trench – caused by slip on the North Anatolian Fault Zone.

It is thus suggested here that western Anatolia may be an example of a region which has experienced two modes of extension: a 'core-complex mode' and a 'wide-rift mode' (cf. Buck 1991). The first phase may be related to the latest Oligocene–Early Miocene gravitational collapse of the orogenically thickened crust and core-complex formation, following the Palaeogene collision across the Neotethyan ocean (Seyitoğlu & Scott 1991, 1992). This event is considered to have commenced *c.* 18–20 Ma [for details see Seyitoğlu & Scott (1992), Seyitoğlu *et al.* (1992) and Hetzel *et al.* (1995)]. It has been suggested that, during this collapse, the Menderes Massif was exhumed in the lower plate of presently low-angle normal fault(s) (Bozkurt & Park 1994, 1997; Hetzel *et al.* 1995, 1998). Deposition of the oldest sediments, which now crop out along the Büyük Menderes Valley, may have occurred in the hanging wall of such a fault during the 14–20 Ma time interval (Seyitoğlu & Scott 1991, 1992; Emre & Sözbilir 1995; Emre 1996; Seyitoğlu 1997). Collapse and related extension ceased by the mid-Late Miocene (*c.* 12 Ma) according to $^{40}Ar–^{39}Ar$ biotite cooling ages from the synextensional Salihli and Turgutlu Granodiorites emplaced along the footwall of the north facing, presently low-angle, normal fault at the southern margin of the Gediz Graben (Hetzel *et al.* 1995). This timing is also consistent with the age of the Miocene sediments (14–20 Ma; Seyitoğlu & Scott 1992; Seyitoğlu 1997).

A comparison of the present-day graben-fill sediments (Pliocene–Pleistocene) and radiometric constraints from the extensional shear zone mylonites (*c*. 12–20 Ma: Hetzel *et al.* 1995), and the hanging-wall basin fill (*c*. 14–20 Ma; Seyitoğlu & Scott 1991, 1992; Seyitoğlu 1997), suggest a time gap lasting *c*. 6–8 Ma, separating the development of major normal faults and the associated metamorphic complexes from the development of the horst-and-graben systems currently observed in western Anatolia. The latter interval may reflect a change in style of extension or correspond to a time of regional erosion that created the major disconformity between unit A and B sediments, and the folding of Miocene sediments, in the Büyük Menderes Graben and in other Miocene basins in western Turkey. This time interval also corresponds to a major break in magmatism in western Turkey (*c*. 14–10 Ma; Yılmaz *et al.* 2000) between the calc-alkaline, high-K Oligocene–Early Miocene first stage which died out at *c*. 14 Ma (Yılmaz 1989, 1990, 1997; Altunkaynak & Yılmaz 1999); and also to the second phase of Late Miocene–Pliocene rift-related basaltic volcanism. [Readers are referred to Yılmaz *et al.* (2000) for a detailed discussion.]

Following the initiation of strike-slip movement along the dextral North and sinistral East Anatolian Faults, the Anatolian block began to move westwards. The effect of westward tectonic escape resulted in a north–south upper crustal extension on active east–west normal faults during the Pliocene. The start of neotectonic extension is also marked by a change in the nature of the volcanism from dominantly calc-alkaline in the Middle Miocene to alkaline, including the Pliocene Kula basalts (7.5 $\pm$ 0.22–0.00025 Ma K–Ar ages; Ercan *et al.* 1985).

The above discussion further demonstrates that the east–west and north–south basins in western Turkey have not developed coevally, as was thought, for instance, by Şengör (1987) and Seyitoğlu (1997). In contrast, the north–south basins are cut by later east–west trending high-angle normal faults (see Yılmaz *et al.* 2000). This interpretation is further supported by the fact that the north–south grabens are elevated along the footwalls of the east–west normal faults. Thus, the episodic two-stage extension model suggested for the Gediz Graben by Koçyiğit *et al.* (1999) also seems to be supported in adjacent regions. Firstly, it explains why extension in western Anatolia and the central Aegean have occurred during and since the Pliocene (McKenzie 1978; Dewey & Şengör 1979; Koçyiğit *et al.* 1999). Secondly, it also accounts for the differences between the calculated initial dips of the low-angle normal faults controlling the core-complex mode of extension (*c*. 30° for the Gediz Graben; Hetzel *et al.* 1995: 49 $\pm$ 5° for the Büyük Menderes Graben; this study) and the high-angle graben-bounding faults which are active at present (up to 82°; Hancock & Barka 1987; Westaway 1990*a*, *b*, 1993; Jones & Westaway 1991; Paton 1992; Cohen *et al.* 1995; this study).

Conclusions

- There is no evidence that the red clastic sediments at the base of the Neogene sequence were deposited during neotectonic extension.
- The Büyük Menderes Graben exhibits evidence for two stages of extension – an initial extension on moderately steeply dipping normal faults was superseded by later steeper normal faults. This can support the following interpretation: during the first stage (latest Oligocene–Early Miocene), the Lower Neogene fluvial clastics were deposited in the basin that sits on the hanging wall of the normal fault(s), and the metamorphic rocks of the Menderes Massif were deformed, mylonitized and progressively exhumed in the footwall. The main graben-bounding normal faulting (Pliocene?) that overprints this early phase cuts and offsets the Miocene units and the normal faults.
- The fault that has controlled the early phase of extension may have been reactivated during the second phase.
- Mammal evidence constrains the start of slip on these younger faults to *c*. 1 Ma, as has been believed from other evidence.
- The Pliocene estimate for the start of extension is in close agreement with the start of slip on the North Anatolian Fault Zone, which further suggests that the extension can be explained as a geometrical consequence of this strike-slip movement. This further implies that the whole crust in western Turkey and the Aegean has since undergone extension due to the effect of extrusion processes along the North and East Anatolian Faults (tectonic escape).
- The two phases of deformation appear to reflect significant changes in the tectonic setting of western Anatolia, which can be attributed to orogenic collapse followed by tectonic escape.

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