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A pulsed extension model for the Neogene–Recent E–W-trending Alaşehir Graben and the NE–SW-trending Selendi and Gördes Basins, western Turkey

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Abstract

We have developed a significant body of new field-based evidence relating to the history of crustal extension in western Turkey. We establish that two of the NE–SW-trending basins in this region, the Gördes and Selendi Basins, whose sedimentary successions begin in the early Miocene, are unlikely to relate to late-stage Alpine compressional orogeny or to E–W extension of Tibetan-type grabens as previously suggested. We argue instead that these basins are the result of earlier (?) late Oligocene, low-angle normal faulting that created approximately N–S "scoop-shaped" depressions in which clastic to lacustrine and later tuffaceous sediments accumulated during early–mid-Miocene time, separated by elongate structural highs. These basins were later cut by E–W-trending (?) Plio–Quaternary normal faults that post-date accumulation of the Neogene deposits. In addition, we interpret the Alaşehir (Gediz) Graben in terms of two phases of extension, an early phase lasting from the early Miocene to the (?) late Miocene and a young Plio–Quaternary phase that is still active. Taking into account our inferred earlier phase of regional extension, we thus propose a new three-phase "pulsed extension" model for western Turkey. We relate the first two phases to "roll-back" of the south Aegean subduction zone and the third phase to the westward "tectonic escape" of Anatolia.

Keywords: Rifting; Graben; Neotectonics; Crustal extension; Western Turkey

1. Introduction

The grabens of western Turkey, in the eastern part of the Aegean extensional province (Fig. 1), notably the Alaşehir Graben (also known as the Gediz Graben), represent one of the classic areas for study of continental extension, comparable to the western USA (e.g., Crittenden et al., 1980; Coney, 1980; Hamilton, 1987; Davis and Lister, 1988; Lister and Davies, 1989; Malavielle, 1993; Miller and John, 1999). However, various controversies concerning the regional tectonic interpretation need to be resolved before this potential can be unlocked.

Three main processes have been proposed to explain the extension of western Turkey. The first is

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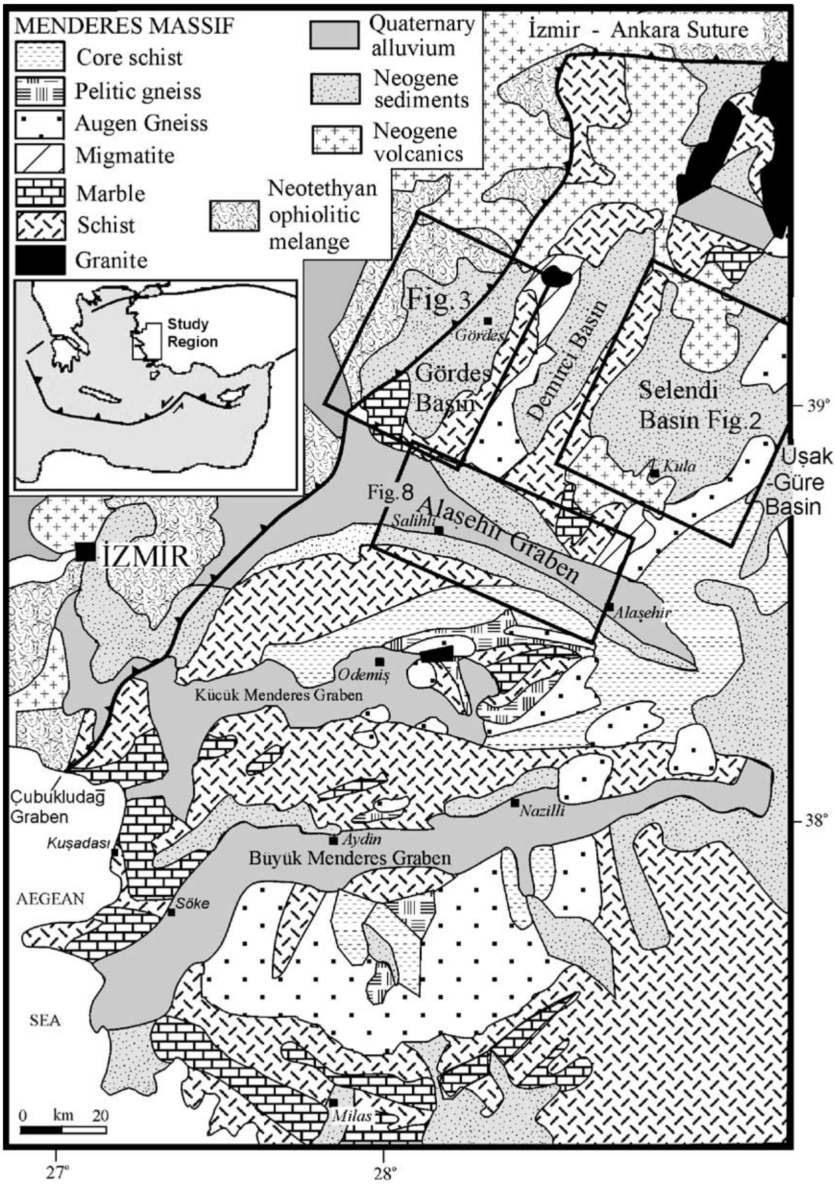

Fig. 1. Geological map of part of Aegean Turkey that was strongly affected by late Cenozoic extension. Note the positions of the Selendi and Gördes Basins and the Alaşehir Graben in relation to the metamorphic rocks of the underlying Menderes Massif (see Figs. 2, 3 and 7). Modified after Hetzel (1995).

orogenic collapse of thickened crust following suturing of the Neotethys Ocean. The second is back-arc rifting behind a Tethyan subduction zone to the south. The third is westward tectonic escape of Anatolia towards the Aegean region. In addition, there are several different interpretations of the timing and direction of extension and compression-related processes have also been proposed (see Bozkurt, 2003 for a recent review). The aim of this paper is to summarise evidence from the Gördes and Selendi Basins and the adjacent Alaşehir Graben (Fig. 1) and to use this to test alternative models for continental extension in this region and thus contribute to the debate concerning extensional processes in general.

2. The Gördes and Selendi Basins

The Gördes and Selendi Basins (Figs. 2 and 3) are the two most studied of the NE–SW-trending basins of western Turkey (e.g., Nebert, 1961; Yağmurlu, 1987; Ercan et al., 1978, 1983a,b; Seyitoğlu and Scott, 1991, 1994a,b; Seyitoğlu, 1992, 1997; Seyitoğlu et al., 1992; Yılmaz et al., 2000; Westaway et al., 2003, 2004). This literature provides a detailed treatment of the local geology and stratigraphical nomenclature used by different workers. Here, we will focus on the tectonic-sedimentary evidence, based on

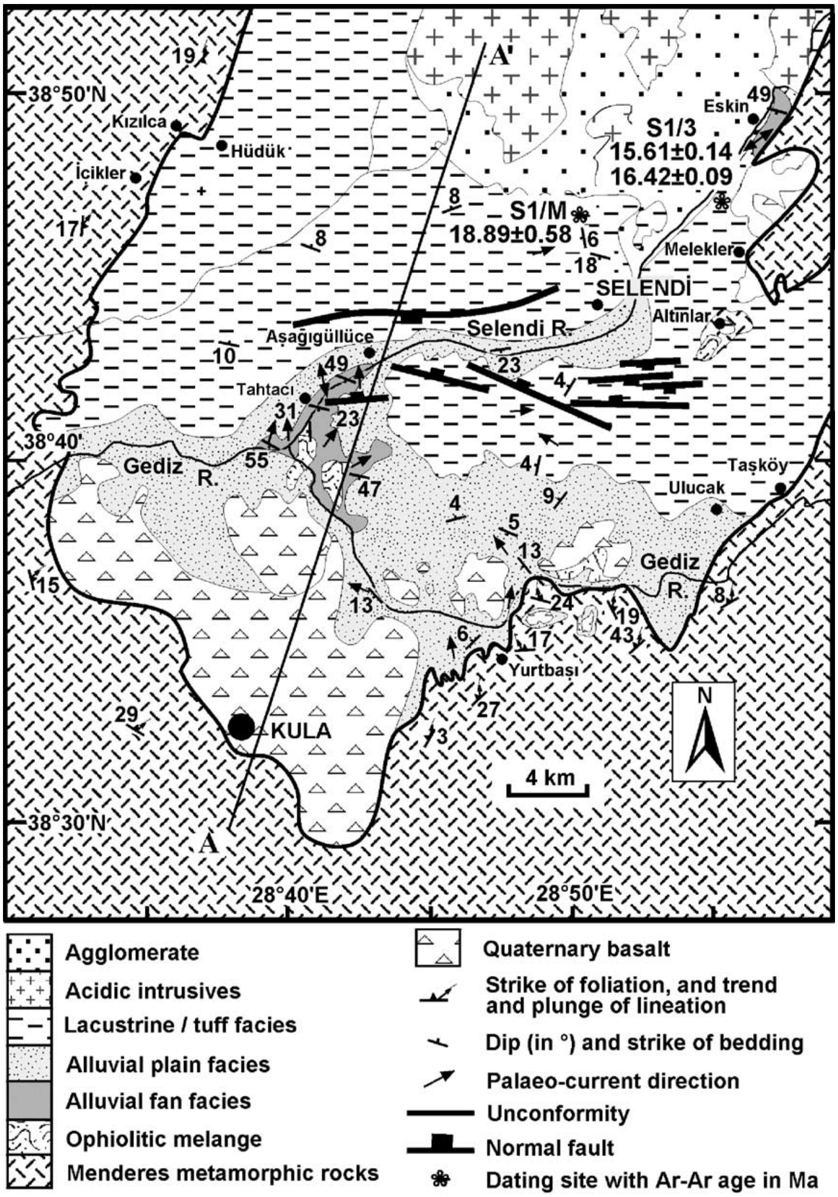

Fig. 2. Map of the Selendi Basin. Note that the margin of this basin is interpreted to have originated as an extensional detachment that was active as a zone of ductile extension prior to deposition of unconformably overlying Miocene sediments. See Fig. 7A for cross-section. Ar-Ar dates obtained during this study are indicated in bold print.

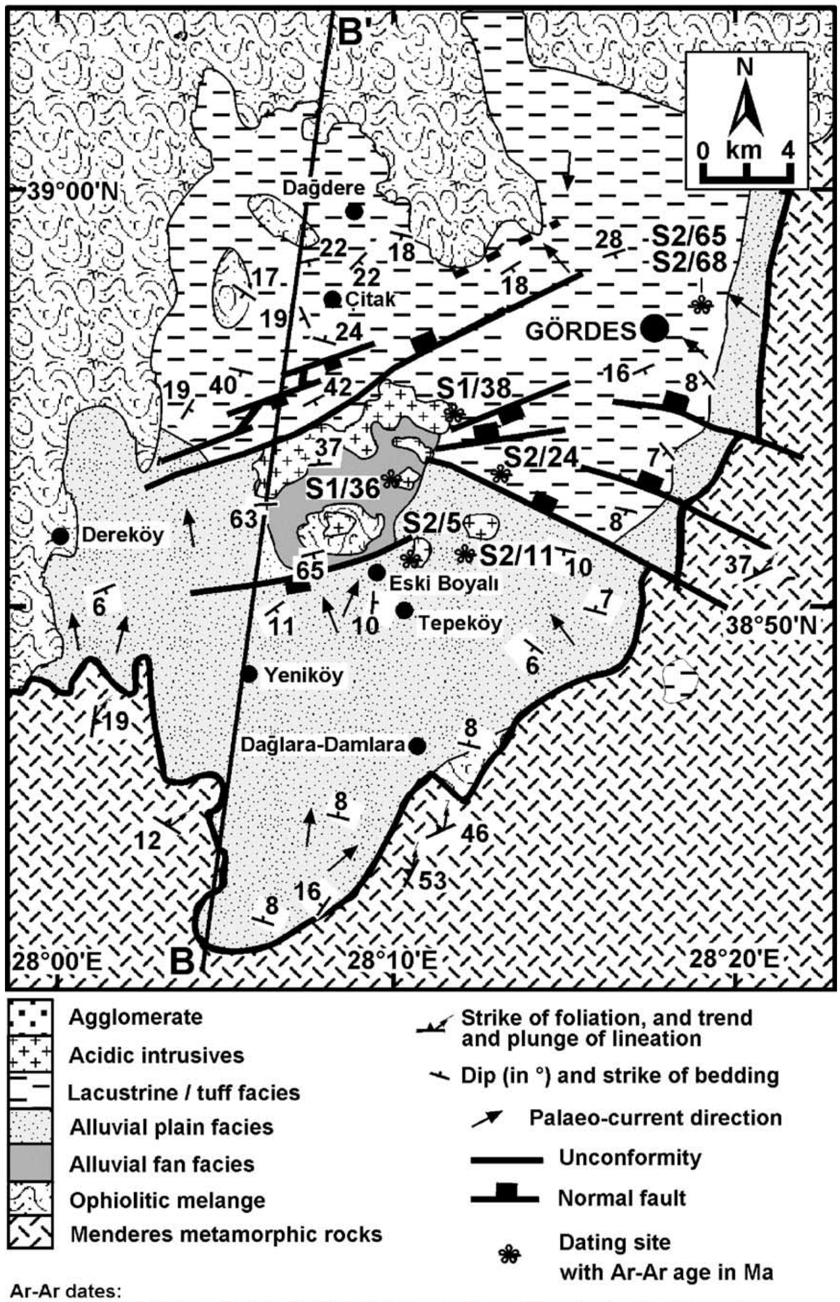

Fig. 3. Map of the Gördes Basin. As in Fig. 2, the margin of the basin is interpreted to have originated as an extensional detachment that was active prior to deposition of unconformably overlying Miocene sediments. Ar-Ar dates (with locations) obtained during this study are listed above.

some 5 months of detailed fieldwork carried out over three main field seasons. Beneath these basins, metamorphic rocks of the Menderes Massif (e.g., Dürr, 1986; Dürr et al., 1978; Figs. 2 and 3) exhibit pervasive deformation, marked by a gently dipping foliation and a stretching lineation plunging to the NE or SW (Verge, 1993, 2000).

2.1. Menderes metamorphic rocks

The metamorphic rocks underlying the Selendi and Gördes Basins (Figs. 2 and 3) locally consist mainly of gneiss, schist and marble (Fig. 1). Their upper surface is widely exposed as an erosion-resistant horizon relative to the overlying unlithified sediments (Figs. 4a,b,d and 5a). Beneath the Selendi Basin, the Menderes metamorphic rocks exhibit pervasive deformation, marked by a gently dipping foliation and an extensional lineation consistently plunging to the NE or SW (Verge, 1993, 2000; Fig. 6A,B). Top-to-the-NNE shear-sense indicators (e.g., rotated feldspar porphyroblasts; Fig. 4c) are ubiquitous. These high-stain features are well developed near the top of the exposed metamorphic succession, directly beneath the

Fig. 4. Photographs of the Selendi Basin. (a) View, looking ESE at the western margin of the basin near Hüdük, showing Menderes metamorphic rocks, with a gently inclined foliation, which plunges northwards beneath the Miocene basin fill. (b) View, looking SW, of the southern margin of the basin, near Yurtbaşı. Note the corrugated morphology of the basement–sediment interface, on a scale of hundreds of metres. This probably reflects the undulating geometry of a corrugated detachment surface. (c) Rotated porphyroblasts in gneissose rocks at the eastern margin of the basin near Taşköy, showing top NE shear, oriented towards top right of the photograph. (d) Irregular erosion surface near Yurtbaşı between Menderes metamorphic rocks and unconformably overlying poorly sorted breccias at the base of the Miocene sedimentary succession.

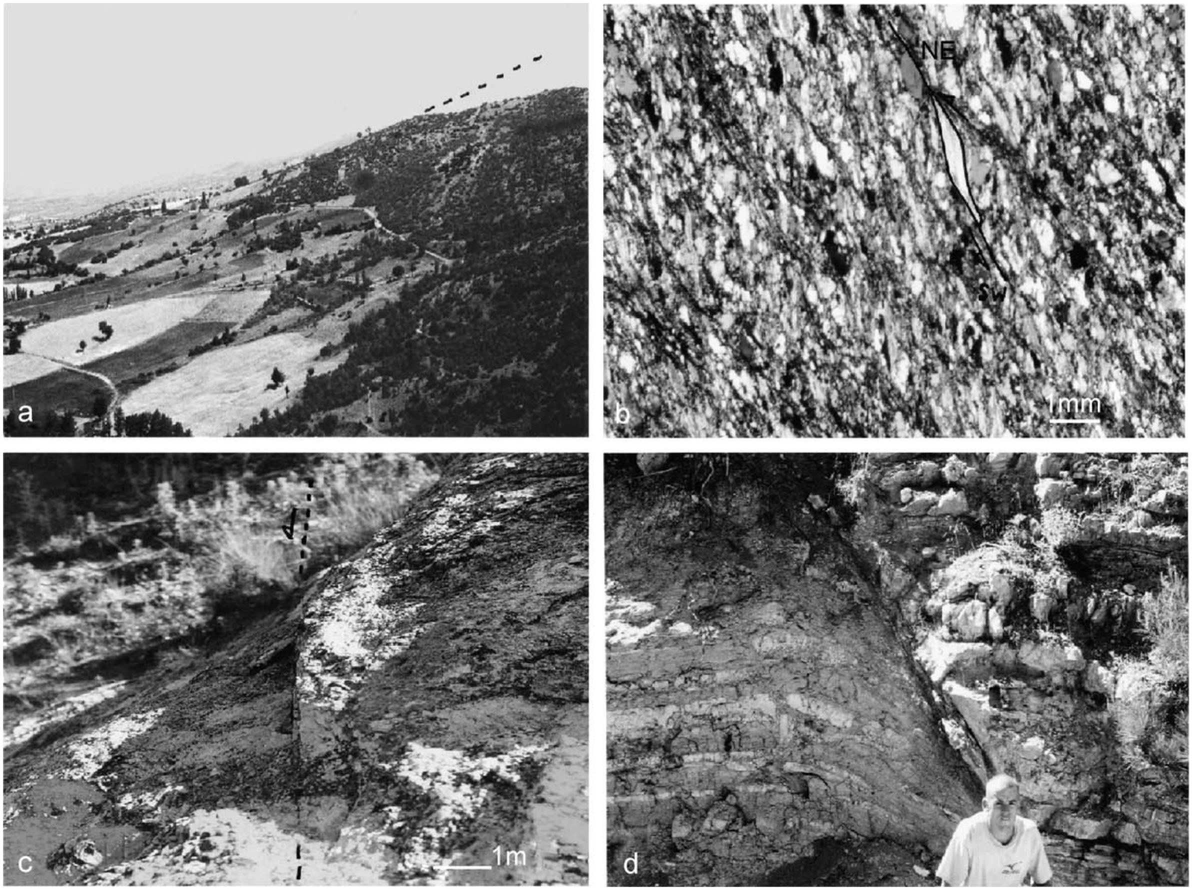

Fig. 5. Photographs of the Gördes Basin. (a) View to the northeast of the eastern margin of the basin (east of Eski Gördes), showing a gently dipping undulating contact with the Menderes metamorphic basement (dashed). (b) Sheared fine-grained schist from the southern margin of the basin (southeast of Dağlara–Damlara). Deformed mica fish, viewed with crossed polars, show a top NE shear sense (i.e., top up and to the left in the field of view). (c) Minor high-angle normal faults cutting the southern margin of the Gördes Basin near Poyraz. (d) High-angle north-dipping normal fault with downthrow to the north showing drag-folding in fine-grained tuffaceous sediements, southwest of Gördes, showing drag folds in fine-grained tuffaceous sediments.

overlying sedimentary rocks (for instance, near Yurtbaşı in the south; Fig. 2). We interpret these shear fabrics as the result of ductile to brittle extension related to unroofing of the Menderes Massif. Because the vertical extent of the exposure that can be examined in single sections is limited by the depths of river gorges, we are unable to determine if these fabrics are restricted to the exposed sections or may extend to greater depths, and thus whether the inferred extension affected a relatively narrow high-strain zone or a broader zone of ductile deformation. However, it is well known that other ductile shear zones related to crustal extension are up to hundreds of metres thick, as documented from the SW USA (e.g., Crittenden et al., 1980) and elsewhere in the Aegean region (e.g., Avigad and Garfunkel, 1991; Avigad et al., 1997).

An excellent section through the uppermost levels of the metamorphic rocks is exposed in the Gördes river gorge near Dağlara–Damlara (Fig. 3). Gneiss is overlain by strongly sheared schist, with lenticular marble intercalations. The uppermost 10–20 m of these metamorphic rocks are mylonitic and cataclastically deformed, with micro- and mesoscale faulting. North-dipping normal faults cut the exhumed basement. The exposed upper surface of the metamorphic rocks locally dips at ~23° towards the north or NNE.

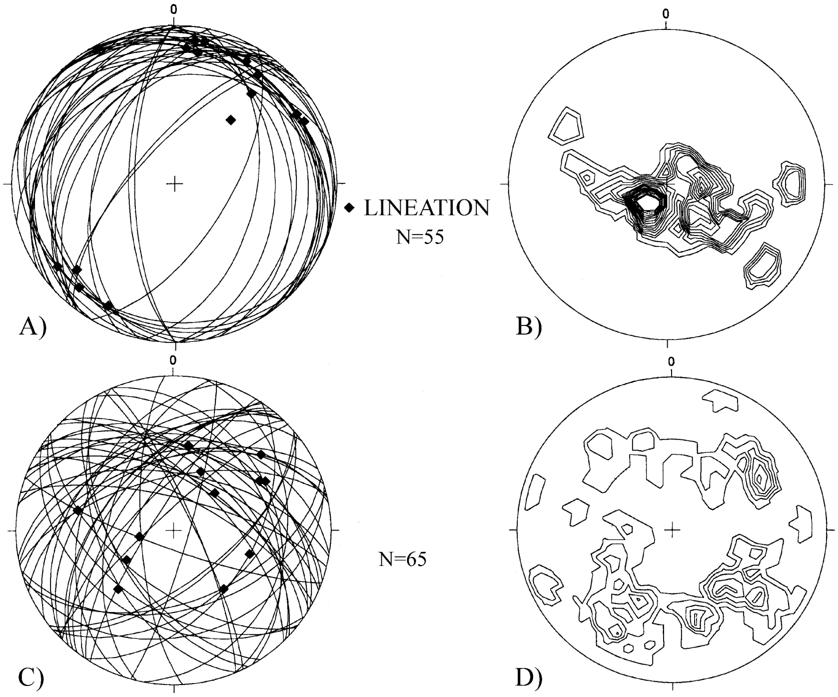

Fig. 6. Structural data from the combined Gördes and Selendi Basins plotted on stereonets, with equatorial projection. (A) Foliation and mineral elongation data from the surface of metamorphic basement, which we have interpreted as an extensional detachment fault system. (B) Contoured poles for the foliation data in (A). (C) High-angle normal fault planes with limited slickenside data indicating slip sense. (D) Contoured poles of the fault data in (C). See text for discussion.

This surface is depositionally overlain by clastic sediments, as summarised below. The top of the metamorphic rocks along the southern margin of the Gördes Basin as a whole dips northwards at shallow angles (<20°), forming a distinctive planar surface (Fig. 5a). Shear-sense indicators in the metamorphic rocks (e.g., mica "fish", rotated porphyroblasts and mineral lineations) indicate a regionally consistent top-to-the-NE extension direction (Fig. 5b).

The NW margin of the Gördes Basin is dominated by unmetamorphosed ophiolitic melange that structurally overlies the Menderes metamorphic rocks (Fig. 3). This Neotethyan ophiolitic melange was emplaced from the İzmir–Ankara suture zone to the north in late Cretaceous–early Tertiary time (e.g., Şengör and Yılmaz, 1981). Large east–west-striking high-angle normal faults locally cut this ophiolitic melange and also offset the underlying metamorphic basement. The ophiolitic rocks, which include lava, serpentinite and radiolarian chert, are unconformably overlain by Miocene sediments of clastic, lacustine and tuffaceous facies in different areas. We interpret the contact between metamorphic basement and ophiolitic melange as the trace of a regional low-angle extensional detachment, as discussed further below.

2.2. Nature and chronology of the sedimentary cover

The Neogene sedimentary rocks of the Gördes and Selendi Basins consist of coarse basal deposits, fluvial sandstones, lacustrine limestones and marls, and interbedded volcanics including lava flows and tuffs (e.g., Ercan et al., 1978, 1983a,b; Yılmaz et al., 2000). These sequences are exposed in this region's characteristic "badlands" landscape (notably in the southern part of the Selendi Basin) that reflects the ~400 m of fluvial incision that has occurred since the latest Miocene or Pliocene (Westaway et al., 2003, 2004). Representative sections of both basins are given in Fig. 7. Previous authors (e.g., Ercan et al., 1978, 1983a,b; Yılmaz et al., 2000) have subdivided the stratigraphy of these basins differently and concluded

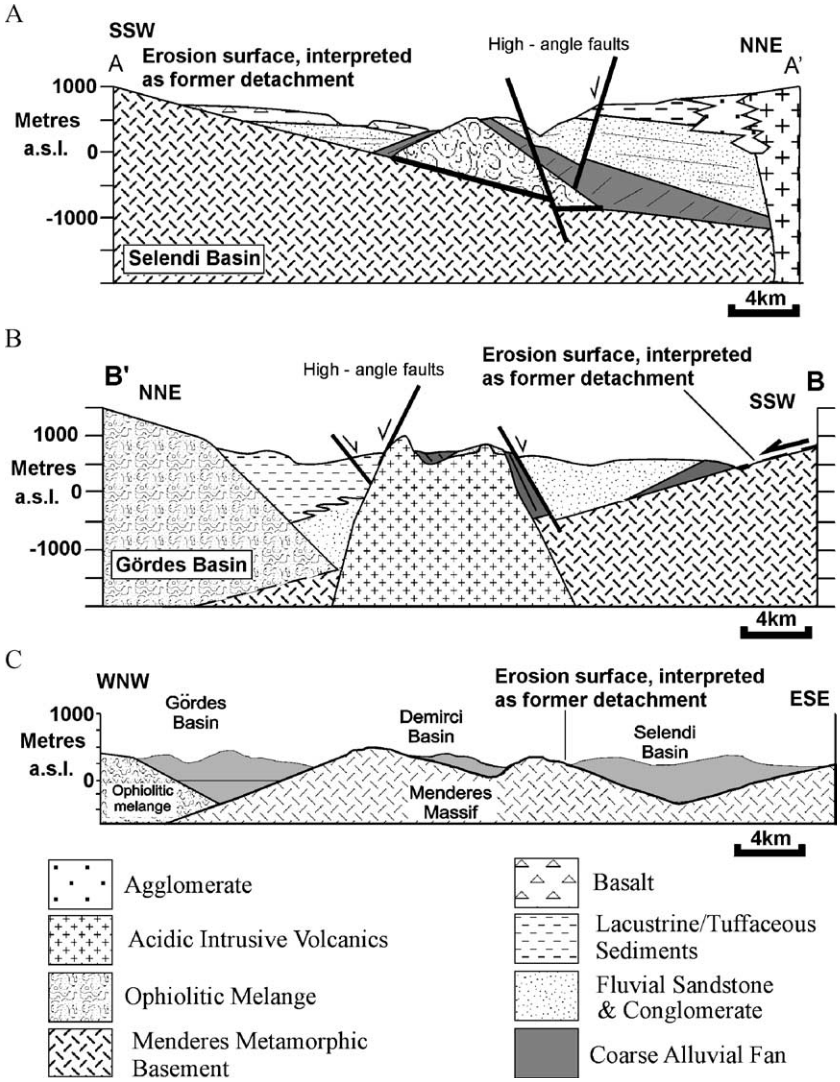

Fig. 7. Interpreted structural development. (A) Selendi Basin, showing a schematic interpretation along line A–A′ in Fig. 2. An inferred extensional detachment above the Menderes metamorphic rocks is overlain by locally preserved ophiolitic melange and Miocene sediments. This inferred structure is then cut by high-angle normal faults. (B) Gördes Basin, showing a schematic interpretation along B–B′ in Fig. 3. Gneiss and schist of the Menderes Massif are overlain by unmetamorphosed ophiolitic melange along an interpreted low-angle extensional detachment. This inferred detachment was eroded then unconformably overlain by Miocene sediments, accompanied by intrusion of silicic plutonic rocks and dissection by E–W-striking high-angle normal faults. (C) Generalised cross section illustrating the inferred relationship of the Gördes, Demirci and Selendi Basins, oriented perpendicular to the inferred top-NNE′ extension direction during the inferred phase of detachment faulting. Note the undulations in the upper surface of the Menderes metamorphic rocks, which can be interpreted as large-scale corrugations of the inferred extensional detachment. We suggest that these corrugations provided the accommodation space for the later sediment infill.

that they experienced different sedimentary and tectonic histories. However, our results indicate a very similar evolution of both of these basins that we would tentatively extend also to the adjacent Demirci Basin (Fig. 1) that we studied briefly, although our new dating results (see below) suggest that the history of the Uşak-Güre Basin farther east (Fig. 1), previously considered equivalent to the Selendi Basin (e.g., Ercan et al., 1978, 1983a,b; Metin et al., 1978), may have been rather different (see below).

The earliest deposits in the Gördes and Selendi Basins (the Hacıbekir Group of Ercan et al., 1978, 1983a,b, in the Selendi Basin, considered equivalent to the Borlu Formation of Yılmaz et al., 2000, in the Demirci Basin) consist of coarse red-brown to grey fluvial conglomerates that are only locally exposed beneath younger sediments (Figs. 2 and 3). In the Gördes Basin, the thickness of these sediments is estimated at 250 m, although exposure is too limited in the Selendi Basin to constrain thickness based on surface evidence alone. Bedding in the lower parts of these sequences is everywhere at least moderately inclined, rarely reaching 65°. In both basins extensive palaeocurrent evidence, mainly clast imbrication and cross bedding, indicates flow towards the north or NNE (Figs. 2 and 3). These sediments are mainly coarse, very poorly sorted conglomerates with angular to sub-angular clasts, up to 1 m in size, set within a red-brown sandy to muddy matrix. In most areas, the clasts are entirely metamorphic in origin, although locally, as in the centre of the Gördes Basin, ophiolite-related clasts are also present. Moderately sorted conglomerates, in places steeply dipping (~50°), are exposed in the deepest, central parts of the Selendi Basin (notably in the Gediz River gorge north of Kula; Fig. 2), where they overlie metamorphic basement and ophiolitic melange and are overlain by younger sediments and volcanics.

These basal clastic sediments are unconformably overlain in both basins by a gently inclined to sub-horizontal succession of moderately consolidated, yellow-brown fluvial sandstones and gravelstones (the Balçıklıdere Member of the Ahmetler Formation of the İnay Group of Ercan et al., 1978, 1983a,b, in the Selendi Basin). Such sediments unconformably overlie the Menderes metamorphic rocks in the Selendi Basin, as seen for instance at Yurtbaşı (Fig. 4b,d), Taşköy (Fig. 2) and along its western margin (Fig. 4a). In the Gördes Basin, basal conglomerates, exposed for example near Dağlara–Damlara (Fig. 3), are unconformably overlain by gently dipping, fine-grained fluvial sediments that onlap metamorphic rocks along the southern, eastern and western margins of the basin, without evidence of faulting (Fig. 5a). Both basins are thus characterised by an overall fining-upward succession of trough-cross bedded, channelised pebblestones within fine-to medium-grained sandstones. Iron-pan is present within coarser units, creating resistant horizons. Lignite horizons are locally developed. Palaeocurrent indicators (pebble imbrication and cross bedding) indicate palaeoflow, generally towards the north in both basins (Figs. 2 and 3).

This fluvial clastic succession is conformably overlain by gently dipping, typically subhorizontal, pale brown to white, fine-grained, lacustrine silts, sands and marly limestones, with interbedded pale tuffaceous sediments. A general fining-upward trend is seen, with an upward increase in fine-grained siliceous air-fall tuff. The lower part of this succession (the Gedikler Member of the Ahmetler Formation of the İnay Group of Ercan et al., 1978, 1983a,b, in the Selendi Basin) consists of fine-grained sandstone and siltstone, with extensive trough cross-bedding. Higher up (in the Ulubey Formation of the İnay Group of Ercan et al., 1978, 1983a,b), the sediments are dominated by carbonaceous muds, which exhibit mud cracks, abundant plant material and common palaeosols. Symmetrical ripples are seen in coarser-grained siltstones. Characteristic freshwater molluscs in this lacustrine unit include *Unio* (a bivalve), and *Viviparus* and Neritinids (gastropods).

The stratigraphically lowest unit of coarse clastic sediments within the Selendi Basin, the Hacıbekir Group, is generally accepted as early Miocene (e.g., Seyitoğlu, 1997; Westaway et al., 2003, 2004). The metamorphic lithologies on the eastern margin of the Gördes Basin are locally cut by undeformed leucocratic dykes containing coarse quartz, feldspar, muscovite and tourmaline These intrusions were K–Ar dated as 24.2±0.8 and 21.1±1.1 Ma (Seyitoğlu, 1992). The ductile shearing of this basement, reflected in its foliation, thus pre-dates the early Miocene; assuming this reflects extensional detachment faulting, this can only have occurred at an earlier stage. Clasts of similar leucocratic lithologies within the basin fill indicate a lower limit for the age of these sediments. Silicic intrusive rocks, cutting the centre of the Gördes Basin and disrupting the earlier relatively steeply dipping part of its sedimentary fill (thus again providing a lower age constraint for the ages of these sediments) were dated as late early Miocene (18.4±0.8 and 16.3±0.5 Ma), close to the range of ages determined during this study for the stratigraphically lowest bedded tuffaceous sediments higher up the sequence in both basins (see below).

However, the age of the overlying İnay Group remains controversial. A late early Miocene–early middle Miocene age was assigned by Seyitoğlu (1997) and Seyitoğlu and Benda (1998) based on pollen and K–Ar dating of interbedded lavas, whereas mammalian biostratigraphic and magnetostratigraphic evidence may support a late Miocene (~7 Ma) age for the upper part of the Balçıklıdere Member in the Uşak-Güre Basin, suggesting that the stratigraphically higher part of the İnay Group could be younger (see Bozkurt, 2003; Westaway et al., 2004 for discussion).

The field relationships between the sediments of the Selendi Basin and the local volcanism have also been disputed. Ercan et al. (1978, 1983a,b) identified four different phases of volcanism, but Seyitoğlu (1997) reduced this number to two: a phase that he K–Ar dated to 18.9±0.6 Ma, which he placed between the end of deposition of the Hacıbekir Group and the start of deposition of the İnay Group; and a younger phase (represented by the agglomerate that crops out in the northern part of the basin; Fig. 2) that he K–Ar dated to 14.9±0.6 Ma and described as "interfingering" with the İnay Group. However, it is clear from his paper (see his Fig. 6) that he placed this phase of volcanism near the base of the İnay Group, just above the unconformity above the Hacıbekir Group. Volcanism with essentially the same K–Ar age occurred near the base of the İnay Group in the Uşak-Güre Basin, where the exposure of the field relationships is very clear (see, e.g., Seyitoğlu, 1997; Westaway et al., 2003, 2004). In the Demirci Basin, Yılmaz et al. (2000) reported that the basal Borlu Formation passes gradually upward, without a major unconformity, into the lacustrine and in part tuffaceous Köprübaşı Formation that is associated with the local Okçular volcanics. There is thus no local equivalent of the thick fluvial sequence in the Selendi Basin that has been assigned by Ercan to his Balçıklıdere Member, and the tuffaceous sediments in the Demirci Basin are reported to be folded (Yılmaz et al., 2000), whereas those in the Selendi Basin are subhorizontal. Furthermore, the Köprübaşı Formation has previously been assigned an age of ~18–14 Ma based on the age span of volcanism elsewhere in the region considered equivalent to the Okçular volcanics (Yılmaz et al., 2000). Until now, it has thus been thought much older than the tuffaceous unit in the Selendi Basin (Ercan's Gedikler Member), which has been considered late late Miocene in age (cf. Westaway et al., 2004).

However, our re-mapping (Purvis, 1998; Figs. 2, 3) suggests a different sequence of events for the Selendi and Gördes Basins. We found no evidence that the agglomerate that crops out in the northern Selendi Basin interfingers with the lower part of the sediment attributed in previous studies to the İnay Group. As illustrated in Fig. 7A, we regard it instead as synchronous with the deposition of the tuffaceous sediment in the upper part of the fluvial sequence (Ercan's Balçıklıdere Member) and the overlying lacustrine sequence (Ercan's Gedikler Member).

We believe that these apparent discrepancies arise from two main factors. The first is regional correlation of volcanism between the individual basins. The reported presence of intercalated volcanic rocks in the Uşak-Güre Basin should not be taken as representative of the basins as a whole, as we did not recognise such volcanics in the lower part of either the Gördes or Selendi Basins. The second is structure. Previous workers (e.g., Seyitoğlu, 1997) have seen the N–S margins of all four basins as being bounded by N–S faults whereas during this study we observed transgressive contacts along the margins of each of the Selendi, Gördes and Demirci Basins; Westaway et al. (2004) note a similar transgressive relationship along the margins of the Uşak-Güre Basin. In addition, in the past important E–W-striking normal faulting, with cumulative throws of hundreds of metres (see below), has not previously been taken into account when interpreting the stratigraphy of the Selendi and Gördes basins. We therefore regard the fine-grained tuffaceous sediments from both the Selendi and Gördes basins as documenting the earliest explosive calc-alkaline volcanism in the vicinity of these basins.

High precision $^{40}Ar/^{39}Ar$ dating of biotite and feldspar was carried out during this study on mineral grains separated from the stratigraphically lowest tuffaceous sediments in both the Gördes and Selendi Basins. Details of the stratigraphy, sample locations, dating methods and the results are given by Purvis et al. (in press). An early Miocene age (21.7±0.04–16.42±0.1 Ma) is thus assigned to the lacustrine and tuffaceous sediments overlying the thick fluvial sediments of both the Selendi and Gördes basins, based on this work. These results indicate that the underlying fluvial sediments of the lower part of the basin are older than the dated tuffs and an early Miocene age is thus assigned, taking account of the age constraints from the leucocratic rocks mentioned above.

These results suggest that the stratigraphic correlation scheme, deduced for these SW–NE-trending basins by Ercan et al. (1978, 1983a,b) and used in subsequent publications on them, is problematic. They indicate, for instance, that the mainly fluvial unit in the Uşak-Güre Basin that was assigned by Ercan to the Balçıklıdere Member and whose deposition spanned ~15 to ~7 Ma (cf. Westaway et al., 2004), is much younger than the apparently similar unit in the Selendi Basin that Ercan also assigned to the Balçıklıdere Member, deposition of which now appears to have ended by ~19 Ma given the oldest of our new Ar–Ar dates from the overlying tuffaceous lacustrine unit (Fig. 2). They also suggest that these tuffaceous sediments in the Selendi Basin that were assigned to Ercan's Gedikler Member are also much older than previously thought, being dated to ~19–16 Ma (Fig. 2), suggesting that they correlate with the Köprübaşı Formation of (Yılmaz et al., 2000) in the adjacent Demirci Basin, even though the former is subhorizontally bedded and the latter is reported to be folded (although this may be extension related; see below).

These indications of difficulties with the existing stratigraphic schemes are a major reason why we do not use previously defined stratigraphic terms, but prefer instead to designate "facies associations" except when summarising how our descriptions relate to those by previous workers in key localities. Further work is evidently needed to re-correlate the rocks assigned by Ercan to his İnay Group in different basins in this region. This difficulty may relate to the inadequacy of the existing stratigraphic units, which are presently defined as a layer-cake sequence despite the pronounced lateral variations in thickness and facies that clearly exist. Such sedimentary variation is indeed recognised in the existing stratigraphy of the Gördes Basin (Seyitoğlu and Scott, 1994a,b), in which the coarse marginal facies, termed the Tepeköy Formation, are seen as spanning the Miocene, whereas more basinal facies are subdivided into the coarse clastic Dağdere Formation and the overlying finer gained tuffaceous to lacustine Kulukköy Formation, both assigned an early Miocene age. However, here again such formation names tend to obscure much of the facies variation that actually exists. Detailed sedimentary logs for the Selendi and Gördes Basins will be presented elsewhere. However, in the context of the present paper, the most important point is that the stratigraphically lowest sediments above the Menderes metamorphic basement are accepted as being of early Miocene age.

2.3. Deformation

Both the Selendi and Gördes basins exhibit a phase of north–south extension, characterised by high-angle (>35°) E–W-striking normal faults, with both northerly and southerly downthrow (Fig. 6C,D). Some small normal faults also cut the basin margins (Fig. 5c) but in most places these margins are not normal-fault-bounded.

The E–W-striking faulting is concentrated near the centres of both the Selendi and Gördes basins. The directions and amounts of throw on individual faults can be directly measured by offsets of correlated strata within the Selendi Basin (notably, near Selendi in Fig. 2). In both basins, these faults cut all the sedimentary units and thus cannot be dated accurately. However, such faults are reported to pre-date the adjacent Quaternary lava flows exposed around Kula in the southern part of the Selendi Basin (Seyitoğlu et al., 1997; Yılmaz et al., 2000, 2001), as supported by R. Westaway (pers. comm., 2004) who finds no evidence that these faults cut these lava flows. The probable age of this high-angle faulting is thus late Miocene–early Pliocene, assuming a mainly middle Miocene age for the volcanogenic units in the upper part of the basin. This later-stage faulting created asymmetrical grabens up to ~10 km wide in the centre of both basins, as best exposed in the Selendi Basin (Fig. 2). Some of these fault blocks are tilted towards the axes of these high-angle extensional fault zones as shown by measured dips of sediments in the individual fault blocks (Purvis, 1998).

Individual faults typically have observed offsets of <20 m and exhibit only subtle topographic expressions. Adjacent lithologies have experienced local

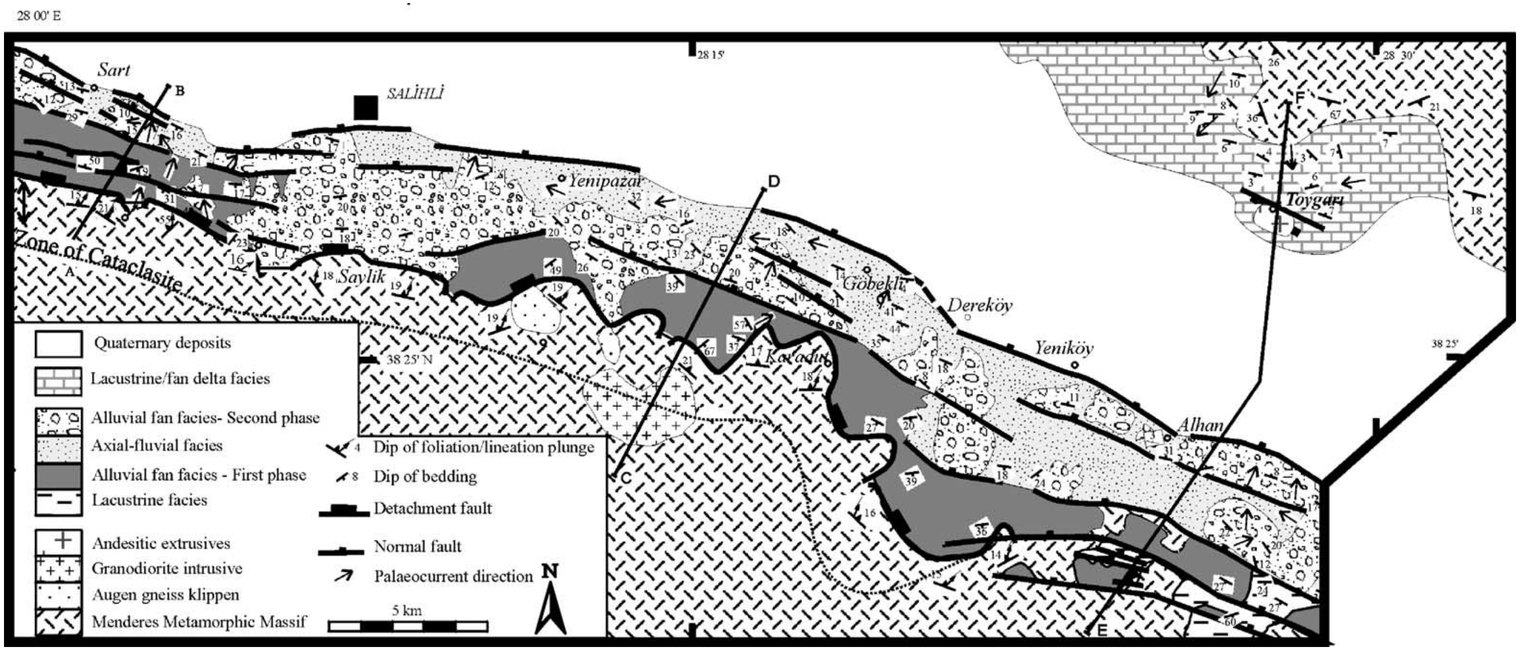

Fig. 8. Geological map of part of the southern margin of the Alaşehir Graben based on detailed mapping carried out during this work. The map shows the main lithologies, faults and gives a summary of palaeocurrent evidence. The sections marked are shown in Fig. 9.

drag folding and brecciation (Fig. 5d). Limited slickenside data from fault planes indicate an overall dip-slip sense of displacement, with a minor dextral component. However, slickensides are rarely observable in these poorly consolidated sediments.

During the progressive incision of these basins, erosion products have been removed from the Selendi and Gördes basins by the Gediz River and transported towards the Aegean Sea. The present westward course of this river is essentially at right angles to the palaeocurrents determined within the clastic sediments of the Selendi and Gördes Basins (Figs. 2 and 3), providing an indication of the drastic geomorphological reorganisation that has occurred since the Miocene (Purvis, 1998; Purvis and Robertson, 2000; Westaway et al., 2004).

3. The Alaşehir Graben

We now investigate, for comparison, the Alaşehir (Gediz) Graben further south, an E–W-trending structure that is >150 km long and >40 km wide in the west but narrows eastwards (Fig. 1). The Menderes Massif to the south forms a regional-scale culmination (200 km N–S by 300 km E–W), mainly composed of augen gneiss, schist, phyllite, quartzite and marble (e.g., Dürr et al., 1978; Akkök, 1983; Dürr, 1986; Şengör et al., 1984; Dora et al., 1990; see Bozkurt and Oberhansli, 2001 for recent review). The sedimentary fill of the Alaşehir Graben was initially explored for minerals, providing an outline of the stratigraphy and structure (Yılmaz, 1986; Yağmurlu, 1987). More recently, the metamorphic and extensional history of the adjacent Menderes Massif has been investigated in some detail (e.g., Hetzel, 1995; Hetzel et al., 1995a,b; Gessner et al., 2001; Lips et al., 2001). The overlying sediments and their related structural geology have also been investigated (e.g., Emre, 1988, 1996; Itzan and Yazman, 1990; Paton, 1992; Cohen et al., 1995; Emre and Sözbilir, 1997; Koçyiğit et al., 1999; Yılmaz et al., 2000; Sözbilir, 2001, 2002; Seyitoğlu et al., 2002). However, the sedimentology has been largely overlooked until now. Various workers have set up different stratigraphical nomenclatures for the Neogene sediments of this graben (e.g., Koçyiğit et al., 1999; Sözbilir, 2001; see Bozkurt, 2003 for a recent review). However, as for the N–S-trending basins further north, these schemes do not fully reflect the wide range of facies variation that exists.

This study involved detailed mapping and collection of sedimentary and structural evidence from a 40-km long segment of the southern margin of the Alaşehir Graben, between Salihli and west of Alaşehir (Fig. 8). A 15-km long segment of the northern margin of the graben, around Toygarı was also studied (Fig. 8, upper right corner). This northern margin has lower relief and is much less dissected by erosion than the southern margin. Three representative cross-sections of the graben were constructed based on the field mapping and structural data (Fig. 9).

3.1. Structure of the southern margin

Many studies have noted the abrupt contact between metamorphic basement and Neogene sediments along the southern margin of the Alaşehir Graben. Most of these studies have interpreted this contact as a regionally extensive, presently low-angle (10–20°), north-dipping normal fault surface, separating the metamorphic rocks of the Menderes Massif from the overlying Neogene sedimentary fill of the Alaşehir Graben (e.g., Hetzel et al., 1995a; Hetzel et al., 1995b; Purvis and Robertson, 1997, 2001; Koçyiğit et al., 1999; Yılmaz et al., 2000; Sözbilir, 2001, 2002; Seyitoğlu et al., 2002). This structure is exposed for several kilometres in both the dip and the strike directions and has a slightly convex-upward profile (Fig. 10a,b). A sheared, locally cataclastic, surface dips at 12–20° towards the north and is laterally continuous for >30 km. This surface is disrupted by a number of structural discontinuities oriented at a high angle to its average trend (Fig. 8). The irregular trace of this fault plane is thus not merely a topographic effect but results from structural disruption into discrete fault segments that are internally smooth and regularly dipping.

These individual fault segments are slightly offset laterally and vertically along these structural breaks, which are marked by complex zones of deformation up to 1 km wide (Purvis, 1998). Sedimentological evidence outlined below indicates that these structural features influenced clastic sediment supply to the graben and are not merely the result of later faulting.

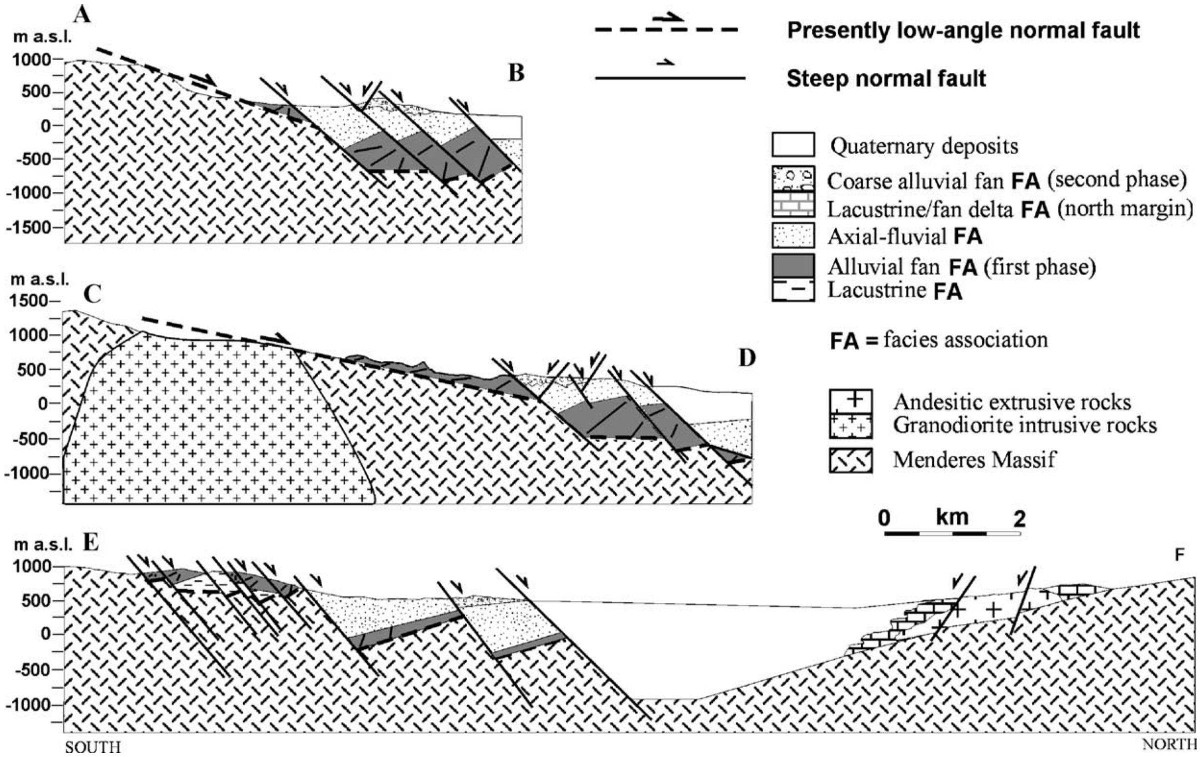

Fig. 9. Geological cross sections of the Alaşehir Graben showing the inferred relationships of facies to the low-angle detachment fault (dashed line), relatively high-angle faults and basement lithologies. For location of sections, see Fig. 8.

Additional, larger (kilometre-scale) offsets, bounded by N–S high-angle faults, have also been reported, both to the west (20 km west of Salihli; Seyitoğlu and Scott, 1996) and east (south of Alaşehir; Yılmaz et al., 2000).

Basement lithologies of the Menderes Massif beneath this presently low-angle normal fault surface are foliated, mainly mica-schist, phyllite, quartzite and marble (Figs. 10a, 11A), typical of the higher levels of the metamorphic tectono-stratigraphy as a whole (Hetzel, 1995). These lithologies exhibit intense shearing over a vertical interval of several tens of metres beneath the fault surface. Early ductile fabrics, as indicated by the presence of small-scale isoclinal folds, are overprinted by brittle fabrics, as recorded by the presence of cataclasis, mylonites and fracturing (Hetzel et al., 1995b). Shear-sense indicators (e.g., C–S fabrics) and mineral lineations indicate a consistent top-to-the-NE sense of shear. In addition, several small fault-bounded slivers of high-grade augen gneiss are exposed structurally above the lower-grade metamorphic lithologies. These may represent slivers of the original hanging wall that were detached and transferred to the footwall during exhumation.

Sub-rounded to elongate plutons (for instance the Salihli Granodiorite; Fig. 8) are present south of the Alaşehir Graben. The internal structural fabric within these plutons is oriented sub-parallel to the regional foliation within the high-grade metamorphic rocks in the central part of the area mapped (Hetzel et al., 1995b). These granodiorites show no metamorphic imprint older than this fabric, in contrast to the Menderes country rocks, which show extensive evidence of earlier compressional deformation including foliations, folding and cleavage. However, both the country rocks and the granodiorite exhibit similar low-angle foliation, stretching lineation (towards the NE) and localised cataclasis (Hetzel et al., 1995b; Purvis, 1998). It is therefore unlikely that the structural fabrics in the Menderes metamorphic rocks and the granites are entirely unrelated and we suggest that that they developed as a result of extensional

Fig. 10. Field photographs of the contact between metamorphic basement and Neogene sediments at the southern margin of the Alaşehir Graben. (a) View of this smooth low-angle north-dipping surface looking westward from south of Salihli, showing its low-angle dip to the north. (b) Close-up view of this surface from south-east of Salihli (near Keskınler), showing its low-angle dip to the north.

processes that occurred simultaneously with granite intrusion. These deformed granitic rocks have been radiometrically dated by the Ar–Ar method, yielding plateau ages of 19.5$\pm$1.4 Ma (Hetzel et al., 1995b) for amphibole (thought to represent the intrusion age) and 12.6$\pm$0.4 (Hetzel et al., 1995b) to 7.1$\pm$1.0 Ma (Lips et al., 2001) for biotite (inferred cooling ages). These dates record closure of the minerals to argon as a result of cooling through temperatures of ~500 to ~300 °C. Given the inferred local geothermal gradient of ~35 °C km^{-1} (e.g., İlkışık, 1995), these data imply exhumation from depths of ~10–15 km since early to middle Miocene (R. Westaway, pers. comm., 2004).

In addition, fluid inclusion data indicate that accelerated cooling took place in Plio–Quatenary time (Gessner et al., 2001). Such cooling implies exhumation, which could in principle have several causes, including erosion and crustal extension. In view of the consensus that rifting of the Alaşehir Graben was active during Plio–Quaternary time, it is likely that movement on one, or several, normal faults was active during this time. However, the cooling evidence cannot by itself confirm whether the extension occurred along the presently exposed low-angle normal fault along the southern margin of the graben or a deeper-level structure.

The footwall is directly overlain by coarse clastic sediments. The presently low-angle normal fault surface is locally seen to be deformed, associated with post-depositional shearing along it. In many places, it is separated from overlying conglomerates by a zone of friable gouge, 1–2 m thick. This gouge contains clasts from both the metamorphic rocks beneath and sedimentary rocks exposed above. Rare slickensides indicate a NNE sense of slip (Purvis, 1998).

Our observations show that the sedimentary fill is everywhere underlain by a presently low-angle normal fault. In contrast, Cohen et al. (1995) and Dart et al. (1995) reported an unconformity offset by high-angle normal faults in the southeast part of the area we mapped (Fig. 8). However, we have observed that numerous high-angle normal faults cut the metamorphic basement in this area, but a pre-existing low-angle normal fault surface and associated cataclasite zone are still locally observable. We correlate this structure with the presently low-angle normal fault that separates the Neogene sedimentary rocks from the Menderes Massif further west.

In the southeasterly area (Fig. 8), the metamorphic footwall is cut by a series of high-angle normal faults, creating a series of mainly back-tilted blocks. Similar high-angle faults cut the sedimentary fill of the graben further west where lower-plate metamorphic rocks are less widely exposed (Fig. 8). Taking the area as a whole, north-dipping high-angle (>40–60°) normal faults transect the Menderes Massif, its low-angle surface and the overlying sedimentary succession. Nowhere have we observed high-angle faults sealed by overlying sediments, which places constraints on the timing of the high-angle faulting (see below).

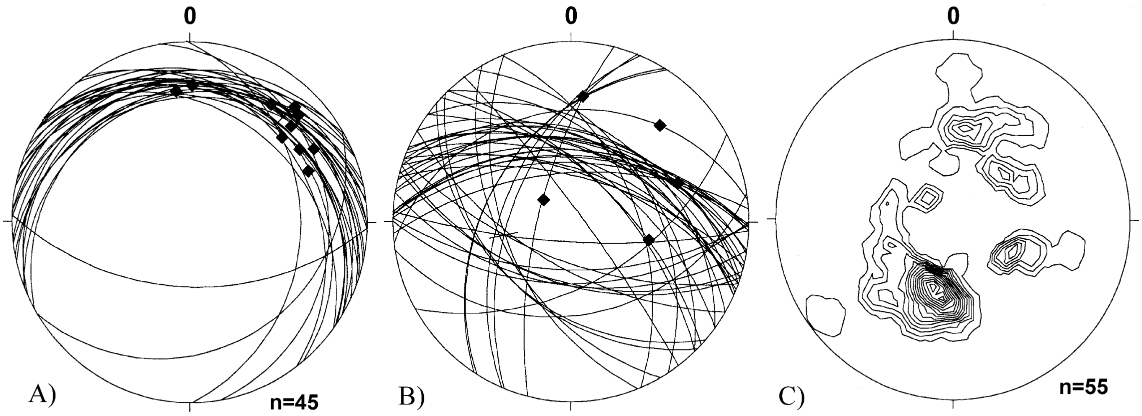

Fig. 11. Structural data from the southern margin of the Alaşehir Graben, plotted on stereonets with equatorial projection. (A) Equal area projection of foliation data from metamorphic rocks of the metamorphic rocks of the Menderes Massif beneath the regional presently low-angle normal fault bounding the overlying sediments. (B) High-angle normal fault planes from the overlying Neogene sediments. These faults post-date all the sedimentary units of the southern margin. Rarely observed slickensides are marked as diamonds; (C) contour plot of poles to the high-angle faults in (B). See text for discussion.

These high-angle normal faults vary from north- (synthetic) to south- (antithetic) dips and are associated with both forward- and back-tilting of strata (Figs. 9, 11B,C). In some areas early alluvial fan conglomerates (commonly reddish) are juxtaposed directly against younger sediments of axial fluvial facies (commonly greenish). Major faults cutting the sedimentary cover (with more than tens of metres of displacement) are generally eroded and poorly exposed. By contrast, smaller faults are commonly well exposed and exhibit extensional kinematic indicators.

The boundary between the exhumed Neogene sediments and the Quaternary–Recent alluvial plain is defined by an important active normal fault, which slipped during the 1969 Alaşehir earthquake (e.g., Eyidoğan and Jackson, 1985). The footwall escarpment of this fault is heavily eroded due to the unconsolidated nature of the sands and gravels. However, in some localities this fault can be traced as a prominent topographic break along the southern margin of the Salihli plain, or from the juxtaposition of sediments of different ages. The present-day margin of this plain is delineated by the last active fault break, which occurred during the 1969 Alaşehir earthquake.

The initial dip of the presently low-angle normal fault separating the Menderes metamorphic rocks from the oldest Neogene sediments is an important question. Elsewhere, presently low-angle normal faults (<20°), similar to those of Aegean Turkey, either resulted from progressive low-angle extensional faulting (Abers, 1991; Livacarri et al., 1995), or tilting of initially high-angle normal faults to a lower angle as extension continued (Axen, 1992; Buck, 1988). The initial dip of this normal fault bounding the Alaşehir Graben can be estimated from its present-day dip of ~16° and the dip of the overlying sediments which reaches ~60° (see below). Using the standard distributed vertical simple shear construction of Westaway and Kusznir (1993), it can be estimated as ~64°. It is thus inferred that this fault formed at quite a steep initial dip, and that its presently low-angle dip resulted from a combination of back-tilting while it was active and subsequent back-tilting when, as is likely, extension switched to other faults in the vicinity (see below).

3.2. South-margin sediments

Four main lithofacies are recognised (Fig. 12), that we designate: the Lacustrine Facies Association (1), The Muddy-to-Sandy Alluvial Facies Association (2), the Axial-Fluvial Facies Association (3) and the Coarse Alluvial Facies association (4). These correspond to the Alaşehir Formation (1), the Kurşunlu Formation (2 and 3) and the Sart Formation (4) of Seyitoğlu et al. (2002) and earlier publications (see Sözbilir, 2001; Bozkurt, 2003 for correlations with

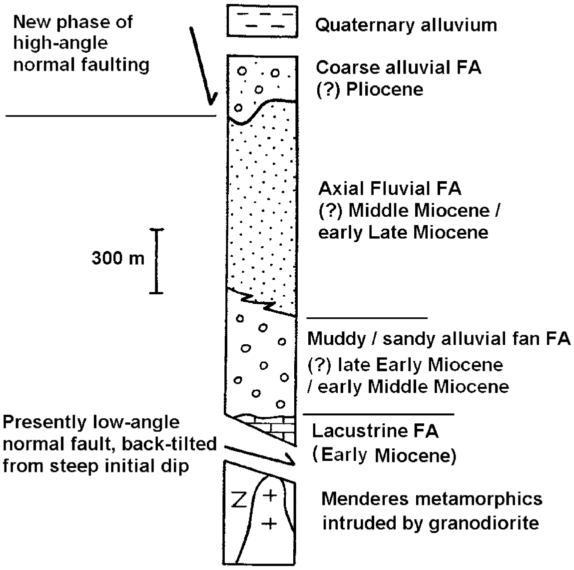

Fig. 12. Stratigraphic nomenclature for the late Cenozoic sediments of the Alaşehir Graben used in this paper. Strong lateral facies variation exists and we thus prefer to describe units as informal facies associations, rather than as formally defined formations. See text for explanation.

other stratigraphic schemes). The maximum dip of these sediments decreases up-section towards the present-day active basin margin, from ~60° in the lowest sediments to ~30° in the upper sediments. However, there is much local variation associated with variably tilted fault blocks.

The Lacustrine Facies Association forms the stratigraphically lowest sedimentary rocks. These sediments are mainly finely laminated mudstones, marls and limestones. There are also occasional sandstones and conglomerate intercalations, individually up to 3 m thick. The main clast lithologies are schist, marble, phyllite and quartzite. Sediments close to the footwall are deformed by shear bands and occasional north-verging asymmetrical folds. The upper levels of the lacustrine sediments are intercalated with poorly sorted conglomerates, with mainly angular to sub-angular clasts. In most areas this lacustrine facies is not exposed and instead the footwall is directly juxtaposed against extensive alluvial conglomerates. These dip southwards at up to 60° towards the basal fault surface. Normal faults cut these alluvial sediments and are back-tilted towards this presently low-angle normal fault.

The Lacustrine Facies Association is separated from the overlying Muddy-to-Sandy Alluvial Facies Association by a low-angle unconformity. Within the overlying interval, Cohen et al. (1995) recognised numerous lithological subdivisions, based on detailed facies and colour variations. However, our comparison with the wider area further west suggests that these sediments can be simplified into two clastic facies association. In particular, two phases of later stage alluvial deposition, as reported by Cohen et al. (1995), are re-interpreted by us as proximal to distal equivalents of the same alluvial fan system, offset by later stage E–W-striking high-angle normal faults. It should be noted that colour variation, especially reddening, is commonly related to diagenesis and should not, by itself, be used to identify different facies or infer the relative age of depositional units. Reddening is frequently patchy and commonly corresponds to local zones of fluid flow within different sedimentary lithologies.

The Muddy-to-Sandy Alluvial Facies Association (~500 m thick) comprises coarse, angular, poorly sorted conglomerates, with typically sub-angular grains and clasts. These are mainly schist, quartzite and marble, ranging from millimetres to metres in size. Both clast- and matrix-supported conglomerates are present, the latter set in muddy sandstone and sandy mudstone. Dips are variable (20–60°), mainly towards the south (Fig. 8).

Conglomerate bodies within the Muddy-to-Sandy Alluvial Facies Association, which are up to several km wide and long, were fed from point sources. The positions of these depositional conduits correspond to the areas where the trend of the basal fault surface departs markedly from its regional orientation, as shown in Fig. 8 (e.g., SW of Salihli). We interpret these features as structural breaks that already disrupted the presently low-angle normal fault surface at an early stage and thus influenced sedimentation (Fig. 9). The clasts in the conglomerates were derived from the Menderes Massif exposed to the south, as indicated by palaeocurrent data (mostly clast imbrication and cross-bedding) that is shown in summary form (as arrows) in Fig. 8, and is presented in full elsewhere (Purvis and Robertson, in press). The conglomerates are coarsest in the axial zones of the individual lobes and pass laterally into finer conglomerates with a greater concentration of sand and silt.

Interlobe areas are dominated by sand and silt, with only rare conglomerate. In some areas these lobes are partly obscured by younger conglomerates of the Coarse Alluvial Fan Facies Association (see below), or by E–W-striking high-angle normal faulting. However, their original presence is confirmed by palaeocurrent patterns (Fig. 8). The dominant flow was northwards, at right angles to the strike of the Alaşehir Graben during accumulation of this Muddy-to-Sandy Alluvial Facies Association (Cohen et al., 1995; Purvis and Robertson, in press).

The Muddy-to-Sandy Alluvial Facies Association is unconformably overlain by the Axial-Fluvial Facies Association, ~500 m thick. Typically, there is an angular discordance of ~20° between these two units, but in places they interfinger on an outcrop scale. These higher sediments are poorly consolidated and erosionally dissected, giving rise to a rugged badlands topography, with >100 m relief. Dips range from 0–40° and are dominantly southward. These sediments are mainly coarse conglomerates with sand-to gravel-sized intercalations, locally with well-developed clast imbrication. The dominant clast type is quartzite, with minor schist, phyllite and marble. Clasts are mainly sub-angular to moderately well-rounded and are slightly more texturally mature than in the underlying conglomerates. The gravels and sands commonly exhibit cross stratification of both planar and trough type. Palaeocurrents are mainly west directed; thus mainly axial with respect to the present Alaşehir Graben (Fig. 8; Purvis and Robertson, in press). Channels, individually up to several metres in amplitude by 1 m deep, include sub-angular clasts of lithified clastic sediment (e.g., pebblestone) that were recycled from the subjacent Muddy-to-Sandy Alluvial Facies Association. Minor lignite interbeds with freshwater gastropods also occur within these higher conglomerates. Pollen is dominated by swamp cypress and Mediterranean pines (see Purvis, 1998).

In some areas of the southern margin of the Alaşehir Graben a contrasting unit of alluvial conglomerates (up to 350 m thick), termed the Coarse Alluvial Facies Association, is found unconformably overlying, or in faulted contact, with the Axial-Fluvial Facies Association (the Alluvial fan facies-Second phase, in Fig. 8). Its basal contact is locally strongly erosive into the underlying depositional units. This younger facies is exposed from just above the level of the modern Salihli plain, up to several hundred metres higher further south, forming erosional caps to underlying conglomeratic facies (for instance, near Sart). This upper unit consists of poorly sorted, coarse conglomerates, with mainly well-rounded clasts. Five separate fan lobes are recognised in the area mapped (Fig. 8; Purvis and Robertson, in press). These sediments were again supplied through discontinuities in the footwall of the presently low-angle normal fault, as with the Muddy-to-Sandy Alluvial Facies Association.

The conglomerates of this Coarse Alluvial Facies Association are only weakly indurated and vary between clast- and matrix-supported. Clasts, up to 25 cm across, are set within a fine sandy matrix, and exhibit excellent imbrication, yielding abundant palaeocurrent evidence (Fig. 8; Purvis and Robertson, in press). Clast lithologies are similar to those within the underlying conglomerates, with the addition of numerous recycled clasts.

The youngest sedimentation of the Alaşehir Graben is dominated by an axial-fluvial system comprising the modern Gediz and Alaşehir Rivers. Minor alluvial fans are being shed laterally from exhumed Neogene sediments (e.g., Paton, 1992). Lateral drainage continues to be controlled by streams issuing from the pre-existing breaks in the footwall of the presently low-angle normal fault bounding the metamorphic basement.

3.3. North-margin sediments

The northern margin of the Alaşehir Graben exposes metamorphic rocks, mainly marble, schist and quartzite, Neogene/Quaternary sediments and local calc-alkaline igneous rocks (Fig. 8). The stratigraphy and structure of this margin were poorly understood in part due to extensive cover of Quaternary alluvium. Regional mapping has either shown the contact between the Menderes Metamorphic rocks and the adjacent sediment as a depositional contact (Emre, 1996; Yılmaz et al., 2000), or as a high-angle normal fault (Westaway, 1990; Paton, 1992; Cohen et al., 1995; Kocyiğit et al., 1999). Previous work has focused on a small area north of Salihli on the banks of the Gediz River, which incises the sedimentary succession (see Westaway et al., 2004). According to Emre (1996), lacustrine and fluvial sediments, up to 400 m thick, of inferred late Miocene to Plio–Quaternary age, lie directly on the Menderes metamorphic rocks above a gentle erosional surface. By contrast, Yusufoğlu (1996) mapped a high-angle normal fault termed the Kırdamları Fault between the Menderes metamorphic rocks and the Neogene sediments It is possible that this discrepancy results from not distinguishing older layer-cake (stacked) sediments that are commonly faulted from inset river terraces that are relatively young and commonly show less evidence of faulting. More work on this area near Salihli is clearly needed (see Westaway et al., 2004).

During this study we focused instead on an exceptionally well-exposed area north of Alaşehir, around Toygarı (Fig. 8). Previous regional mapping suggested that the contact between the metamorphic basement and cover is locally faulted (Cohen et al., 1995; Dart et al., 1995) and was mapped as the Karataş Fault by Koçyiğit et al., (1999). This fault is not well exposed due to concealment by young sediments, including slope wash, but the existence of a major fault in this area is supported by the steep profile of the Gediz River (see Fig. 20 of Westaway et al., 2004). We have no doubt that this area is transected by important high-angle faults, as shown in Fig. 8. However, our detailed mapping indicates the presence of an irregular unconformity, rather than a fault, between the Menderes Massif (locally marble) and the overlying late Cenozoic sedimentary succession. A ductile foliation within the metamorphic basement dips southwards at a shallow angle (10–20°) beneath the sediments (Fig. 8). Extensional lineations dip towards the southwest, with shear-sense indicators showing local top-to-the-northeast displacement. The metamorphic rocks directly underlying these sediments are brecciated, with clasts up to several centimetres in size. These clasts preserve an interlocking jig-saw-type fabric suggesting that they are of tectonic rather than sedimentary origin. We thus infer that sediments were initially deposited unconformably along the present northern margin of the Alaşehir Graben and that both the basement and cover were later cut by an array of high-angle faults. Locally (due north of Toygarı; Fig. 8), the contact with the Menderes basement swings sharply northwards but the contact with the overlying sediments remains depositional. This feature is interpreted as a local palaeo-valley that is ~100 m wide and up to 10 m deep, incised into this metamorphic basement.

The lowest part of this north-margin succession is dominated by trough cross-bedded carbonate-rich siltstones and sandstones, which are heavily bioturbated, with rootlets and plant fragments. These sediments coarsen upwards and pass into angular, poorly sorted conglomerates, consisting almost entirely of marble clasts, up to 6–8 cm in size. These clasts are self-supported and arranged in beds up to 60 cm thick. Foreset geometries, cross bedding and clast imbrication indicate persistent southwestward palaeoflow (Fig. 8; Purvis and Robertson, in press). These sediments are cut by SW-dipping high-angle normal faults, as also mapped by Cohen et al. (1995) and Koçyiğit et al. (1999). The offsets of these faults are estimated to be <100 m, considerably less than at the southern margin of the graben.

We therefore propose that the northern margin of the Alaşehir Graben originated as part of the same zone of ductile extension that we infer to underlie the NE–SW-trending basins further north including the Selendi and Gördes Basins. This surface was later eroded and then unconformably overlain by late Cenozoic sediments. The sediment cover and the adjacent metamorphic basement were later cut by southward dipping high-angle extensional faults, probably correlating with the Plio–Quaternary high-angle faults that cut the southern margin of the graben.

3.4. Age control

Unlike in the Selendi and Gördes Basins, there has been no local volcanism coeval with the deposition of the south margin sequence, and so little potential for isotopic dating. Age determination is thus at present largely dependent on the study of sporomorphs and mammalian biostratigraphy. However, a notable feature of the northern margin is a circular outcrop of pink andesite (1 km^2) near Toygarı, interpreted as a small lava flow, although relationships with the adjacent sedimentary units are not clear (Fig. 8). This lava overlies the metamorphic basement, but pre-dates the overlying sediments (Purvis, 1998; Koçyiğit et al., 1999; Yılmaz et al., 2000). Recent Ar–Ar dating has yielded biotite ages of 14.65 ± 0.06 to 16.08 ± 0.91 Ma for this igneous rock (Purvis et al., in press), confirming an earlier K–Ar date of 14.8 ± 0.4 Ma

(Ercan et al., 1997). These results therefore indicate that the sediments overlying this igneous body are no older than mid-Miocene. However, the temporal correlation with the sedimentary succession of the southern margin of the Alaşehir Graben is debatable as there is no continuity of exposure across the graben floor. In our opinion (and that of Koçyiğit et al., 1999), the north margin sediments near Toygarı are likely to correlate with the Coarse Alluvial Fan Facies Association (i.e., the uppermost part of the south margin succession).

The ages of the sediments of the Alaşehir Graben have been and remain the subject of major controversy (see, e.g., Bozkurt, 2003, for a recent review). For instance, Seyitoğlu and Scott (1992, 1996) and Seyitoğlu et al. (2002) considered the sediments at its southern margin to represent more or less continuous deposition during steady-state extension that began in late early Miocene time, whereas Yılmaz et al. (2000) and others have argued that deposition was discontinuous with the early Miocene sediments relating to deposition in N–S-striking grabens that formed earlier and independently of the Alaşehir Graben. These authors believe that a significant hiatus of up to ~7 million years' duration exits between the lacustine sediments and the overlying red alluvial fan clastics which were inferred to be of late middle Miocene or younger age.

It is now generally accepted that the oldest sediments present, our Lacustrine Facies Association (the Alaşehir Formation of others) are late early Miocene to early middle Miocene from the pollen present, known as the Eskihisar assemblage, of inferred Burdigalian to mid-Serravallian age: 20–14 Ma (Benda and Meulenkamp, 1979; Seyitoğlu and Scott, 1996; Ediger et al., 1996). Seyitoğlu and Scott (1992, 1996) believe that the overlying two units (our Muddy-to-Sandy Alluvial Facies Association and Axial-Fluvial Facies Association) are also early to middle Miocene mainly based on the occurrence of a similar sporomorph assemblage. However, it was later reported (e.g., Sarıca, 2000; Yılmaz et al., 2000) that the pollen used in this analysis was from reworked lignite, thus casting doubt on (but not disproving) an early–middle Miocene age. These authors instead argued, based on limited in situ pollen evidence (the Kızılhisar assemblage; late Miocene) and mammalian biostratigraphy that the clastic succession of these two units is mainly late Miocene–Pliocene. In addition, the uppermost unit, our Coarse Alluvial Fan Association (Sart Formation of others), is thought to be Pliocene to early Pleistocene, based on its mammalian biostratigraphy (see, e.g., Yılmaz et al., 2000) and a comparison with the Büyük Menderes Graben further south (Sarıca, 2000).

Additional evidence (e.g., from magnetostratigraphy) is still needed to clarify the age of the clastic sediments of the Alaşehir Graben. However, the facies relationships point to a depositional transition rather that a major hiatus between our early Miocene Lacustrine Facies Association and the overlying Muddy-to-Sandy Alluvial Fan Facies Association and also of continuity between this and the overlying Axial-Fluvial Facies Association. We therefore favour an early–middle Miocene age for both the lower clastic facies associations and a (?) late Miocene–Pliocene age for the uppermost clastic unit (Coarse Alluvial Fan Facies Association).

4. Discussion

It is important to establish how and when the high-grade metamorphic rocks underlying these sedimentary basins were exhumed. Some authors believe that this was by prolonged erosion from depth (e.g., Yılmaz et al., 2000; Westaway et al., 2004). However, we favour an alternative, as we propose that the contact between the metamorphic basement and the unmetamorphosed late Cenozoic sediments originated as a low-angle ductile extensional detachment, as observed in the northern Gördes Basin and the southern Selendi Basin (Figs. 2 and 3), where unmetamorphosed remnants interpreted as originating from the detached upper plate lie directly on the Menderes metamorphic rocks. We consider that the pervasive stretching affecting these rocks reflects a broad zone of ductile extension over which exhumation took place. If correct, we would predict that the ductile extension would diminish at greater depth although the restricted range of depth of exposure does not allow this to be confirmed at present within this study area. Conversely, if unroofing occurred purely by erosion, a vertical gradation between lower and higher grade metamorphic rocks, or important thrust discontinuities would be expected. An abrupt metamorphic break is indeed present between the metamorphic basement and the unmetamorphosed overlying ophiolitic melange. However, this discontinuity is unlikely to be related to thrusting during "Alpine" contractional deformation as the kinematic indicators are north directed, the opposite of the well documented southward direction of Alpine emplacement from the İzmir–Ankara suture in this region (e.g., Şengör and Yılmaz, 1981; Collins and Robertson, 1998; Okay et al., 2001). Assuming our interpretation is correct, the northern Menderes Massif, including the basement of the Selendi and Gördes basins, is characterised by down-to-the-north extension.

In contrast, the southern Menderes Massif is associated with down-to-the-south extension, as documented along the northern margin of the Büyük Menderes Graben (e.g., Bozkurt, 2000) and within the Menderes Massif further south (e.g., Bozkurt and Park, 1994; Bozkurt, 2001). Two extensional shear zones are thus interpreted as having developed with opposite vergence on both margins of an exhumed metamorphic complex (e.g., Hetzel et al., 1995a; Lips et al., 2001; Gessner et al., 2001). This situation contrasts with core complexes that exhibit a single, or predominant, extension direction (e.g., Buck, 1988; Wernicke and Axen, 1988; Malavielle and Taboada, 1991; Westaway, 1999).

In many previous studies these NE–SW-trending basins were regarded as grabens bounded by NE–SW faults (e.g., Şengör, 1987; Seyitoğlu, 1992; Seyitoğlu and Scott, 1994a). En echelon bounding faults were, for example, reported from the Demirci Basin, between the Selendi and Gördes Basins (Yılmaz et al., 2000). The basins have also been interpreted simply as the effects of post-orogenic subsidence, without active fault control, for instance as a result of sediment loading (e.g., İnci, 2002). In this interpretation, the trend of the NE–SW basins is seen as being sub-parallel to that of the Palaeogene İzmir-Ankara suture, so that sediment loading might have accentuated initial irregularities in this pre-existing structural trend (Westaway et al., 2004).

We suggest instead that the NE–SW-trending highs between these basins are long-wavelength undulations (wavelength ~30 km, amplitude >1.5 km) oriented in the direction of extension of our inferred extensional detachment surface (see Fig 7C). We envisage the intervening basins as corresponding "scoop-shaped" depressions on a similar scale. Where well exposed, as around the eastern margin of the Gördes Basin, we have observed gentle progradation of the sediments that we infer to be of early Miocene age (see earlier discussion of age constraints) over the metamorphic basement, consistent with an overall broad "scoop-shaped" basin morphology. However, we do not exclude the possibility that the margins of the individual "scoop shaped" depressions could be locally compartmentalized into faulted segments. Some later high-angle faults also cut the basin margins in some areas in view of literature reports, for example in the Demirci Basin (Yılmaz et al., 2000). Assuming regional N–S extension, such faults would be expected to be of strike-slip to oblique-slip nature.

We consider that our explanation of basin formation related to extensional exhumation of the Menderes Massif fits the regional setting. Their NE–SW-trend significantly differs from that of the Ankara–İzmir Zone and we thus believe that the origin of these basins post-dates regional alpine contractional deformation. In addition, there is no obvious reason for the existence of two independent phases of first E–W then later N–S extension in the region (cf. Yılmaz et al., 2000). A similar syn-extensional corrugated morphology (wavelength 10–20 km, amplitude 1–2 km) is known from detachment faults in other regions (Friedmann and Burbank, 1995), notably the western U.S.A. (e.g., John, 1987; Davis and Lister, 1988; Spencer and Reynolds, 1991; Miller and John, 1999). Such large-scale detachments may reflect the existence of deep-seated structural controls (such as, inherited zones of structural weakness) or may relate to isostatic uplift of the lower plate in response to tectonic denudation (Spencer, 1984).

The nature of the contact between the metamorphic rocks and the sedimentary cover can be established at a number of localities. We have observed in many areas that the lowest exposed conglomerates prograded directly onto the metamorphic basement with no intervening fault scarps. These conglomerates are lithologically similar to the conglomerates exposed at the base of the Miocene succession beneath the alluvial plain facies (e.g., near Yurtbaşı in the SE Selendi Basin; Fig. 2). Also, in many places (e.g., the SE margin of the Selendi Basin), faults either never existed, or they were eroded (degraded) before being covered by coarse sediments. A similar conclusion was reached from study of the SE margin of the Uşak-Güre Basin (Westaway et al., 2004). Where the metamorphic basement is overlain by unmetamorphosed rocks, mainly ophiolitic mélange, we infer the contact to be the actual detachment fault. However, where an unconformity with overlying sediments is exposed this contact may not represent the actual extensional detachment surface, but rather this surface was eroded, perhaps by up to tens of metres during a relatively brief interval prior to deposition.

The lower parts of the overlying alluvial sediments of both basins are gently warped, although as many studies (e.g., Ercan et al., 1978; Yılmaz et al., 2000; Westaway et al., 2004) have also noted, their upper parts are sub-horizontal. Possible explanations for dip variations include differential compaction above an irregular basement, the presence of buried faults, tilting of buried fault blocks, or magmatic activity. High-resolution seismic data and drill cores would be needed to resolve such alternatives.

Folding of similar Miocene sediments within the Uşak-Güre Basin, further southeast, was observed by us during reconnaissance and has been reported by a number of workers (e.g., Ercan et al., 1978; Seyitoğlu, 1997; Westaway et al., 2004). Bozkurt (2003) believes that the early Miocene sediments in this basin were compressionally folded about N–S axes and unconformably overlain by (unfolded) younger sediments. Seyitoğlu (1997) reported similar folding in the Selendi Basin and Yılmaz et al. (2000) reported north-directed folding in the Demirci Basin. Our observations in the Selendi and Gördes basins do not, however, require the existence of compressional folding. More work is needed to distinguish truly compressional structures from extension-related ones (e.g., drape folds and roll-over anticlines). However, our results emphasise the importance of the E–W-striking faulting that post-dates the accumulation of the Miocene sediments (e.g., in the central Selendi Basin) and we suspect that the tilting and other deformation are extension related, other than in the vicinity of the intrusions cutting the Gördes and other basins. Comparable tilting and folding have been reported elsewhere, including the South Aegean islands (e.g., Mykonos, Paros and Naxos). Avigad et al. (2001) argue that in this area NNE–SSW extension, similar to that in western Turkey, can be compatible with an overall generally N–S extensional regime.

We regard the cause of this region's abundant late early Miocene–early middle Miocene magmatism as still debatable. Calc-alkaline magmatism (mainly 21–19 Ma) is documented by tuffs, lava flows and intrusions over a wide area, especially within the Gördes and Selendi basins (Seyitoğlu et al., 1992; Yılmaz et al., 2001). It may have been triggered by regional extension and decompression melting of lithosphere/upper mantle that had undergone chemical contamination related to the pre-existing subduction of Neotethys in the region, as suggested by Yılmaz et al. (2001), although contemporaneous subduction (Okay and Satır, 2000) cannot be easily ruled out as the cause.

We have seen little evidence of strong N–S faulting to justify the view that eruptions occurred in grabens generated in an E–W extensional stress field (cf. Yılmaz et al., 2000). On the other hand, we confirm that the volcanic rocks are restricted to the basins rather than the intervening highs, suggesting that the underlying crustal structure exerted a control on magmatic plumbing. There is also insufficient evidence of strike-slip faulting bounding the Selendi and Gördes basins to justify their interpretation as strike-slip pull-apart basins with associated magmatism (see Bozkurt, 2003). We thus suspect that the magmatism may reflect regional N–S extension, with the magmatic products being channelled into the NE–SW-trending basins by deep-seated structures, possibly those that helped to localise these basins in the first place.

4.1. The Alaşehir Graben

As already noted, a key issue regarding the Alaşehir Graben concerns whether it developed during a single on-going phase of extension (as suggested by Seyitoğlu and Scott, 1992, 1996), with all the sediment deposition related to this single phase, or whether it represents multiple phases of extension separated by hiatuses (e.g., Purvis and Robertson, 1997; Kocyiğit et al., 1999; Yılmaz et al., 2000; Bozkurt, 2000, 2001, 2002). We favour a pulsed evolution of this graben for both structural and sedimentary reasons.

First, in the southeast, we observe that the presently low-angle normal fault was cut by later high-angle normal faults, indicating that a major reorganization of fault geometry has taken place. Such a reorganization may be natural consequence of how the stress field is expected to change (e.g., Westaway, 1998) and may be influenced by both tectonic (e.g., strain rate) and environmental effects (e.g., climatic variation). Secondly, we have observed a depositional hiatus, with obvious erosion in some places, between the Axial-Fluvial Facies Association and the overlying Coarse Alluvial Fan Facies Association. This marked facies change might have been either tectonically induced (e.g., by accelerated footwall uplift), or climatically controlled (e.g., by accelerated erosion and run-off during a humid period; Purvis and Robertson, 1997). One possibility is that the fault style changed to accommodate continuing extension, whereby the presently low-angle normal fault tilted from a steeper initial dip such that further slip was impeded (mainly by friction), triggering the nucleation of new high-angle faults towards the axis of the basin.

As the bounding fault rotated to near its present dip (10–20°), it was abandoned and cross-cut by high-angle faulting, soling out into still-active deeper level faults. In some other areas, including the western U.S.A., a similar change from low-angle to high-angle faulting is known to have taken place (e.g., Profett, 1986) following the onset of extension by a few million years (Friedmann and Burbank, 1995; Fedo and Miller, 1992; Fowler et al., 1995). Alternatively, Westaway (1998, 1999) proposed a dominantly climatic control and explained the change from extension on moderately steeply dipping normal faults, to extension on a new set of deeper faults, as the result of changing rates of erosion, sediment transport and sedimentation affecting the stress field, and thus facilitating the switch to slip on a new set of steeper normal faults.

We have observed that the high-angle normal faults cut the Coarse Alluvial Fan Facies Association. This contrasts with Seyitoğlu et al. (2002) who map these sediments (their Sart Formation) as being formed in the hanging wall of several major E–W-trending high-angle normal faults. Koçyiğit et al. (1999) see these faults as actual topographic fault scarps along which alluvial fans were shed during Plio–Quaternary time. However, our mapping (e.g., south of Salihli; Fig. 8) shows that the Coarse Alluvial Fan Facies Association does not relate depositionally to the E–W-striking high-angle faults but instead forms larger lobes sourced directly from the Menderes Massif through breaks in the presently low-angle normal fault. The high-angle normal faults cut and thus post-date this alluvial facies. This implies a late Pliocene–early Quaternary age for the establishment of the later phase of high-angle faulting, possible linked to the initiation of the present axial graben and its seismically active bounding faults, in agreement with Westaway (1998).

It is thus unlikely that this later phase of coarse deposition (latest Miocene (?)–Pliocene) reflects the initiation and early evolution of the Alaşehir Graben (cf. Yılmaz et al., 2000). We suspect instead that the dominant control of the later stage alluvial fan facies was a change in climate resulting in accentuated run-off and fluvial incision. A similar climatic effect was proposed by Westaway (2002) for the Quaternary evolution of the Gulf of Corinth in central Greece.

A related key question is whether this entire sedimentary fill developed within an E–W basin, or whether only its upper part, of essentially Plio–Quaternary age, formed in the Alaşehir Graben, with the underlying alluvial and lacustrine sediments forming in a pre-existing unrelated basin, as favoured by Yılmaz et al. (2000) and some others (e.g., Görür et al., 1995). In this alternative hypothesis, the early Miocene lacustine facies and possibly the lower part of the overlying alluvial sediments formed in N–S-trending basins, essentially extensions of the NE–SW-trending basins to the north. A major structural and depositional hiatus then followed before onset of clastic deposition within the E–W-trending Alaşehir Graben. The sedimentary fill of the Alaşehir Graben is reported to exhibit N–S segmentation, based on Turkish Petroleum seismic reflection profiles (Yılmaz et al., 2000). The presence of N–S fault-bounded highs and lows is also suggested by gravity modelling and electrical resistivity data (Gürer et al., 2001a,b). However, in detail, these inferred basins and highs appear to be offset westwards by ~20 km relative to the prolongation of the Selendi, Demirci and Gördes basins to the north, more than can be explained by a possible oblique component of rifting of the Alaşehir Graben. Such fault-bounded features could instead be related to an early to mid-Miocene initiation of the Alaşehir Graben for instance as transfer faults within an overall zone of N–S extension. In addition, Bozkurt (2003) discusses the existence of "trapped" remnants of the NE–SW-trending basins along the northern margin of the Alaşehir Graben (north of Salihli) and takes these deposits as evidence that the E–W grabens post-date the N–S ones. However, the large-scale NE–SW-trending corrugations that we envisage as forming the Selendi and Gördes basins could well have extended southwards beneath the present Alaşehir Graben and thus formed sinks for early–middle Miocene sediments, similar to those exposed within the more northerly basins.

Şengör (1987) and Yılmaz et al. (2000) considered several geometric settings that would permit the development of N–S-trending grabens within the hanging wall of an overall zone of N–S crustal extension. However, the Selendi and Gördes Basins and any possible prolongations southwards beneath the Alaşehir Graben in our interpretation are developed in the footwall rather that the hanging wall of the zone of N–S extension.

Towards the Aegean coast, red clastics of inferred late middle–late Miocene age, located within the Çubukludağ Graben, near İzmir (Fig. 1), are reported to have been shed from originally N–S-striking faults, again apparently supporting E–W rather than N–S regional extension prior to the Pliocene (Genç et al., 2001). However, these N–S-striking faults could also be interpreted as large transverse faults within an overall N–S extension system. Elsewhere, in the prolongation of the Büyük Menderes Graben to the coastal area near Söke and Kuşadası (Fig. 1), pre-Pliocene red clastic sediments and related N–S-striking faults are reported to be transected by later E–W-striking faults (Gürer et al., 2001a,b). Such E–W-striking faults could be equivalent to the late-stage high-angle E–W-striking faults that cut the sedimentary fill of both the Büyük Menderes and Alaşehir Grabens.

Taking all the evidence into account, we favour the N–S extension hypothesis for the following main reasons.

Firstly, the presence within the Lacustrine Facies Association of conglomerates, interpreted as coarse alluvial fans, supports a genetic link with the alluvial fans of the overlying Muddy-to-Sandy Alluvial Facies Association. A major hiatus at any stage between the start of deposition of the Lacustrine Facies Association is thus unlikely. Secondly, the limited palaeocurrent data from the Lacustrine Facies Association and the much more extensive palaeocurrent evidence from the Muddy-to-Sandy Alluvial Facies Association (Fig. 8; Purvis and Robertson, in press) indicate derivation from the south with no evidence of any derivation from possible N–S-trending basin margins, as would be implied by the existence of N–S-trending fault-controlled basins in this area. Thirdly, although some authors believe that a phase of regional compressional folding (NNE–SSW- and E–W-trending) affected the Alaşehir Graben in late Miocene time (e.g., Koçyiğit et al., 1999; Yılmaz et al., 2000), rare folds observed by us in the graben fill (at the southern margin of the Alaşehir Graben only) could instead be interpreted as extension-related features (e.g., Purvis, 1998; Seyitoğlu et al., 2000; Sözbilir, 2002). It is, therefore, unlikely that the deposition of the Coarse Alluvial Fan Facies Association is related to regional uplift accompanying compressional deformation.

We thus infer that our Muddy-to-Sandy Alluvial Facies Association and our Axial-Fluvial Facies Association were deposited during a single phase of north–south extension which probably lasted from the early Miocene to the (?) late Miocene (Fig. 13). During this time, extension occurred on the normal fault bounding the northern margin the Menderes Massif, which initiated at a relatively steep dip (Fig. 13a) and tilted to progressively lower-angle dips as extension proceeded, leading to the strong back-tilting of the oldest sediments that were deposited in its hanging wall (Fig. 13b). After deposition of the entire clastic fill exposed on the southern margin of the graben, these sediments were cut by a later phase of high-angle faults that also cut the previously back-tilted normal fault bounding the northern margin of the basin, as observed in the south-east of the area (Fig. 13c). This faulting is inferred to be of Plio–Quaternary age as it cuts all the exposed clastic sedimentary units of the southern margin of the Alaşehir Graben. It is also likely to be approximately contemporaneous with the faulting of the late Cenozoic sediments at its northern margin (e.g., in the Toygarı area; Fig. 8), because these faults also cut the entire preserved sedimentary succession, which is inferred to be of at least partly Pliocene age (see earlier discussions of age controls). It may also possibly be contemporaneous with the E–W-trending normal faults cutting the Selendi and Gördes Basins, as discussed earlier.

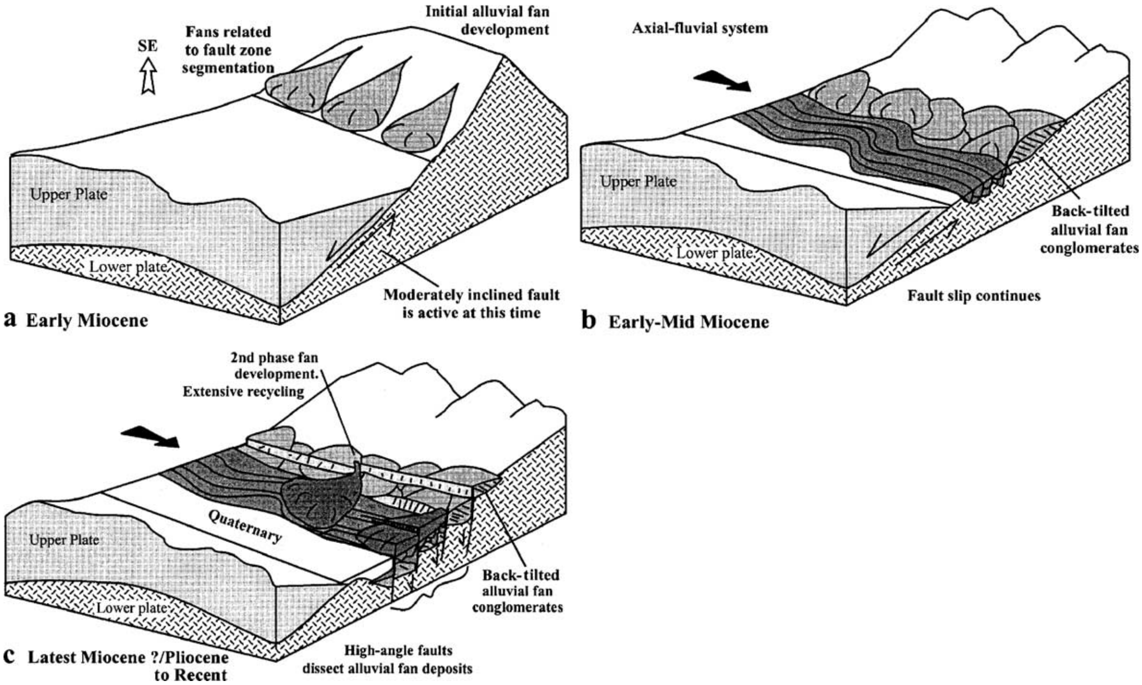

Fig. 13. Block diagrams illustrating the inferred tectonic evolution of the Alaşehir Graben. (a) Its initial phase of extension, early Miocene, showing normal faulting and related alluvial-lacustrine deposition. (b) Later during the same phase of extension, in the early-mid Miocene, showing further extension on the same normal fault set and development of an axial fluvial system. (c) From the (?) Pliocene, a new phase of high-angle normal faulting dissected earlier sediments and the back-tilted normal fault plane that was active during the earlier phase, combined with a second phase of alluvial fan development.

Fig. 14 illustrates the inferred geometrical relationship in the early Miocene between the low-angle zone of detachment that we infer to underlie the Gördes and Selendi basins and the normal fault zone that we suggest controlled extension across the Alaşehir Graben from the early Miocene. In this interpretation,

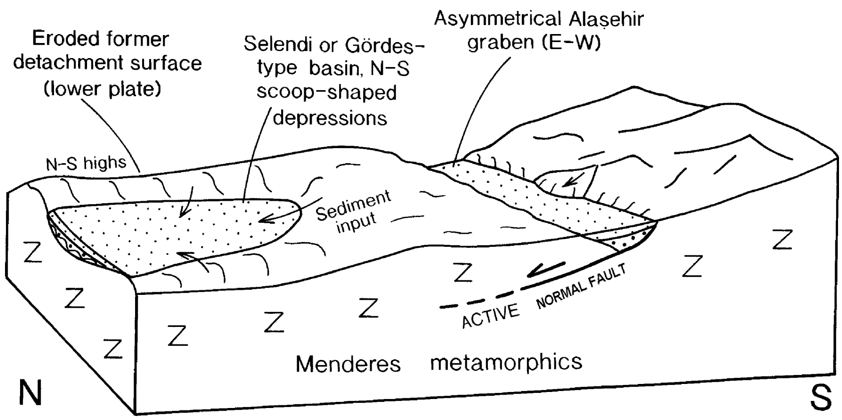

Fig. 14. Block diagram illustrating the inferred regional setting in early–middle Miocene time when the E–W Alaşehir Graben was first active in the present interpretation. The presently low-angle fault bounding the southern margin is inferred to sole out at depth to the north towards the Selendi, Demirci and Gördes Basins. Miocene sediments accumulated in depocentres within large-scale corrugations that are inferred to have formed during earlier, presumably late Oligocene, extensional unroofing of the Menderes Metamorphic Massif.

any slip on the inferred regional low-angle detachment zone was over by the time the Alaşehir Graben became active. As already noted, we infer that the early Miocene and younger sediments of the Gördes and Selendi Basins are not directly extension related, but were deposited in large-scale "scoop-shaped" depressions (Fig. 15) that we attribute to the earlier low-angle normal slip. Any such phase of low-angle slip must have pre-dated the early Miocene and we thus tentatively place it in the late Oligocene.

We therefore suggest that this large region of western Turkey has experienced three phases of extension: an initial phase that we infer took place by dominantly low-angle normal faulting and ductile

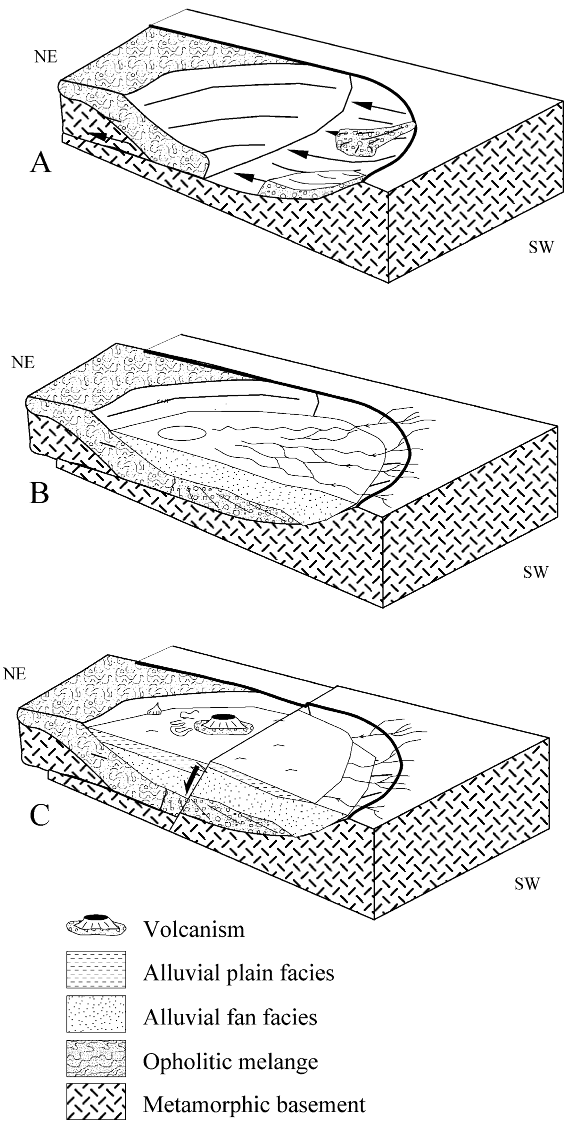

Fig. 15. Possible tectonic evolution of the Selendi Basin, assuming that the study region is underlain by an extensional detachment system, compartmentalised along strike into a series of scoop-shaped depressions, with wavelength ~30 km and amplitude ~1.5 km, one of which developed into this basin, separated by structural highs. (A) During slip on this detachment, coarse clastic deposits were shed from its footwall of this detachment, forming the earliest part of the late Cenozoic succession. (B) Fluvial environments later developed. (C) At a later stage, lacustrine/tuffaceous environments developed and the region was later cut through by minor normal faulting. See text for discussion.

shear in the (?) late Oligocene, a second phase that occurred from the early Miocene to the (?) late Miocene–early Pliocene and a final phase from the Pliocene to the present. We call this our pulsed extension hypothesis.

5. Regional controls of pulsed extension

The crustal-scale extension that we envisage to have formed the NE–SW-trending basins was possibly triggered by crustal thickening related to the final closure of Neotethys (Şengör et al., 1984). In one interpretation, the crust in western Turkey was thickened to an estimated 50–55 km (Şengör et al., 1984), followed by gravity thinning to the present average thickness of ~40 km (Makris and Stobbe, 1984; Mindevalli and Mitchell, 1989). Alternatively, Okay and Satır (2000) argue that the crust was not greatly thickened by suturing of Neotethys in western Turkey.

Assuming the second interpretation is more correct, simple gravitational "orogenic collapse" is an unlikely explanation of orogenic exhumation. An alternative trigger is gravity spreading and "roll-back" of a south-Aegean subduction zone (e.g., Robertson and Grasso, 1995; Jolivet and Patriat, 1999; Avigad and Garfunkel, 1991; Avigad et al., 2001). Following suturing by Eocene time of the Mesozoic ocean basin, known as the northerly Neotethys in Turkey and the Vardar ocean in Greece (e.g., Şengör and Yılmaz, 1981; Robertson and Dixon, 1984; Okay et al., 2001) a southerly Neotethyan ocean remained open in the south during the Oligo–Miocene, allowing the orogen to spread southwards as Africa–Eurasia convergence continued. We see this as the dominant regional control of prolonged N–S extension. At present, there is no agreement on when northward subduction of the present Hellenic subduction zone and possible roll-back began in the south Aegean. Estimates vary from early Tertiary, based on seismic tomography (Spakman et al., 1988; Taymaz et al., 1990, 1991), to late Eocene–early Oligocene based on fission-track dating (Thomson et al., 1988), to late Oligocene (~33 Ma), based on arguments concerning the timing of subduction/accretion (Kastens, 1991), to early Miocene (~26 Ma or earlier), based on knowledge of Cretan geology (Meulenkamp et al., 1988), to late Miocene (~13 Ma; Le Pichon and Anglier, 1979) or late Miocene–Pliocene (~5 Ma; Jackson and McKenzie, 1984), based on mainly theoretical arguments. However, geological evidence from Western Turkey, especially the timing of calc-alkaline magmatism, can be regarded as consistent with subduction roll-back being active in the Oligocene (Okay and Satır, 2000). If correct, thus could be seen as the driving force of inferred regional exhumation of the Menderes Massif during Oligocene time. We presume that roll-back then continued to dominate the regional tectonic setting during the Miocene and this is seen as the regional control of our second pulse of the crustal extension that initiated the Alaşehir Graben. We then interpret our third pulse of N–S extension, dominated by high-angle faulting of inferred Plio–Quaternary age, as being related to the well-documented westward tectonic escape of Anatolia during Plio–Quaternary time (e.g., McKenzie, 1978; Şengör et al., 1985; Reilinger et al., 1997).

6. Conclusions

We have developed a significant body of new field evidence, coupled with Ar–Ar dating, relating to the history of crustal extension in western Turkey. We show that two of the NE–SW-trending basins in this region, the Gördes and Selendi Basins, are unlikely to be related to a phase of regional E–W extension as in some previous models. We propose instead that these basins are the result of earlier, (?) late Oligocene, low-angle normal faulting and ductile shear that created basin-scale "scoop-shaped" depressions in which sediments later accumulated, initially during the early Miocene. These basins were later modified by relatively minor E–W-trending high-angle normal faulting of inferred Plio–Quaternary age. We interpret the Alaşehir Graben in terms of two phases of extension, an early phase lasting from the early Miocene to the (?) late Miocene or early Pliocene and a young Plio–Quaternary phase. Taking account of our earlier phase of regional extension that we infer unroofed the Menderes Massif, we thus propose a new three-phase "pulsed extension" model for western Turkey.

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