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A N–S-trending Active Extensional Structure, the Şuhut (Afyon) Graben: Commencement Age of the Extensional Neotectonic Period in the Isparta Angle, SW Turkey

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Abstract: The Şuhut graben is an 8–11-km-wide, 24-km-long, N–S-trending, active extensional structure located on the southern shoulder of the Akşehir-Afyon graben, near the apex of the outer Isparta Angle. The Şuhut graben developed on a pre-Upper Pliocene rock assemblage comprising pre-Jurassic metamorphic rocks, Jurassic–Lower Cretaceous platform carbonates, the Lower Miocene–Middle Pliocene Afyon stratovolcanic complex and a fluvio-lacustrine volcano-sedimentary sequence.

The eastern margin of the Şuhut graben is dominated by the Afyon volcanics and their well-bedded fluvio-lacustrine sedimentary cover, which is folded into a series of NNE-trending anticlines and synclines. This volcano-sedimentary sequence was deformed during a phase of WNW–ESE contraction, and is overlain with angular unconformity by nearly horizontal Plio–Quaternary graben infill. Palaeostress analyses of slip-plane data recorded in the lowest unit of the modern graben infill and on the marginal active faults indicate that the Şuhut graben has been developing as a result of ENE–WSW extension since the latest Pliocene. The extensional neotectonic period in the Isparta Angle started in the latest Pliocene.

All margins of the Şuhut graben are determined and controlled by a series of oblique-slip normal fault sets and isolated fault segments. More active faults which are capable of creating destructive earthquakes with magnitudes of Mw=6.3 and Mw=6.5 include the Ağzıkara, Güneytepe, Çobankaya and the Yarışlı faults, as in the occurrence of two devastating historical earthquakes, those of 1766 and 1862, which were seismic events with intensity of VII and X, respectively. The Yarışlı fault, however, still remains seismically inactive. In addition, the finer-grained modern graben infill is thixotropic and so these active faults and the finer-grained alluvial sediments have to be taken into account in both earthquake risk analysis and city planning design in Şuhut County.

Key Words: Isparta Angle, fold, Şuhut graben, oblique-slip normal fault, palaeostress analysis, SW Turkey

K–G Gidişli Aktif Bir Genişleme Yapısı, Şuhut (Afyon) Grabeni: Isparta Açısında Genişlemeli Yenitektonik Dönemin Başlama Yaşı, GB Türkiye

Özet: Şuhut grabeni 8–11 km genişlikte, 24 km uzunlukta, K–G gidişli, aktif bir genişleme yapısı olup, Akşehir-Afyon ana grabeninin güney omuzunda ve dış Isparta Açısı'nın kuzey uç kesiminde yer alır. Şuhut grabeni Pliyosen öncesi bir kaya topluluğu üzerinde gelişmiştir. Bu kaya topluluğu başlıca Jura öncesi yaşlı metamorfik kayalar, Jura–Alt Kretase yaşlı platform karbonatları, geç Erken Miyosen–Orta Pliyosen yaşlı Afyon volkanik karmaşığı ve aynı yaşlı akarsu-göl ortam ürünü volkano-sedimanter bir istiften oluşur.

Şuhut grabeninin doğu kenarı Afyon volkanitleri ve onun örtüsünü oluşturan göl-akarsu ürünü sedimanter bir istif ile şekillenir. İyi gelişmiş katmanlanma sunan örtü istifi KKD gidişli bir seri antiklinal ve senklinal ile deformasyona uğramıştır. Bu volkano-sedimanter istif BKB–DGD yönlü bir daralma fazıyla deformasyon geçirmiştir. Deformasyon geçirmiş volkano-sedimanter istif Pliyo–Kuvaterner yaşlı ve hemen hemen yatay konumlu modern graben dolgusu tarafından açılı uyumsuzlukla örtülür. Gerek grabenin en alt dolgusu içinde kayıd edilmiş, gerekse grabeni sınırlayan aktif kenar fayları üzerinde gelişmiş olan kayma vektörlerinin (kayma düzlemi ve kayma çizikleri) eskigerilim analizi, Şuhut grabeninin, DKD–BGB yönelimli bir genişlemenin denetiminde, en Geç Pliyosen'den beri gelişmekte olduğunu gösterir. Genişleme türündeki neotektonik dönem Isparta Açısı ve özellikle Şuhut bölgesinde, Geç Pliyosen'de başlamıştır.

Şuhut grabeninin tüm kenarları bir seri verev atımlı normal fay seti ve tekil faylar tarafından belirlenmekte ve denetlenmektedir. Çok daha aktif ve büyüklüğü Mw=6.3 ve Mw=6.4 arasında değişen yıkıcı deprem üretme potansiyeline sahip ana faylar Ağzıkara, Güneytepe, Çobankaya ve Yarışlı faylarıdır. Bu durum, özellikle Ağzıkara ve Güneytepe ana faylarından kaynaklanan ve şiddetleri VII-X arasında değişen iki yıkıcı tarihsel depremle de (1766 ve 14 Kasım 1862 depremleri) kanıtlanmıştır. Bununla birlikte, Yarışlı fayı, sismik boşluk özelliğini günümüzde de korumaktadır. Ayrıca, ince taneli güncel graben dolgusu, yüksek sıvılaşma kapasitesine sahiptir. Bu nedenle, anılan ana aktif faylar ve sıvılaşma kapasitesi yüksek olan ince taneli gevşek zemin (alüvyon) gerek deprem risk analizinde, gerekse Şuhut ilçesi gelişim planlamasında dikkate alınmalıdır.

Anahtar Sözcükler: Isparta Açısı, kıvrım, Şuhut grabeni, verev-atımlı normal fay; eskigerilim analizi, GB Türkiye

Introduction

Major active structures governing the neotectonics of Turkey and its surroundings include the dextral North Anatolian Fault System (NAFS), the sinistral East Anatolian Fault System (EAFS) and Dead Sea Fault System (DSFS), and the south Aegean-Cyprus Subduction Zone (ACSZ) (Figure 1). The NAFS and the EAFS determine the outline of the Anatolian microplate that is escaping west-southwestwards overriding the subducting oceanic lithosphere of the Eastern Mediterranean Sea. In addition to these major structures, second-order contractional and extensional fault systems and fault zones also traverse the Anatolian microplate, deforming and dividing it into a number of smaller blocks. Most of these second-order structures are strike-slip faults that splay off from the NAFS and are located in the eastern part of the Anatolian microplate. These include the Yağmurlu-Ezinepazarı Fault System (YEFS), the Central Anatolian Fault System (CAFS), the Yakapınar-Göksun Fault System (YGFS), and the Malatya-Ovacık Fault System (MOFS) (Figure 1). Two other structures, the İnönü-Eskişehir Fault System (İEFS) and the Akşehir-Simav fault system (ASFS) in the southwest part of the Anatolian microplate, are extensional (Figure 1). The İEFS is an approximately 15-km-wide, 430-km-long, NNW-trending active deformation zone composed of numerous short (0.2 km) to long (up to 30 km), closely-spaced (i.e., the interval among fault segments ranges from 100 m to 500 m), parallel to sub-parallel, southerly- and northerly-dipping normal fault segments linked to each other by a number of intervening relay ramps of varying size. It also forms a transitional boundary between areas undergoing continental extension in the south and strike-slip faulting in the north (Figure 1). The ASFS is an average 10–30-km-wide, 550-km-long and NW–SE-trending seismogenic belt within the Anatolian microplate, and is characterized by a series of grabens to horsts and their margin-boundary oblique-slip normal faults. Apart from these two major extensional structures, a number of E–W-, NW-, NE- and N–S-trending grabens and horsts occur in the southwestern part of the Anatolian microplate. Their activity is indicated by both the focal mechanism solutions of large and devastating seismic events and palaeostress analysis of slip-plane data on the fault arrays (Ergin *et al.* 1967; Ambraseys & Tchalenko 1972; Angelier *et al.* 1981; Soysal *et al* 1981; Eyidoğan & Jackson 1985; Ambraseys & Finkel 1987; Taymaz & Price 1992; Taymaz 1993; Temiz *et al.* 1998; Wright *et al.* 1999; Yılmaztürk & Burton 1999; Koçyiğit *et al.* 2000a; Özalaybey *et al.* 2000; Taymaz *et al.* 2002; Koçyiğit & Özacar 2003; Bozkurt & Sözbilir 2004; Koçyiğit 2005).

The southwestern part of the Anatolian microplate is the most rapidly deforming continental region in the world (e.g., Arpat & Bingöl 1969; Ambraseys & Tchalenko 1972; McKenzie 1972, 1978; Dumont *et al.* 1979; Koçyiğit, 1984a, 2000; Eyidoğan & Jackson 1985; Şengör 1987; Westaway 1990; Paton 1992; Seyitoğlu *et al.* 1992; Taymaz & Price 1992; Zanchi & Angelier 1993; Price & Scott 1994; Cohen *et al.* 1995; Hetzel *et al.* 1995; Koçyiğit *et al.* 1999, 2000a, b; Yusufoğlu 1998; Bozkurt 2000, 2001, 2002; Yılmaz *et al.* 2000; Koçyiğit & Özacar 2003; Bozkurt & Sözbilir 2004; Erkül *et al.* 2005; Koçyiğit 2005; Tokçaer *et al.* 2005; Aldanmaz 2006). Widely distributed and frequent shallow-focus earthquakes in the continental crust (Ambraseys & Tchalenko 1972; McKenzie 1972; Eyidoğan & Jackson 1985; Westaway 1990; Zanchi & Angelier 1993; Kalafat 1998; Yılmaztürk & Burton 1999; Koçyiğit 2000; Koçyiğit *et al.* 2000a; Ambraseys 2001; Koçyiğit & Özacar 2003) indicate that continental extension has continued in this region above the ACSZ since at least the late Pliocene (e.g., Koçyiğit *et al.* 1999, 2000a; Bozkurt 2000, 2001, 2002; Yılmaz *et al.* 2000; Koçyiğit & Özacar 2003; Koçyiğit 2005; Aldanmaz 2006). However, some critical geological problems remain disputed, as the lack of detailed field data prevents plausible explanations being universally accepted. These problems include: (a) the nature of extension in southwest Turkey: is it uni- or multi-directional?, (b) whether the evolutionary history of horst-graben system in southwest Turkey is continuous or episodic in nature?, (c) whether the active horsts and grabens only trend E–W or whether there are other active horst and graben trends?, (d) establishing when the extensional neotectonic regime was initiated in southwest Turkey?, and (e) marking the easternmost limit of the '*west Anatolian extensional neotectonic province*'. Most previous papers (Boray *et al.* 1985; Seyitoğlu & Scott 1994; Barka *et al.* 1995; Yılmaz *et al.* 2000; Alçiçek *et al.* 2005; Kelling *et al.* 2005) reported that the west Anatolian continental extension province is bounded by the western flank of the Isparta Angle to the east, and that its incipient age ranges from early Miocene to late Pliocene. However, one of major areas which should be included in the west Anatolian extensional

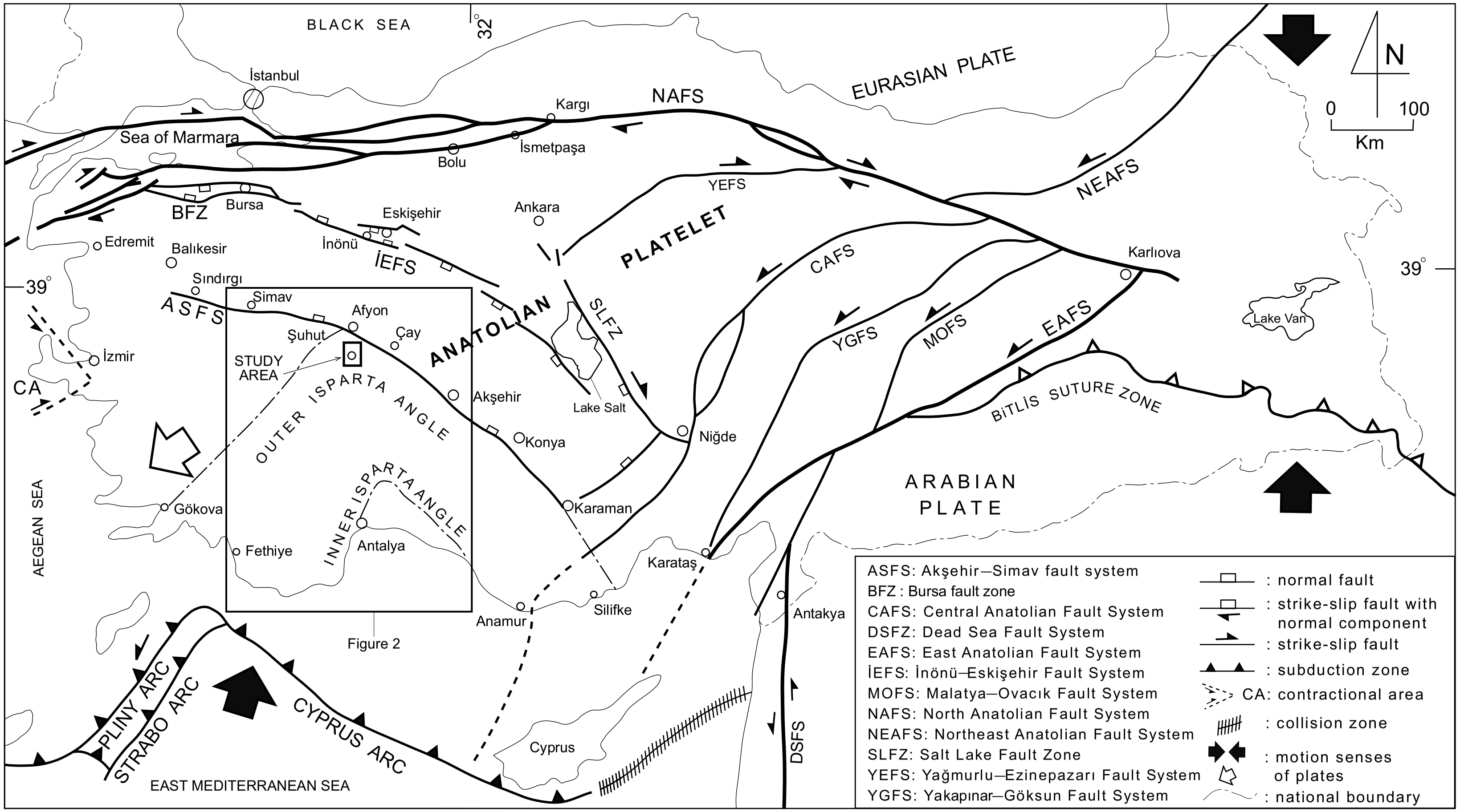


Figure 1. Map showing major active fault systems and fault zones governing neotectonics of Turkey and its neighborhood.

province is the Isparta Angle. There are two models for the type and initiation age of the neotectonic regime in the Isparta Angle: (1) the present-day configuration of the Isparta Angle is a contractional structure, and the initiation age of its contractional neotectonic regime is late Miocene (Boray *et al.* 1985; Barka *et al.* 1995; Altunel *et al.* 1999), and (2) the Isparta Angle is an extensional structure and initiation of the extensional neotectonic regime was during the latest Pliocene (Koçyiğit 1996; Glover & Robertson 1998b; Koçyiğit *et al.* 2000a; Koçyiğit & Özacar 2003; Poisson *et al.* 2003; Koçyiğit 2005).

The main aim of this paper is to present new field data from a well-defined neotectonic structure, the Şuhut graben, and to discuss solutions of the problems mentioned above in the light of both newly-obtained field data and literature. All these problems and their solutions are illustrated in the Isparta Angle, which is therefore one of the critical areas in which to seek more reliable data. Also, as the evolutionary history of the horst-graben system in southwest Turkey is episodic (e.g., Koçyiğit *et al.* 1999; Bozkurt & Sözbilir 2004; Bozkurt & Rojay 2005), most older contractional and *phase-I* extensional structures (older reverse and normal faults) were reactivated but deformed in a different way during the extensional neotectonic regime (*phase-II* extension). For this reason, the Palaeogene and early Neogene geological outline of the Isparta Angle will, first of all, be presented below.

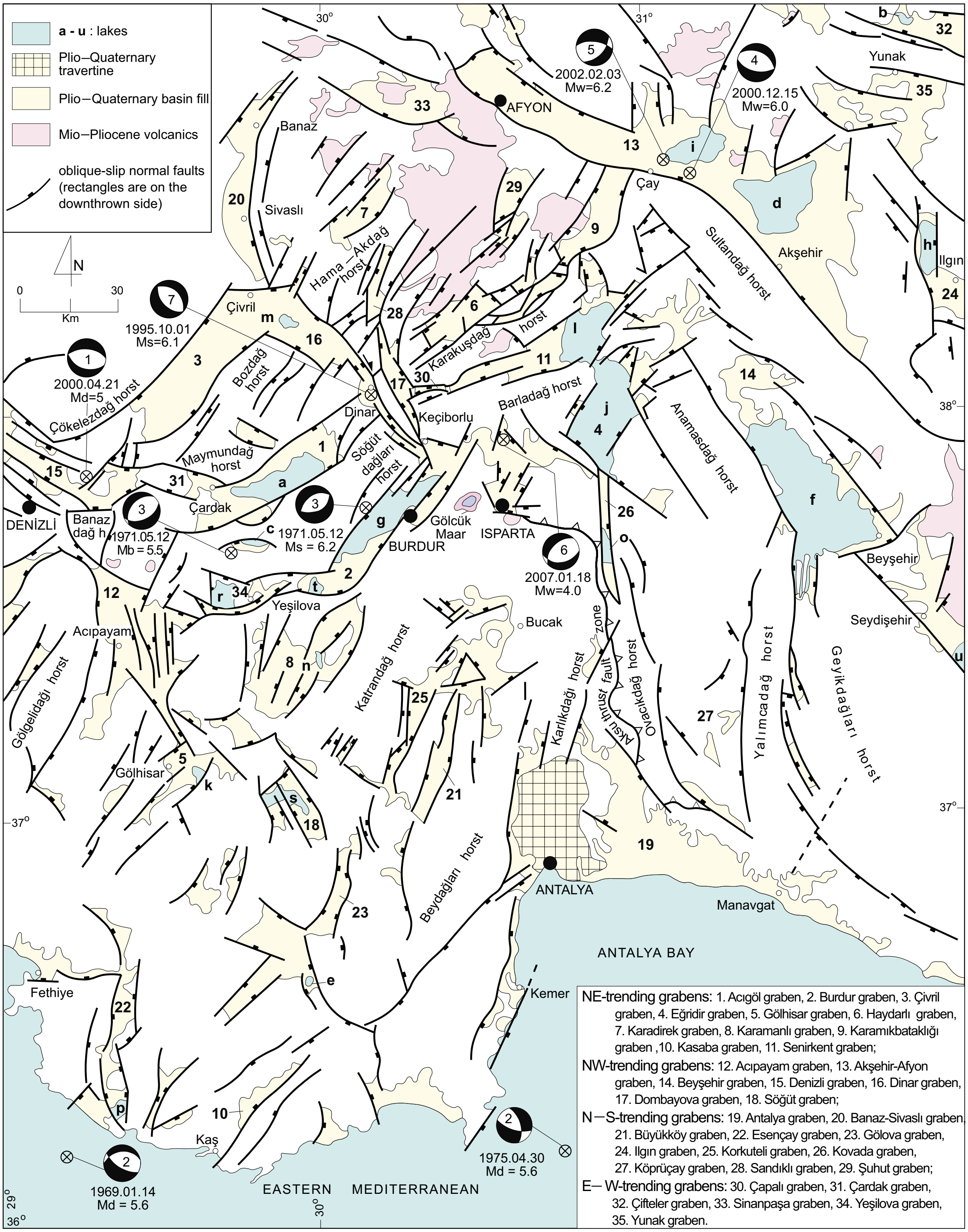
Isparta Angle

The present-day configuration of the Isparta Angle is a reverse V-shaped morphotectonic structure located north of Antalya Bay (Figure 1). Originally defined and reported by Blumenthal (1951, 1963), the first attempt to explain its origin was made by Dumont (1976). Kelling *et al.* (2005) defined the Isparta Angle as a regional embayment into the Menderes-Tauride microcontinental units that separated the Taurides into discrete eastern and western crustal fragments. However, its origin remains controversial, although most authors accept it as a palaeotectonic structure originating from the northward curvature of the originally E–W-trending Tauride orogenic belt due to nappe emplacement and related clockwise and anti-clockwise rotations in the early Paleocene to early Pliocene (Poisson 1977; Akay & Uysal 1985; Kissel & Poisson 1987; Kissel *et al* 1993; Glover & Robertson 1998a; Piper *et al.* 2002; Poisson *et al.* 2003).

The complexity of the Isparta Angle is indicated by the regional overthrusting of nappes, of dissimilar age and origin, on to an extensive Menderes-Tauride Platform or microcontinent that rifted from the Gondwanan (northernmost African) margin and was later accreted to Eurasia as the intervening Tethys closed (Dercourt *et al.* 1993). The para-autochthonous Menderes-Tauride platform consists of a thick, predominantly carbonate, Mesozoic sequence overlain by Upper Cretaceous shallow marine to pelagic sediments and Lower Eocene nummulitic limestone, and flysch that includes ophiolitic detritus related to nappe emplacement (Gutnic *et al.* 1979; Koçyiğit 1983, 1984b). This approximately E–W-trending and curvilinear submarine shelf or carbonate platform was flanked north and south by Mesozoic oceanic crust known as the northern Neotethys and the southern Neotethys, respectively (Şengör & Yılmaz 1981; Robertson 2002). Closure of the ocean north of the Menderes-Tauride microcontinent, with the formation of both the İzmir-Ankara-Erzincan and the Inner Tauride sutures between latest Cretaceous and Early Eocene time, is revealed by the presence of a regional angular unconformity and an overlying Lower Eocene polygenetic basal conglomerate composed of clasts drived directly from both the underlying pre-Eocene metamorphics of the Menderes Massif and the ophiolitic rocks (Kaya 1972; Okay 1984; Baş 1986; Koçyiğit *et al.* 1991). After complete ocean closure and continent-continent collision along the two sub-branches (the İzmir-Ankara and the Inner Tauride oceans) of the northern Neotethys, the Lycian and the Beyşehir-Hoyran ophiolitic nappes developed and were thrust southward into their present position (Özgül 1976; Poisson 1977; Gutnic *et al.* 1979; Koçyiğit 1983, 1984b; Waldron 1984; Flecker *et al.* 2005). The Beyşehir-Hoyran nappes contain ophiolitic rocks and both shallow and deep marine tectonic units of Permian–Cretaceous age which originated in the Inner Tauride oceanic setting and were thrust south-southwestward on to the central Tauride platform (eastern flank of the Isparta Angle) in Campanian and late Lutetian times. The Lycian Nappes are similar to the Beyşehir-Hoyran nappes in origin and composition and were thrust south-southeastward onto the western Tauride platform (western flank of the Isparta Angle) in two stages during the late Oligocene and late Langhian. However, closure of the southern Neotethys, initated in the latest Cretaceous in the east, still continues in the modern Eastern Mediterranean Sea. This long-term and diachronous closure history of the southern Neotethys has been controlled by two major palaeogeographic features, the Arabian promontory in the east and the Isparta Angle in the west (Kelling *et al.* 2005). Among the diagnostic tectonic units that relate to the early closure history of the southern Neotethys are the Antalya Nappes (AN). They are located near the core of the Isparta Angle and comprise a complex imbrication of Palaeozoic (Ordovician–Permian), Mesozoic and Lower Tertiary sediments and igneous rocks. They originated in southern Neotethys and were first thrust northwards onto the autochthonous Menderes-Tauride platform during the late Early Paleocene (Uysal *et al.* 1980). Episodic thrusting of the Antalya Nappes then continued in varying directions until the late Pliocene (Lamotte *et al.* 1995; Poisson *et al.* 2003), and as a result a range of vergences such as SE, SW, W and S is seen. These include the latest contractional event and its related structure, the Aksu contractional phase and the Aksu thrust zone, located near the core of the Isparta Angle (Figure 2).

As a consequence of the diachronous closure history of both the northern and southern Neotethys, the latest Palaeogene (Oligocene) and the Neogene sedimentary basins developed within and bordering the Isparta Angle are classified into two categories: (1) subductional to locally syn-collisional southern basins and (2) post-collisional northern basins. The first group of basins, from west to east, includes the Kasaba-Karakuş-Darıören, Aksu, Köprüçay and Manavgat basins. The Kasaba-Karakuş-Darıören basin, on the western flank of the Isparta Angle, trends approximately northeast, sub-parallel to the front of the Lycian Nappes. The Aksu and the Köprüçay basins, in the core of the Isparta Angle, trend approximately N–S, while the Manavgat basin, on the eastern flank of the Isparta Angle, trends northwest. In general, these basins contain a marine sedimentary fill over 2 km thick that ranges from Aquitanian to early Pliocene in age and consists mostly of boulder-block conglomerate and a flyshoidal facies of shale, siltstone and turbiditic sandstone alternating with patch reef intercalations (Akay & Uysal 1985; Karabıyıkoğlu *et al.* 2005). The Kasaba-Karakuş-Darıören basin in the west and the adjacent Aksu basin underwent flexural subsidence linked to the final (Langhian) south-southeastward thrusting of the Lycian Nappes (Kelling *et al.* 2005), although to the east of Isparta Angle, major south-southwestward emplacement of thrust sheets ended in the late Eocene. Both the N–S-trending Aksu and Köprüçay basins developed as half grabens controlled by extensional faults such as the Kırkkavak and Akbelenli faults (Blumenthal 1951; Dumont & Kerey 1975; Koçyiğit *et al.* 1997). During the mid-Miocene the Köprüçay and the Manavgat basins subsided in response to regional northward subduction and slab retreat of a remnant of the southern Neotethys ocean (Kelling *et al.* 2005). Later, the Aksu basin and, to some extent, the Köprüçay basin were affected by late Miocene contraction (Aksu Phase) that locally continued until the mid-Pliocene (Poisson *et al.* 2003). In the latest Pliocene, the last compression and its related structures were replaced by crustal extension and related structures, marking the start of the neotectonic period (Koçyiğit *et al.* 1997; Glover & Robertson 1998b; Koçyiğit *et al.* 2000; Koçyiğit & Özacar 2003; Poisson *et al.* 2003).

The post-collisional basins are located within the northern half of the Isparta Angle. They include the axial Tauride molasse basin, the Çameli, Acıpayam, Karamanlı, Burdur, Isparta, Senirkent, Dinar, Dombayova-Sandıklı, Karadirek, Sinanpaşa, Haydarlı-Karaadilli, Gelendost, Beyşehir-Yarıkkaya, Şuhut basins and the Akşehir-Afyon basin (Ercan *et al.* 1978; Koçyiğit 1984a, b; Boray *et al.* 1985; Karaman 1986; Şenel *et al.* 1989; Price & Scott 1991; Yağmurlu 1991; Akgün & Akyol 1992; Koçyiğit *et al.* 2000, 2001; Koçyiğit & Özacar 2003; Alçiçek *et al.* 2005). At present, discrete and variably sized outcrops of a molasse sequence of Oligocene–Aquitanian age are exposed between Andırın (Kahramanmaraş) in the east and Kale-Tavas (Denizli) in the west along the axial line of Tauride orogenic belt. The sequence consists of a 2.5-km-thick shallow-marine to terrestrial, unsorted, thick-bedded (60 cm – 6 m) to massive boulder-block conglomerate with patch reef intercalations. This molasse sequence was accumulated in an approximately E–W-trending linear to curvilinear trough or foreland basin developed in front of the northerly-derived ophiolitic nappes, that came from the northern Neotethys and its sub-branch (the Inner Tauride ocean) and were then thrust on to the Menderes-Tauride microcontinent (Koçyiğit 1981,1983,1984b; Öztürk 1982; Demirtaşlı *et al.* 1984; Tekeli *et al.* 1984; Yetiş 1987; Okay 1989;



a. Acıgöl, b. Lake Ak, c. Akgöl, d. Lake Akşehir, e. Lake Avlan, f. Lake Beyşehir, g. Lake Burdur, h. Lake Çavuş, i. Lake Eber, j. Lake Eğirdir, k. Lake Gölhisar, l. Lake Hoyran, m. Lake Işıklı, n. Lake Karataş, o. Lake Kovada, p. Lake Ova, r. Lake Salda, s. Lake Söğüt, t. Lake Yarış, u. Lake Suğla. Numbers in focal mechanism solution diagrams of earthquakes refer to: 1. Kalafat (1998), 2. Kocaefe (1982), 3. Taymaz & Price (1992), 4. Taymaz & Tan (2001), 5. Taymaz *et al.* (2002), 6. USGS – NEIC (2007), 7. Wright *et al.* (1999).

Figure 2. Neotectonic map of the Isparta Angle.

Yılmaz *et al.* 2000; Akgün & Sözbilir 2001; Sözbilir 2002; Jaffey & Robertson 2005). The molasse sequence generally unconformably overlies the northerly-derived ophiolitic nappes (Beyşehir-Hoyran Nappes and Lycian Nappes), the Triassic–Upper Lutetian sedimentary sequence and the pre-Triassic basement rocks. This reveals that the latest emplacement age of the northerly-derived nappes north of the Tauride axial line is late Eocene. However it also locally shows both transitional and reverse-faulted contact relationships with the pre-Oligocene rocks.

All the remaining basins in the northern half of the Isparta Angle are terrestrial and characterized by fluvio-lacustrine fill of varying age ranges (from latest early Miocene to mid-Pliocene), thickness (0.5–3 km) and internal facies. The fluvio-lacustrine basin fill with the coal seam intercalations was accumulated on the pre-Aquitanian rocks in the NE-, NW-, N–S- and E–W-trending grabens (Koçyiğit *et al.* 2000; Alçiçek *et al.* 2005; Kelling *et al.* 2005; Koçyiğit 2005). Fluvio-lacustrine sedimentation was also accompanied by calc-alkaline volcanism in places (Keller & Villari 1972; Becker-Platen *et al.* 1977; Besang *et al.* 1977; Koçyiğit 1981, 1983, 1984a; Ercan *et al.* 1985; Çevikbaş *et al.* 1988; Erkül *et al.* 2005; Aldanmaz 2006). The Miocene crustal extension that locally continued into the mid-Pliocene was recorded within the suitable lithofacies (e.g., mudstone to claystone) of the basin fill and on the margin-bounding faults. Their palaeostress analyses indicate an extensional tectonic regime that influenced sedimentation of the basin fill (Dumont *et al.* 1979; Koçyiğit 2005). The post-collisional basins and their fills were diachronously deformed by folding and reverse or strike-slip faulting between late Miocene and mid-Pliocene time (Glover & Robertson 1998a; Koçyiğit *et al.* 1999; Kaya *et al.* 2004; Koçyiğit 2005). In the frame of regional geodynamic processes, the late Miocene–mid Pliocene diachronous contractional events can be linked to the most distant reflections of the final continent-continent collision of the Eurasian and Arabian plates and consequent formation of the Bitlis Suture in southeastern Turkey in the late mid-Miocene Figure 1). In the latest Pliocene, a second phase of crustal extension occurred, here termed the *extensional neotectonic period*, caused by a combination of several geodynamic processes, such as the emergence and west-southwestward motion of the Anatolian platelet, the northward subduction of the southern Neotethys and related slab retreat (Glover & Robertson 1998a; Koçyiğit *et al.* 1999; Kelling *et al.* 2005). This second phase of crustal extension led both to reactivation of most of the older structures and the formation of new extensional faults. These divided the Isparta Angle and its surroundings into a number of horsts and grabens that shape the present-day configuration of the Isparta Angle Horst-Graben System (IAHGS). The IAHGS is characterized by NE-, NW-, N–S- and rarely E–W-trending horsts and grabens and their margin-bounding oblique-slip normal faults. Major grabens and horsts trending in different directions are listed and labelled in Figure 2. Most grabens have two infills, separated by an intervening angular unconformity. These are the exhumed and dissected older fill of Miocene–mid Pliocene age, deformed by folding, reverse faulting and strike-slip faulting, and the nearly horizontal Plio–Quaternary modern fill. These superimposed basins and the very distinct inversion in the geochemical composition of the Plio–Quaternary volcanism (from calc-alkaline to alkaline) that accompanied the modern graben formation, imply an episodic evolutionary history rather than continuous evolution for the IAHGS, similar to the west Anatolian Horst-Graben System (e.g., Keller 1974; Richardson-Bunbury 1996; Nemec *et al.* 1998; Temiz *et al.* 1997; Koçyiğit *et al.* 1999, 2000; Bozkurt & Sözbilir 2004; Beccaletto & Steiner 2005; Bozkurt & Rojay 2005; Erkül *et al.* 2005; Tokcaer *et al.* 2005).

Stratigraphic Outline of the Study Area

Based on their type and age, the rocks exposed in the study area and its neighborhood are subdivided into three categories: (1) older rocks, (2) pre-modern graben fill units, and (3) modern graben fill units (Figure 3). Older rocks consist of pre-Jurassic low-grade metamorphic basement rocks, Jurassic–Lower Cretaceous platform carbonates, Upper Cretaceous ophiolitic mélange, Eocene flysch and volcanic rocks, and the Oligocene molasse. The basement rocks, exposed at the northeastern corner of the study area, consist mainly of quartz-mica-chlorite schist, slate, quartzite and marble (Figure 4). The Jurassic–Lower Cretaceous platform carbonates crop out along the southern margin of the Şuhut graben and consist of thick-bedded to massive and recrystallized shallow marine limestones. Both of these older rock assemblages are overlain with angular unconformity by both the pre-modern graben and modern graben fill

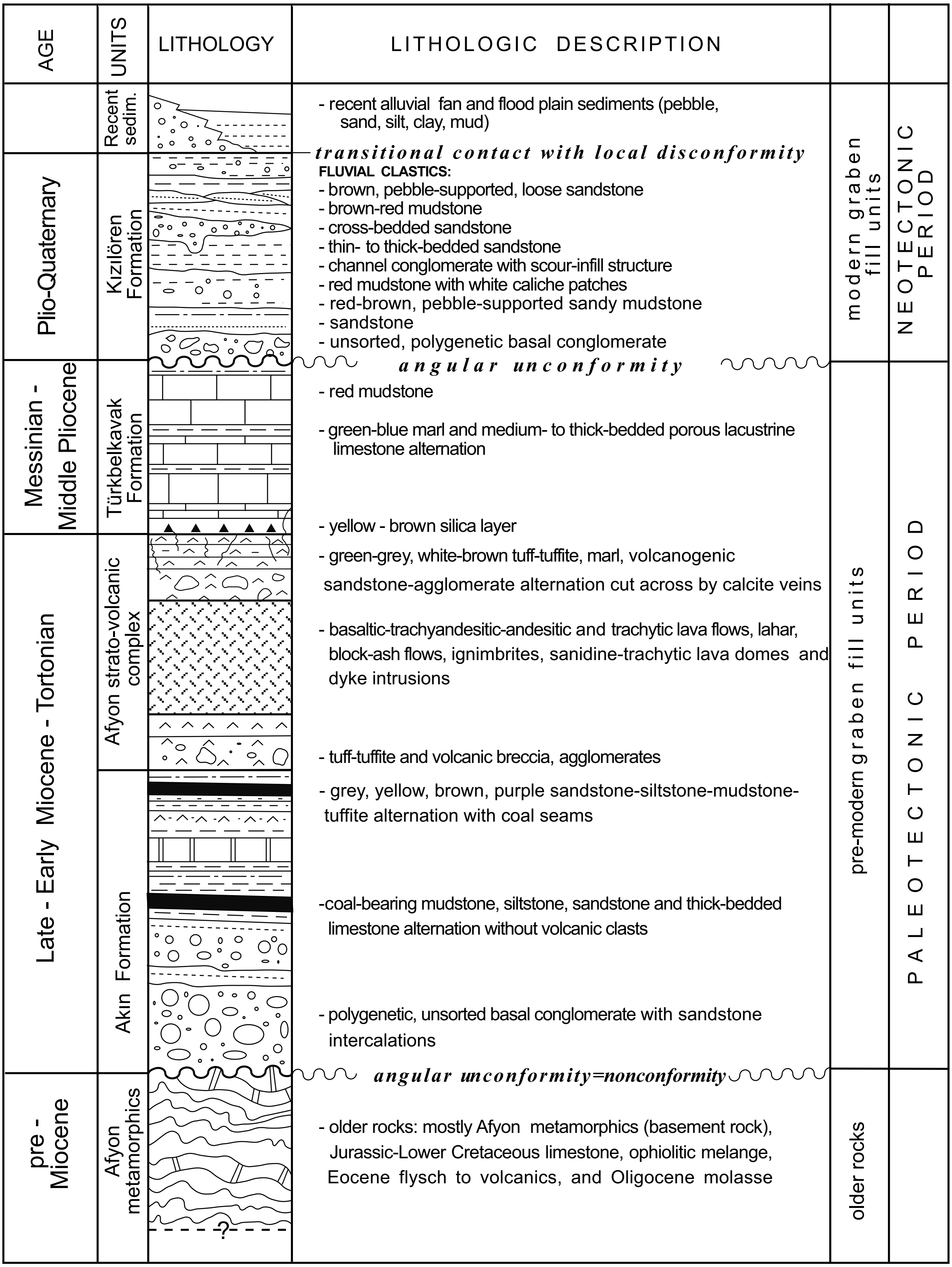


Figure 3. Generalized stratigraphical columnar section of the Şuhut graben.

units. Detailed description of older basement rocks and the platform carbonates is beyond the scope of this paper. However, the pre-modern graben fill units, which are the youngest palaeotectonic units, are described in more detail below to emphasize the distinction between the rock sequences, deformation patterns and related structures of both the palaeotectonic and neotectonic periods.

Pre-Modern Graben Fill Units

A volcano-sedimentary sequence, over 1 km in thickness, crops out over a very broad area in both the western and eastern sides of the Şuhut graben, a N–S-trending extensional neotectonic structure located near the apex of the Isparta Angle (Figures 1, 3 & 4). The volcano-sedimentary sequence, overlain with angular unconformity by the Plio–Quaternary modern graben infill, is subdivided into three rock-sratigraphic units. These are, from bottom to top, the Akın Formation, the Afyon volcanic complex and the Türkbelkavak Formation (Figure 3).

The Akın Formation has a grey to purple, unsorted, polygenetic basal conglomerate immediately above the unconformity, overlain by an alternation of various lithologies such as conglomeratic sandstone, sandstone, siltstone, mudstone, coal seam-bearing claystone, marl and thick-bedded to massive limestone. This 10–60-m-thick basal sequence consists mainly of well-rounded quartz, quartzite, schist and marble clasts, derived directly from the underlying older metamorphic rocks, in a sandy matrix. Grains or pebbles of volcanic origin are absent, so this sedimentation precedes the volcanic activity in the apex of the Isparta Angle. Based on the rich mammalian fossil content (e.g., *Byzantina cariensis*, *Pliospalax* sp., *Amphilagus fontannesi, Alloptox* cf. *gobiensis)*, a latest Early Miocene age (MN-5–7) is assigned to this basal sequence (Sickenberg *et al*. 1975; Saraç 2003).

The basal sedimentary sequence is succeeded by a volcanic rock assemblage of alkaline to calc-alkaline composition over 1 km thick, the Afyon strato-volcanic complex, composed of tuffite, tuff, volcanic breccia, ignimbrite, block-ash flows, lahar, basaltic-andesitic-trachyandesitic- and trachytic lavas, and domes cut by a series of basaltic dykes (Keller & Villari 1972; Besang *et al*. 1977; Keller 1983; Çevikbaş *et al*. 1988; Aydar *et al*. 1996; Akal 2003). It is overlain conformably by a pyroclastic sequence consisting of volcanic breccia, tuff-tuffite, volcanic material-rich sandstone and marl cut by a series of isolated to conjugate veins containing calcite of hydrothermal origin (Figure 3). Based on both the common principles of relative ages and radiometric age determinations (14.75±0.3 Ma and 8±0.6 Ma) from different horizons of the Afyon strato-volcanic complex (Keller & Villari 1972; Çevikbaş *et al*. 1988; Akal 2003), a Middle–Late Miocene age is assigned to it (Sickenberg *et al.* 1975; Becker-Platen 1977; Besang *et al.* 1977; Koçyiğit *et al*. 2001; Saraç 2003). In addition, based on its petrographical and geochemical composition, the Afyon strato-volcanic complex can be correlated with other widespread Miocene volcanic rocks in western Anatolia (e.g., Erkül *et al*. 2005; Aldanmaz 2006).

The uppermost tuffaceous and veined unit of the Afyon strato-volcanic complex is overlain conformably by the yellow-brown siliceous basal unit of the Türkbelkavak Formation. This easily-recognizable key horizon is succeeded by alternating green-blue laminated marl-shale and medium- and thick-bedded to massive and very porous lacustrine limestone. The lacustrine sequence is 10 to 200 m thick and is overlain conformably by fluvial and fine-grained red beds that are very rich in micro-mammalian fossils (e.g., *Turogontherium minus, Mimomys polonicus*, *Mimomys septimamus*, *Mimomys occitanus Canis odessanus, Vulpes alopecoides, Stephanorhinus meparhinus*) (Sickenberg *et al*. 1975; Saraç 2003). This rich fossil content in the topmost fluvial horizon yields a Middle Pliocene age (M16), and the age of the volcano-sedimentary sequence exposed near the apex of the Isparta Angle is between the latest Early Miocene and the Mid-Pliocene (Figure 3).

Modern Graben Fill Units

The Şuhut graben is a Plio–Quaternary extensional structure with only modern infill that consists of: (1) coarser-grained lateral marginal deposits (Kızılören Formation), and (2) finer-grained axial depocentral (alluvial plain) deposits. The marginal succession forms a continuous and margin-parallel blanket of mostly older fan-apron deposits formed by coalescence of alluvial fans and widespread intervening thick slope scree. The older fan-apron deposit is the lowest unit of the modern graben infill. It displays well-exposed outcrops in faulted contact with the pre-modern graben fill units and older rocks along the margins of the Şuhut graben (Figure 4). At some localities, particularly the Bademli-Çobankaya, Mahmutköyü and Ağzıkara areas, the older fan-apron deposits occur as faulted, uplifted and dissected terraces related to step-like normal faulting (Figure 5a). The broadest and thickest (~225 m) fan apron deposits are well-exposed in the Ağzıkara area, where they display proximal, medial and distal facies of a fan sequence. The proximal facies has a basal unsorted, polygenetic boulder-block conglomerate along the unconformity and faulted contacts with the older basement rocks. It continues upwards with the alternation of conglomeratic sandstone, siltstone, clast-supported red mudstone, claystone with caliche patches and lensoidal intercalations of channel conglomerates, indicative of the distal facies of a fluvial drainage system (Figure 5b). The basal facies or conglomerate consists of boulder–block sized (up to 60 cm in diameter) angular to semi-rounded clasts of mostly quartzite, schist, marble, radiolarite, vesicular basalt, andesite, trachyte, marl and also porous lacustrine limestone set in a silty matrix bounded by an iron-rich calcite matrix. Based on their lithofacies, internal structure and stratigraphical position, these fan-apron deposits can be correlated with the modern graben infill, the Plio–Quaternary Kızılören Formation, of the Sandıklı graben to the southwest of and outside the Şuhut graben (Koçyiğit *et al.* 2001). Except for the faulted contact, where it dips steeply (Figure 5a) and/or is tilted towards the margin-boundary fault, the nearly-flat-lying fan-apron deposit overlies with angular unconformity the deformed (folded) pre-graben fill units of late Early Miocene–Middle Pliocene age (Sickenberg *et al.* 1975; Koçyiğit *et al.* 2001; Saraç 2003). A latest Pliocene–Quaternary age is therefore assigned to the marginal facies of the modern graben infill.

Another significant member of the coarser-grained marginal facies consists of the younger alluvial fans up to 16 km^2 in area along the faulted margins of the graben. Some of them are very large and mapped on the scale of 1/100 000. They formed where transverse streams and rivers, such as the Çatak, Ellez and Şuhut drainage systems, enter the graben (Figure 4). The alluvial fans consist of partly lithified, unsorted and polygenetic boulder- to pebble-sized sediments in their proximal parts and coarse-grained sand and silt in their distal parts. Both older fan-apron deposits and younger fan deposits grade into finer-grained axial alluvial plain deposits consisting of sand, silt and organic material-rich mud to clay. The maximum thickness of both recent alluvial fan and depocentral facies is about 60 m. The unconsolidated and water-saturated fine-grained sediments comprising the distal parts of alluvial fans and depocenter of the graben are thixotropic (water-saturated granular material may be readily transformed from a solid to liquid state). During earthquakes, this may result from an increase in pore-water pressure caused by compaction because of intense shaking. Liquefaction of near surface water-saturated silts and sands causes the materials to lose their shear strength and flow (Keller & Pinter 1996). As a result, buildings may tilt or sink into the liquefied sediments, and buried tanks or pipelines may float to the surface.

Structures

Two groups of structures are well exposed both within and outside of the Şuhut graben. These are the contractional palaeotectonic structures, such as folds, and the extensional neotectonic structures such as the Şuhut graben and its margin-boundary normal faults

Folds in Pre-Modern Graben Fill Units

The western margin of the Şuhut graben consists mostly of the Afyon strato-volcanic complex, so does not display well-developed layering and bedding planes, and therefore the plastic deformation pattern of these rocks could not be identified. The eastern margin of the graben consists mainly of a fluvio-lacustrine sequence (e.g., Upper Miocene–Middle Pliocene Türkbelkavak Formation) (Çevikbaş *et al* 1988; Koçyiğit *et al.* 2001) and it displays well-developed bedding planes. Their dips vary between 5° and 50°, and average 23° to both west and east, and so they form a series of anticlines and synclines with the NNE-trending axes (Figures 4 & 6a). A stereographic plot of a series of folded beds indicates that the area has experienced a N76°W-to S76°E-directed contraction (Figure 6b). The age of the last deformational phase of the palaeotectonic period is the end of the Middle Pliocene, and predates the latest Pliocene.

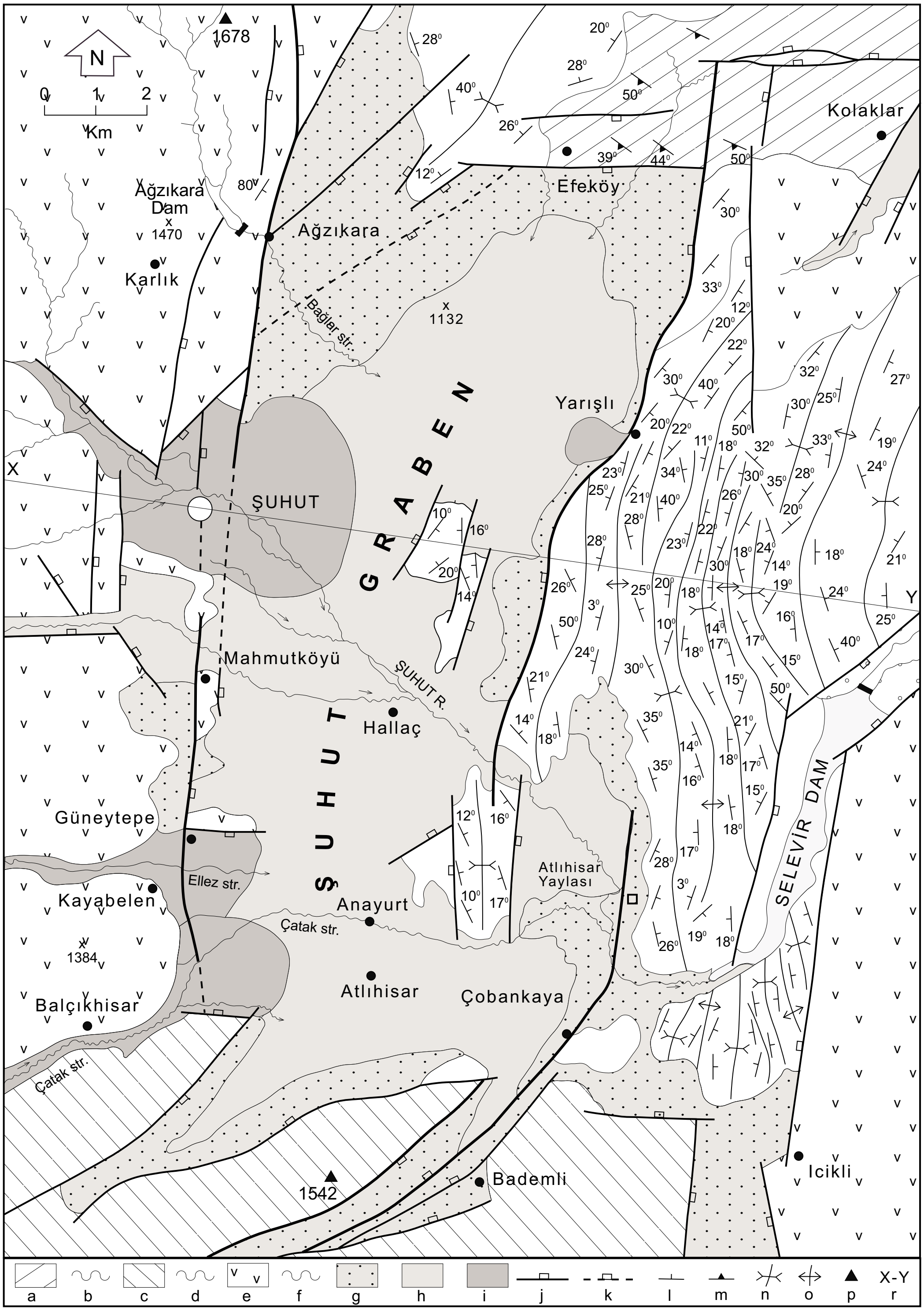


Figure 4. Geological map of the Şuhut graben and its vicinity. a– pre-Jurassic metamorphic rocks; b– nonconformity; c– Jurassic–Cretaceous limestone; d– angular unconformity; e– uppermost Lower Miocene–Middle Pliocene pre-modern graben fill units; f– angular unconformity; g– Plio–Quaternary modern graben infill; h– Holocene modern graben infill; i– alluvial fan; j– oblique-slip normal fault; k– buried normal fault; l– strike and dip of bedding; m– strike and dip of foliation; n– syncline axis, o– anticline axis, p– elevation above sea level and r– line of geological cross section.

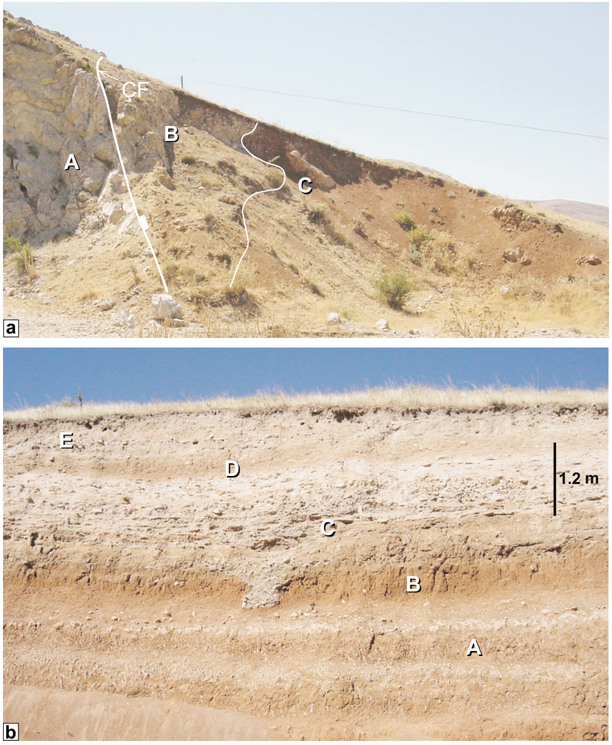


Figure 5. (a) Photograph showing faulted contact between the Jurassic–Lower Cretaceous limestones (A) and the Plio–Quaternary modern graben infill (B– basal conglomerate, and C– flood plain deposits). ÇF. Çobankaya oblique-slip normal fault (200 m SSW of Çobankaya village); (b) photograph illustrating various facies of the Plio–Quaternary modern graben infill. A– red mudstone with caliche patches, B– dark red clayey mudstone-claystone; C– channel conglomerate, D– pebble-supported red mudstone-sandstone alternation, and E– fluvial conglomerate (3 km NE of Ağzıkara village).

Şuhut Graben

The Şuhut graben is about 8–11-km-wide, 24-km-long and trends N–S (Figures 7 & 8). It is located on the southern shoulder of the Akşehir–Afyon graben near the apex of the outer Isparta Angle (Figures 1 & 2). The Şuhut graben developed on both the pre-modern graben fill units and older rocks, such as the pre-Jurassic metamorphic rocks and Jurassic–Lower Cretaceous platform carbonates (Figure 4). The Şuhut graben, particularly in its southern half, is subdivided into several N–S- and NE-trending sub-grabens and horsts including the Didiköreni, Sıtmadağ, Kayrakdağ, Bademli and the İcikli sub-horsts, and the Başyataşı, Çobankaya and the Selevir sub-grabens (Figure 7).

Morphologically, the Şuhut graben is bounded by linear and curvilinear mountain fronts rising steeply from the fault contact between the older, deformed rocks and the undeformed Plio–Quaternary graben infill. The relief differences between the graben floor and the northern, southern, eastern and western margins are 250 m, 422 m, 272 m and 338 m, respectively. These values show that the southern and western margins of the Şuhut graben are higher than its other margins, probably because of faster subsidence on the western and southern boundary faults. This is also reflected by a series of fault-parallel alluvial fans aligned along the western margin (Figure 7). Several transverse drainage systems issue from hilltops and ridge crests along the elevated footwall blocks of both the western and southern boundary faults and flow east-southeastwards into the Şuhut graben. Accordingly, the older fan-apron deposits are deeply incised and overlain unconformably by the newly-forming fans with apices adjacent to graben margin-boundary faults, implying recent fault motion. These fault-controlled drainage systems include the Çatak, Ellez, Cevizli and Başlar streams, and the Şuhut River. The Şuhut River is the longest drainage system issuing from the highest hill tops at the western margin and flows south-eastwards across the Şuhut graben (Figure 7).

Normal Faults

Structures that shaped the Şuhut graben and play a key role in its evolutionary history are mostly step-like oblique-slip normal faults, which are well-exposed as short (1.3–6 km) to long (up to 16 km) fault segments cutting both the Plio–Quaternary modern graben fill sediments and older rocks. They juxtapose all of these units with each other.

Normal faults of varying size occur at the margins of the Şuhut graben (Figure 7). They display a steplike pattern (a kind of stepped land shape produced by the step faults that are a series of parallel, closely spaced faults over which the total displacement is distributed as the downthrown side is on the same side of each fault) dominated by the master fault and a series of synthetic fault segments facing towards the interior of the graben (Figures 7 & 8). The faults that shape the Şuhut graben occur in several discrete (single) faults and fault sets. These are the Şuhut and Bademli fault sets, and the Ağzıkara, Güneytepe, Kocaçal, Gerdek, Kayrakdağ, Çobankaya and the Yarışlı isolated faults (Figure 7).

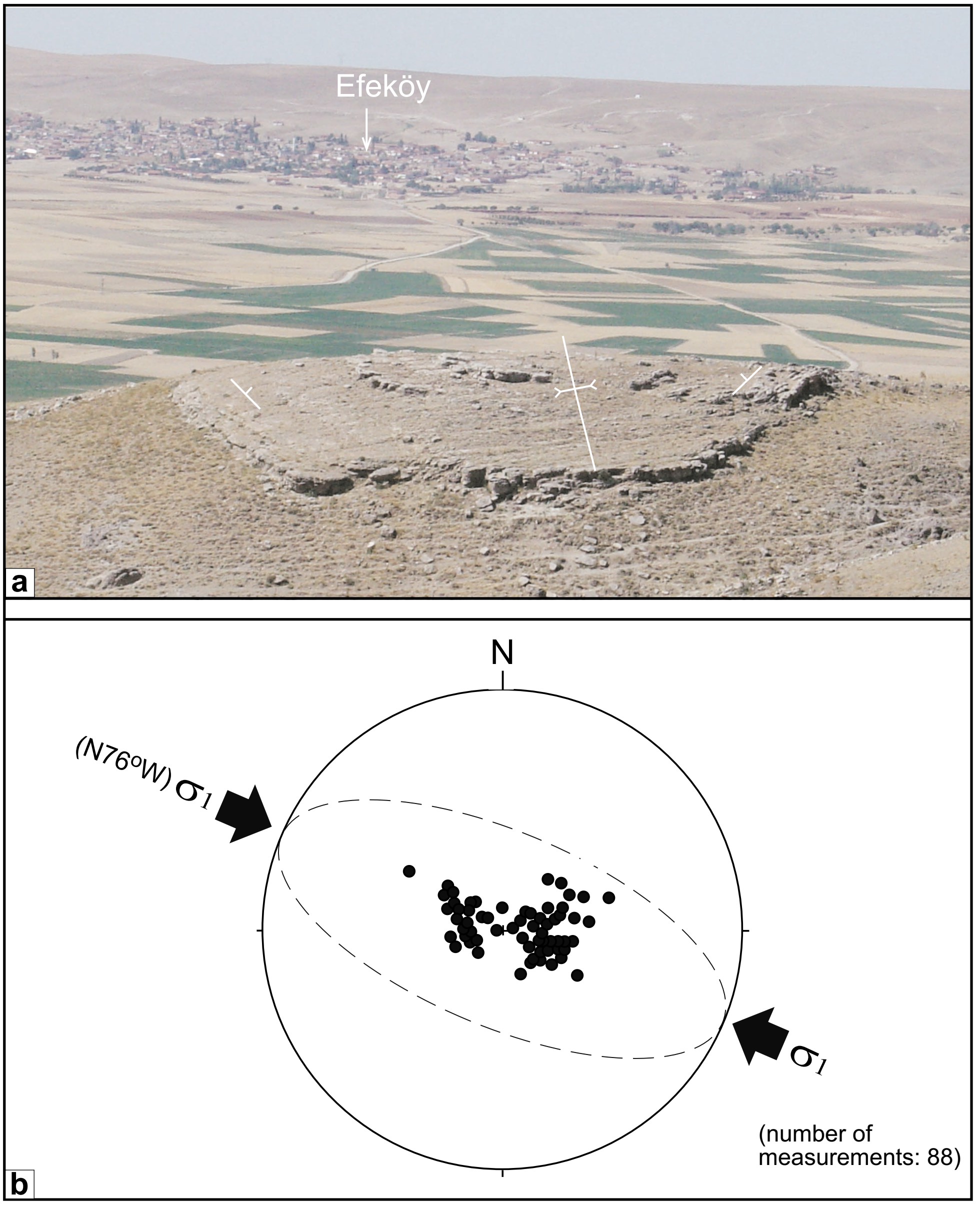


Figure 6. (a) Photograph showing a syncline (in foreground) developed within the transitional sedimentary cover of the Afyon strato-volcanic complex (view to NNW, NE of Yarışlı village). (b) Poles to bedding on the Schmidt's lower hemisphere net. Large black arrows show the shortening direction of the contractional phase that deformed major volcano-sedimentary sequences at the end of the Middle Pliocene: last phase of palaeotectonic period).

The Şuhut fault set is a 2.5-km-wide, 22-km-long normal fault set trending N–S between Belkaracaören (2 km outside the study area) in the north and Balçıkhisar in the south (Figure 7). The Şuhut fault set consists of five 3-16 km long, parallel- to sub-parallel fault segments displaying an eastward-facing stepped land shape that characterizes the western margin of the Şuhut graben (Figure 8). Two of these fault segments are the master faults, while others are synthetic normal faults. Master faults are named here the Ağzıkara fault and the Güneytepe fault (Figure 7). The northern half of the Ağzıkara fault bifurcates into several NE-trending second order fault segments comprising a horse-tail structure.

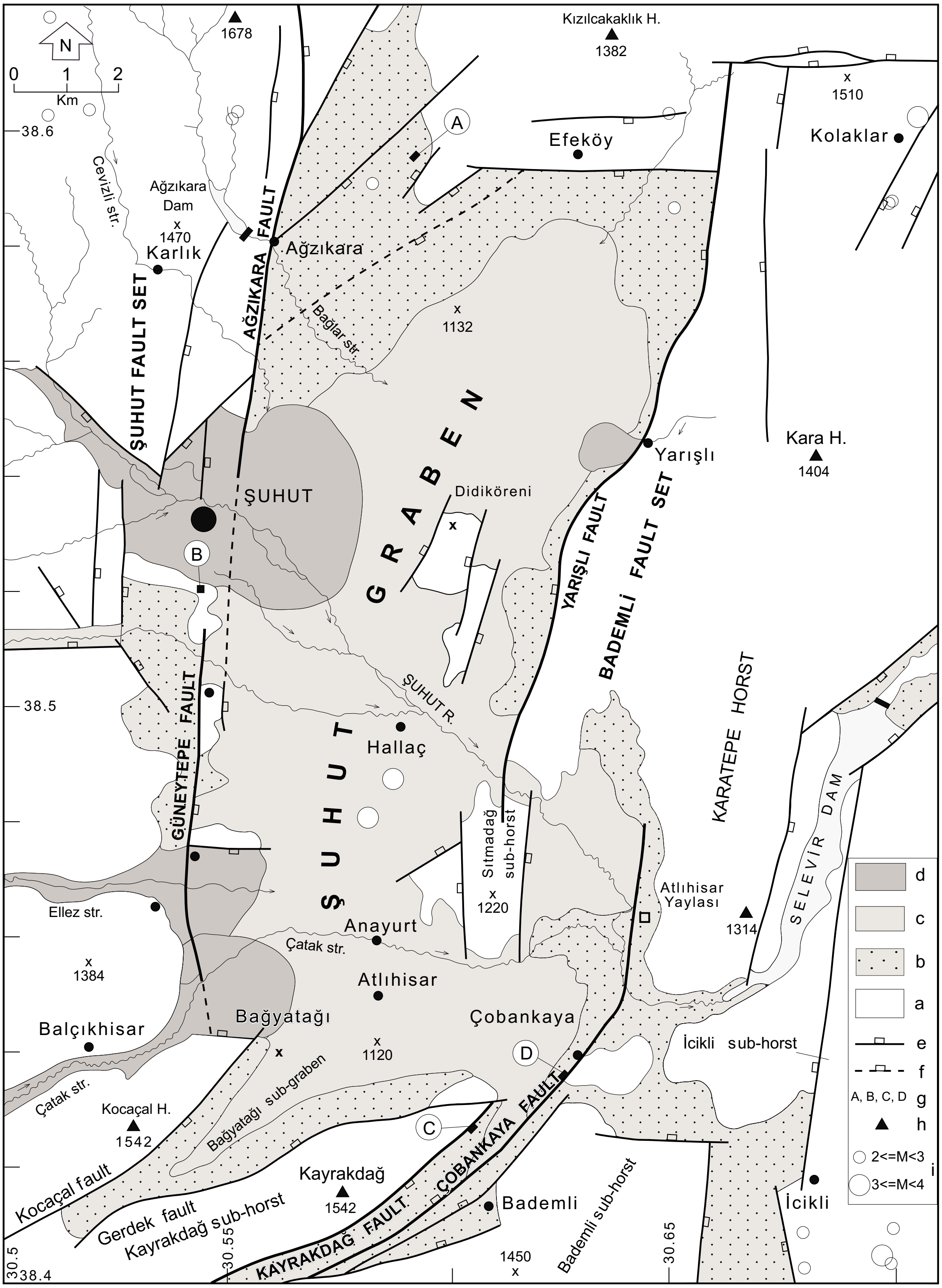


Figure 7. Neotectonic map of the Şuhut graben. **a**– pre-Late Pliocene basement rocks; **b**– Plio–Quaternary modern graben infill; **c**– Holocene modern graben infill; **d**– alluvial fan, **e**– oblique-slip normal fault; **f**– buried normal fault; **g**– sites of slip-plane measurements; **h**– elevation above sea level, and **i**– epicentral locations of both historical and recent earthquakes (parameters of recent earthquakes are taken from KOERİ 2006).

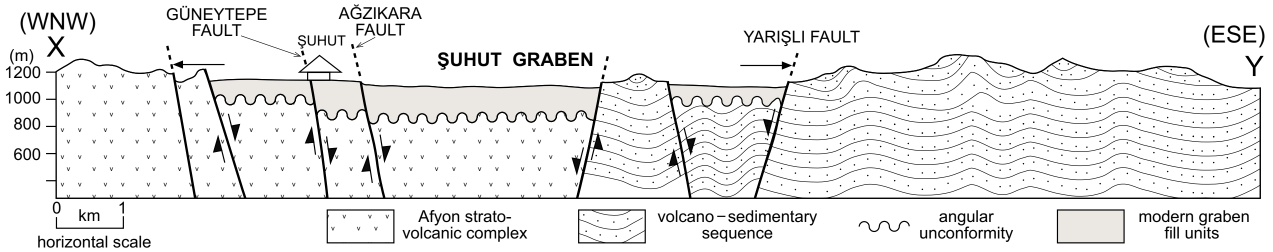


Figure 8. Geological cross-section along the line X–Y that illustrates the present-day configuration of the Şuhut graben (see Figure 4 for its location).

Fault segments comprising the Şuhut fault set cut across both the Afyon strato-volcanic complex and the Plio–Quaternary graben infill, and locally juxtapose them. Evidence for recent activity along these faults include linear to curvilinear fault traces and triangular facets (Figure 9), sudden breaks in slope, fault-controlled drainage systems incised deeply their beds into older fan-apron deposits (e.g., Başlar stream, Cevizli stream, Şuhut River, Ellez and Çatak streams), a series of large alluvial fans aligned along the master faults, faulted, dissected and elevated fault terraces, back-tilting of graben infill, tectonic juxtaposition of older rocks and modern graben infill and intensely crushed and brecciated fault rocks (Figure 7). The master faults of the Şuhut fault set are also seismically active as indicated both by recent micro-seismicity and earthquakes in 1766 and 1862 (Pınar & Lahn 1952; Ergin *et al.* 1967; Öcal 1968; Soysal *et al.* 1981). These two destructive historical earthquakes originated from the Ağzıkara and Güneytepe master faults. The total throw accumulated on the Şuhut fault set is about 400 m, based on the relief between the erosional surfaces of pre-modern graben fill units exposing along the faulted margin and overlain by the 385-m-thick modern graben infill in the graben. This value yields an approximately slip rate of 0.2 mm/yr along the western margin-boundary faults.

The style of deformation and motion direction during the evolutionary history of the Şuhut graben have also been recorded and well-preserved as slip-planes on the faults displacing the mostly clayey mudstone facies of the lowermost modern graben infill. Mesoscopic fault population and slickensides on them are well observed at station A (Figure 7). Kinematic analysis of slip-plane data measured at station A indicates an extensional tectonic regime coeval with the modern graben sedimentation, and an E–W extension (Figure 10a). Likewise, the active western boundary fault, the Güneytepe master fault, also displays well-preserved slickensides. However, stereographic plots of slip-plane data measured at station B reveal ENE–WSW extension (Figure 10b), indicating that the extension direction has rotated up to 15° anticlockwise.

The eastern to southern boundary fault segments are termed the Bademli fault set, which is a 7.5-km-wide, 30-km-long and N–S- to NE-trending normal fault belt extending from the Kayrakdağ district in the south to northeast of Efeköy in the north (Figure 7). It consists of 1.5-km- to 15-km-long fault segments with straight, curved and curvilinear traces (Figure 7). They cut across pre-Upper Pliocene rocks, and divide them into a series of sub-horsts and sub-grabens (Figures 4 & 7). The Bademli fault set also tectonically juxtaposes Plio–Quaternary modern graben infill and older rocks (Figures 5a, 7 & 11a, b), and displays similar morphotectonic features to the Şuhut fault set. The total vertical displacement accumulated along the Bademli fault set is about 390 m, based both on the thickness of the modern graben infill and the relief between the graben floor and its margins.

The Kayrakdağ, Çobankaya and Yarışlı fault segments of the Bademli fault set are longer and more active than its other segments. The 6.5-km-long and NE-trending Kayrakdağ fault segment forms a tectonic contact between the Jurassic–Lower Cretaceous limestone and the Plio–Quaternary fan-apron deposits of the modern Şuhut graben. Slip-plane data measured at station C (Figure 7) reveal ENE–WSW extension (Figure 10c). In the same way, around Bademli and Çobankaya villages, three closely-spaced, parallel to sub-parallel, NE- and N–S-trending and westward concave normal fault segments are well-exposed. They also form the tectonic contacts between the Jurassic–Lower Cretaceous limestone and the Plio–Quaternary modern graben infill

Figure 9. Photograph showing the western boundary fault (Ağzıkara fault) of the Şuhut graben along which Plio–Quaternary graben infill is tectonically juxtaposed with the Middle-Upper Miocene Afyon strato-volcanic complex (view to W, N of Ağzıkara village).

and display a westward facing fault scarp. The 12-km-long Çobankaya fault is here defined as the master fault of the Bademli fault set because its slip rate is faster. Older fan-apron deposit of the Şuhut graben has been cut and uplifted as a fault terrace along the Çobankaya fault (Figure 11b). Slip-plane data measured at station D (Figures 7 & 11c) reveal ENE–WSW extension (Figure 10d).

The longest (15 km) and most active fault segment in the Bademli fault set is the Yarışlı fault. Extending from the Şuhut River in the south to east of Efeköy in the north (Figure 7), it displays a steep, west-facing fault scarp, steplike morphology and a curvilinear fault trace, along which the fan-apron deposits of the modern Şuhut graben are tectonically juxtaposed with various basement rocks, such as metamorphic rocks, the Afyon strato-volcanic complex and its fluvio-lacustrine sedimentary cover. Both the microseismic activity and morphotectonic features such as the uplifted, dissected and fault-suspended terrace deposits, offset and diverted transverse stream beds, very sharp triangular facets and back-tilted blocks indicate that the Yarışlı fault is active (Figures 7, 8 & 11d).

Seismicity

The Şuhut graben is a seismically active extensional structure, as indicated by both historical and recent earthquakes originating in it and at its margins (Table 1) (Ergin *et al.* 1967; Soysal *et al.* 1981; KOERİ 2006). It has been reported that two destructive seismic events, the 1766 and 1862 Şuhut earthquakes, with intensities of VII and VIII-X respectively, based on the Modified Mercalli Intensity Scale (cf. Bolt 1993), occurred in the modern Şuhut graben (Ergin *et al.* 1967; Soysal *et al.* 1981). Their epicenters are located near the western boundary faults, which implies that they originated in the master faults of the Şuhut fault set, the Ağzıkara and the Güneytepe faults (Figure 7). The 1862 event, which was followed by a series of aftershocks, devastated most of the town of Şuhut and caused both surface ruptures and wide-spread liquefaction of water-saturated fine-grained modern graben infill. Over eight hundred people died during this earthquake (Ergin *et al.* 1967). The magnitudes of the most powerful earthquakes originating from both the 16-km-long Ağzıkara and the 12-km-long Güneytepe master faults are Mw=6.47 and Mw=6.3, respectively, based on the equation of Wells & Coppersmith (1994). These estimates are supported by these two historical events.

Nineteen tremors with magnitudes ranging between 2.6 and 3.4 have occurred in and adjacent to the Şuhut graben between 1900 and 2006 (Table 1) (KOERİ 2006). Their epicentral distribution indicates that some fault segments of both the Şuhut and Bademli fault sets are also seismically active. The Yarışlı fault is also active, but no destructive historical earthquakes have been recorded relating to it.

Discussion and Conclusion

After ocean closure and continent-continent collision along the two sub-branches (the İzmir–Ankara and the Inner Tauride oceans) of the northern Neotethys, some

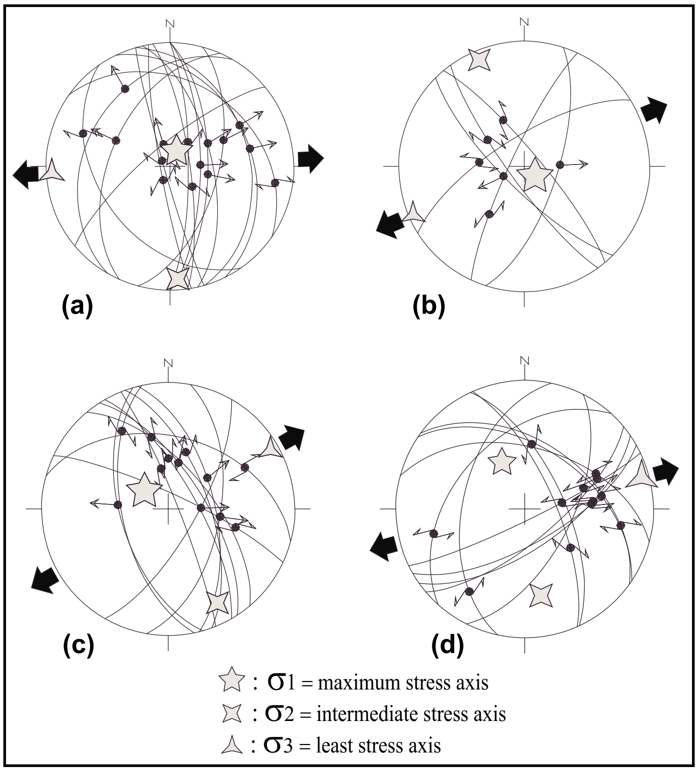


Figure 10. Stereographic plots of both the fault population (a) and boundary faults (b-d) slip-plane data on the Schmidt lower hemisphere net. Large arrows show extension direction in the Şuhut graben.

ophiolitic nappes (such as the Lycian and the Beyşehir-Hoyran nappes) developed and began to move southward (Özgül 1976; Poisson 1977; Gutnic *et al.* 1979; Şengör & Yılmaz 1981; Koçyiğit 1983, 1984b; Waldron 1984; Flecker *et al.* 2005). Continuing (late Palaeocene–early Miocene) intracontinental convergence and progressive nappe emplacements first led to shortening, uplift and overthickening of the crust and, finally to extensional orogenic collapse in southwestern Turkey (e.g., Dewey 1988; Seyitoğlu & Scott 1992; Koçyiğit *et al.* 1999; Bozkurt & Sözbilir 2004). For this reason, this phase of contraction in the late Early Miocene may have played a critical role in triggering the extensional orogenic collapse and, accordingly, the initiation of the *phase-I* continental extension and graben formation in the late Early Miocene (Koçyiğit *et al.* 1999; Bozkurt 2000; Gürer *et al.* 2001; Koçyiğit 2005). In southwestern Turkey, including west Anatolia, west-central Anatolia and the Isparta Angle, the first sediments were marine, but fluvio-lacustrine facies were deposited in the northern part of Isparta Angle under the control of this first phase of extension (extension of palaeotectonic period) (Figure 12a). This first sedimentation and graben formation were accompanied by volcanic activity in some parts of southwestern Turkey (such as İzmir, Uşak, Kütahya and Afyon areas), which resulted in a volcano-sedimentary sequence up to 2 km thick (e.g., Keller & Villari 1972; Becker-Platen *et al.* 1977; Besang *et al.* 1977; Ercan *et al.* 1978; Koçyiğit 1981, 1983, 1984a; Çevikbaş *et al.* 1988; Akal 2003; Erkül *et al.* 2005; Aldanmaz 2006). One of these well-developed sequences (up to 1.5 km in thickness) dominated by a strato-volcanic complex, is exposed in the Afyon region, and includes the Şuhut and Sandıklı areas, comprising the northernmost tip of the Isparta Angle (Figures 2, 3 & 12b) (Keller & Villari 1972; Çevikbaş *et al.*1988; Aydar *et al.* 1996; Koçyiğit *et al.* 2001; Akal 2003). In the Afyon region, and particularly in the Şuhut area, the first sedimentation and graben formation lasted until the mid-Pliocene under the control of an extensional tectonic regime which operated in approximately NW–SE direction (Figure 12a, b).

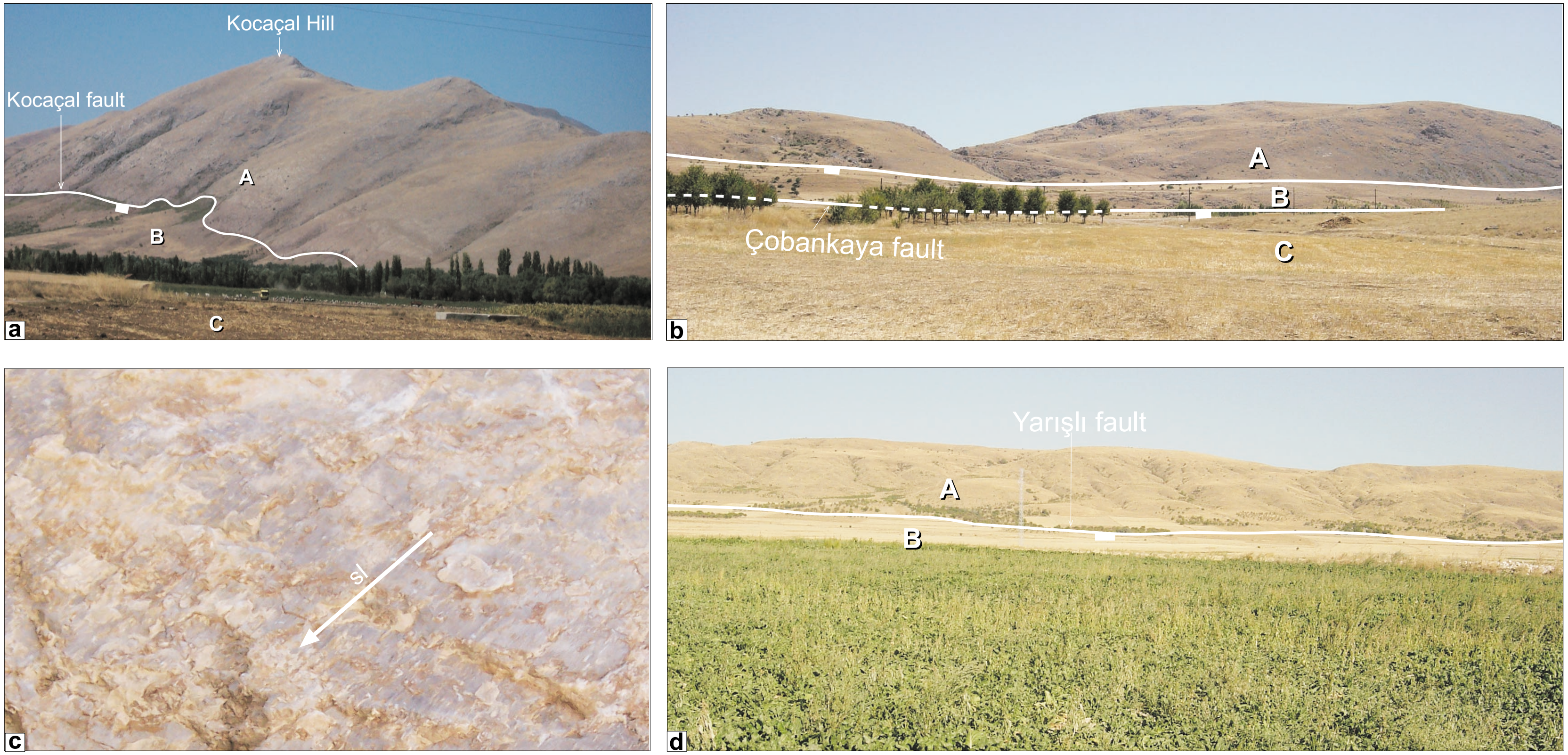


Figure 11. (a) Photograph showing the Kocaçal oblique-slip normal fault. A– Jurassic–Lower Cretaceous limestone, B– Plio–Quaternary modern graben infill, and C– Holocene modern graben infill (view to SSW). (b) Photograph showing Çobankaya fault and the uplifted to suspended fault terrace conglomerate of Plio–Quaternary age (B). A– Jurassic–Lower Cretaceous limestone, and C– Holocene modern graben infill. (c) Close-up view of the Çobankaya fault slickensides illustrating its oblique-slip nature. Long arrow shows motion sense of the hanging-wall block (200 m SSW of Çobankaya village). (d) Photograph showing the northern half of the Yarışlı oblique-slip normal fault along which the pre-Jurassic metamorphic rocks (A) and the Plio–Quaternary modern graben infill (B) are tectonically juxtaposed (view to E; ENE of Efeköy).

Table 1. Historical and recent earthquakes in the Şuhut graben and adjacent areas (Ergin *et al.* 1067; Soysal *et al.* 1981; KOERI 2006).

No	Date	Latitude (N)	Longitude (E)	Magnitude	Region
1	23.05.2004	38.4765	30.6035	2.7	ŞUHUT (AFYON)
2	16.04.2004	38.5895	30.5835	2.9	ŞUHUT (AFYON)
3	05.12.2003	38.5477	30.5007	2.6	ŞUHUT (AFYON)
4	04.06.2003	38.4020	30.6955	3.1	ŞUHUT (AFYON)
5	08.04.2002	38.4000	30.6800	2.8	ŞUHUT (AFYON)
6	16.04.2002	38.4000	30.6980	2.7	ŞUHUT (AFYON)
7	08.04.2002	38.4080	30.6800	2.8	ŞUHUT (AFYON)
8	22.03.2002	38.6020	30.5200	2.9	ŞUHUT (AFYON)
9	22.03.2002	38.6020	30.5100	2.9	ŞUHUT (AFYON)
10	22.03.2002	38.6200	30.5120	2.8	ŞUHUT (AFYON)
11	22.03.2002	38.6000	30.5520	2.9	ŞUHUT (AFYON)
12	22.03.2002	38.6010	30.5530	2.9	ŞUHUT (AFYON)
13	22.03.2002	38.6012	30.6500	2.7	ŞUHUT (AFYON)
14	22.03.2002	38.5852	30.6500	2.8	ŞUHUT (AFYON)
15	22.03.2002	38.5852	30.6982	2.9	ŞUHUT (AFYON)
16	22.03.2002	38.5854	30.6985	2.8	ŞUHUT (AFYON)
17	22.03.2002	38.4800	30.5810	3.4	ŞUHUT (AFYON)
18	22.03.2002	38.4856	30.5882	3.1	ŞUHUT (AFYON)
19	22.03.2002	38.6002	30.7200	3.4	ŞUHUT (AFYON)
20	14.11.1862	38.50	30.55	VIII – X	ŞUHUT (AFYON)
21	1766	38.53	30.55	VII	ŞUHUT (AFYON)

The pre-modern graben infill is deformed by the second contractional phase (last phase of palaeotectonic period) with maximum shortening in an ESE–WNW direction (Figure 12c). This deformation includes: (1) a series of anticlines and synclines with NNE-trending axes developed in the Lower Miocene–Middle Pliocene volcano-sedimentary sequence at the margins of the Şuhut graben (Figures 4 & 6), and (2) an angular unconformity that separates this deformed (folded) pre-modern graben infill from the nearly flat-lying Plio–Quaternary modern graben infill (Figures 3 & 12d). The last contractional phase, which was also recorded in other grabens comprising the southern part of the Isparta Angle and southwestern Anatolia (Koçyiğit *et al.* 1999; Koçyiğit *et al.* 2001; Koçyiğit & Özacar 2003; Poisson *et al* 2003; Koçyiğit 2005), is related to the emergence (formation and appearance of an independent microplate) of the Anatolian microplate and its west-southwestward motion along its boundary faults, the dextral North Anatolian and the sinistral East Anatolian fault systems. Therefore, this last contractional event, which covers the last deformation of the pre-modern graben infill and pre-dates the *phase-II* continental extension (extensional neotectonic period) in southwestern Turkey, particularly in the Şuhut area, is latest mid-Pliocene in age. This conclusion fits well with those of Glover & Robertson (1998a) and Robertson & Comas (1998), who reported that the mid-Pliocene contraction, with sinistral to dextral strike-slip faulting in the Isparta Angle, was followed by pervasive extensional faulting during late Pliocene–Pleistocene times. Thus, the first sedimentation was interrupted by a phase of contraction producing folding, which switched to extensional faulting (second phase of extension: neotectonic period) in the latest Pliocene. The initiation of extension in the Neotectonic period is attributed to the combination of several geodynamic processes, such as the emergence of the Anatolian microplate, its west-southwestward motion,

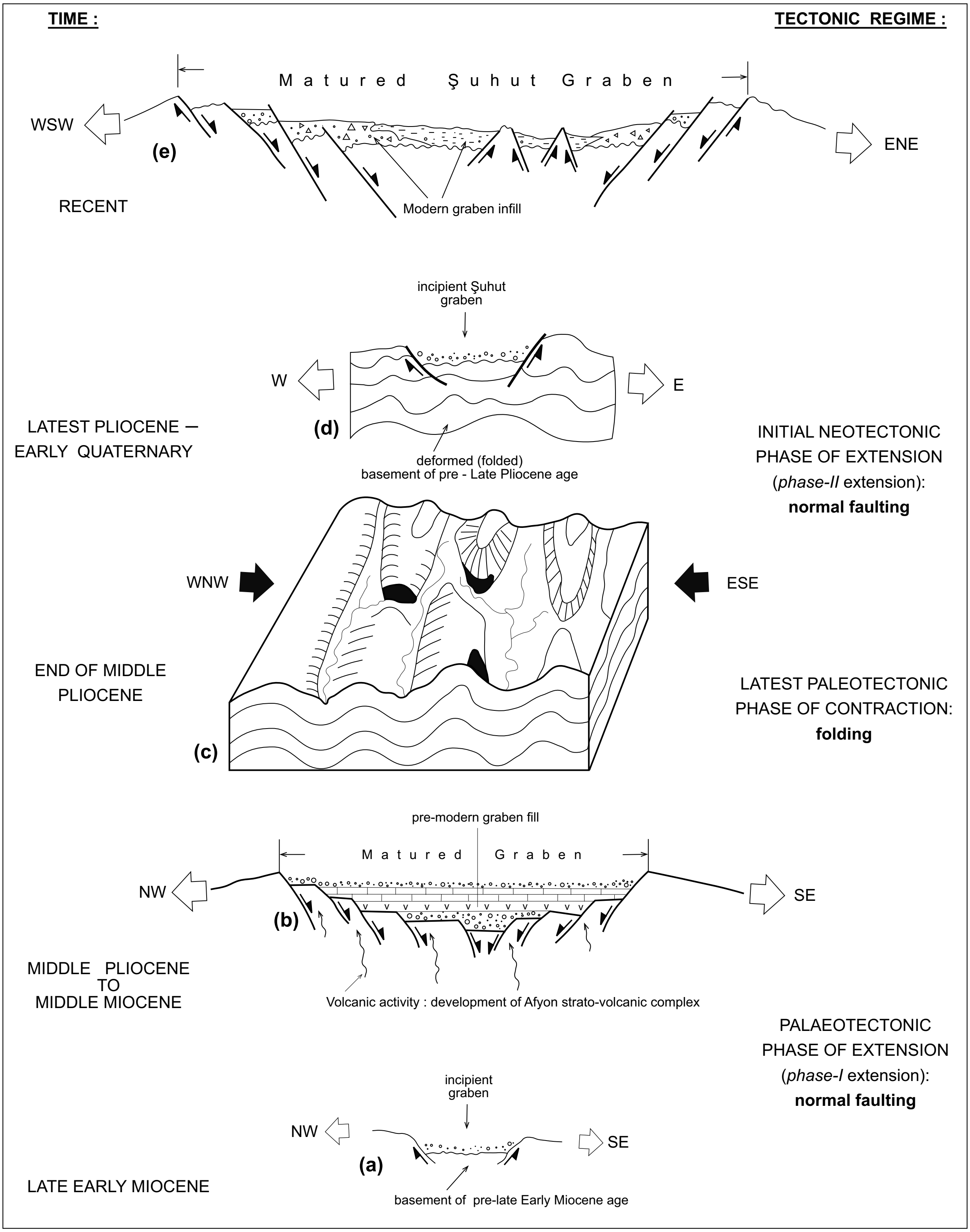


Figure 12. Sketch geological cross-sections and block diagram depicting inversion in palaeotectonic period and the initiation of the extensional neotectonic period in the frame of the development history of the Şuhut graben.

the northward subduction of the southern Neotethys and related slab retreat (Glover & Robertson 1998a; Koçyiğit *et al.* 1999; Kelling *et al.* 2005). This *phase-II* crustal extension led to both the reactivation of most older structures and the formation of new extensional faults. These divided the Isparta Angle and its vicinity into a number of horsts and grabens that shape the present-day configuration of the Isparta Angle horst-graben system (IAHGS) (Figure 2). One of its well-defined members is the Şuhut graben. Based on palaeostress analyses, the σ_3 has had an ENE–WNW direction in the Şuhut graben since the latest Pliocene (Figures 10a–d & 12 d, e).

The Şuhut graben is a N–S-trending and actively growing extensional neotectonic structure about 8–11 km wide, 24 km long, located on the southern shoulder of the major Akşehir-Afyon graben near the apex of the outer Isparta Angle (Koçyiğit 2005). Its current activity is evidenced both by a series of morphotectonic features, and devastating historical earthquakes and tremors. Boundary faults of the graben, particularly the Ağzıkara, Güneytepe, Çobankaya and the Yarışlı normal faults have the potential of creating destructive earthquakes with magnitudes between M=6.3 and Mw=6.5. This was previously proved by the occurrence of two devastating earthquakes, the 1766 and the 1862 seismic events (Ergin *et al*. 1967; Soysal *et al.* 1981) sourced from the Ağzıkara and the Güneytepe master faults of the Şuhut graben. However, the Çobankaya and the Yarışlı faults still retain their long-term seismic quiescence. In addition, Quaternary alluvial fans and the finer-grained alluvial sediments (modern graben infill) can be readily liquified. Therefore, these active faults and the water-saturated modern graben infill have to be taken into account in both the earthquake hazard to earthquake risk analyses and the design of city planning of Şuhut Town, which is located on a large alluvial fan.

Based on data presented and discussions carried out above, the following are concluded: (a) the evolutionary history of the horst-graben system in the Isparta Angle is episodic, i.e. it evolved at two extensional phases (*phase-I* and *phase-II* extensional periods) interrupted by an intervening short-term contractional phase; (b) the *phase-II* extension equates with the neotectonic period in southwest Turkey, and it commenced in the latest Pliocene in the Isparta Angle, (c) the eastern limit of the 'west Anatolian extensional domain' is not confined to the area west of a N–S-trending imaginary line connecting the western margin of the Gulf of Antalya and the eastern part of Sea of Marmara: on the contrary, it continues eastwards up to the Salt Lake and the fault zone bounding its eastern margin, the Salt Lake Fault Zone, (d) the Isparta Angle is characterized by four horst-graben sets and their active bounding faults; they are the E–W-, NW-, NE- and N–S-trending horst-graben sets, and (e) continental extension in the Isparta angle is multidirectional owing to the distributed stress systems revealed by both palaeostress analyses and focal mechanism solutions of earthquakes (Koçyiğit 2005).

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References

Akal, C. 2003. Mineralogy and geochemistry of melilite leucitites, Balçıkhisar, Afyon, (Turkey). *Turkish Journal of Earth Sciences* **12**, 215–239.

Akay, E. & Uysal, Ş. 1985. *Stratigraphy, Sedimentology and Structural Geology of the Neogene Deposits in the West of Central Taurides (Antalya)*. Mineral Research and Exploration Institute of Turkey (MTA) Report no. **7799** [in Turkish, unpublished].

Akgün, F. & Akyol, E. 1992. Yukarıkaşıkara ve Yarıkkaya (Isparta) kömürlerinin karşılaştırmalı palinostratigrafisi ve paleoekolojisi [Correlative palinostratigraphy and palaeoecology of the Yukarıkaşıkara and Yarıkkaya (Isparta) coals]. *Turkish Association of Petroleum Geologists Bulletin* **4**, 129–139 [in Turkish with English abstract].

Akgün, F. & Sözbilir, H. 2001. A Palinostratigraphic approach to the SW Anatolian molasse basin: Kale-Tavas and Denizli molasses. *Geodinamica Acta* **14**, 71–93.

Alçiçek, M.C., Kazancı, N. & Özkul, M. 2005. Multiple rifting pulses and sedimentation pattern in the Çameli Basin, southwestern Anatolia, Turkey. *Sedimentary Geology* **173**, 409–431.

Aldanmaz, E. 2006. Mineral-chemical constraints on the Miocene calc-alkaline and shoshonitic volcanic rocks of western Turkey: disequilibrium phenochryst assemblages as indicators of magma storage and mixing conditations. *Turkish Journal Earth Sciences* **15**, 47–73.

ALTUNEL, E., BARKA, A.A. & AKYÜZ, S. 1999. Paleoseismology of Dinar fault, SW Turkey. *Terra Nova* 11, 297–302.

AMBRASEYS, N.N. 2001. Reassessment of earthquakes 1900–1999 in the Eastern Mediterranean and Middle East. *Geophysical Journal International* 145, 471–485.

AMBRASEYS, N.N. & FINKEL, C.F. 1987. Seismicity of Turkey and neighbouring regions, 1899-1915. *Annales Geophysicae* 87, 701–725.

AMBRASEYS, N.N. & TCHALENKO, J.S. 1972. Seismotectonic aspect of the Gediz, Turkey, earthquake of March 1972. *Geophysical Journal of the Royal Astronomical Society* 30, 229–252.

ANGELIER, J., DUMONT, J.F., KARAMANDERESİ, H., POISSON, A., ŞIMŞEK, Ş. & UYSAL, Ş. 1981. Analyses of fault mechanisms and expansion of southwestern Anatolia since the late Miocene. *Tectonophysics* 75, T1–T9.

ARPAT, E. & BİNGÖL, E.1969. The rift system of western Turkey: thoughts on its development. *Mineral Research and Exploration Institute (MTA) of Turkey Bulletin* 73, 1–9.

AYDAR, E., BAYHAN, H. & ZİMİTOĞLU, O. 1996. Investigation of volcanological and petrological evolution of Afyon stratovolcano. *Earth Sciences* 18, 87–107.

BARKA, A.A., REILINGER, R.E., ŞAROĞLU, F. & ŞENGÖR, A.M.C. 1995. Isparta Angle: Its importance in the noetectonics of the Eastern Mediterranean Region. *In:* PIŞKİN, D., ERGÜN, M., SAVAŞÇIN, M.Y. & TARCAN, G. (eds), *Proceedings of International Earth Science Colloquium on the Aegean Region* 1, 3–18.

BAŞ, H. 1986. Domaniç-Tavşanlı-Kütahya-Gediz yöresinin Tersiyer jeolojisi [Tertiary geology of the Domaniç-Tavşanlı-Kütahya-Gediz area]. *Chamber of Geological Engineers Bulletin* 27, 11–18 [in Turkish with English abstract].

BECCALETTO, L. & STEINER, C. 2005. Evidence of two-stage extensional tectonics from the northern edge of the Edremit graben, NW Turkey. *Geodinamica Acta* 18, 283–297.

BECKER-PLATEN, J., BENDA, L. & STEFFENS, P. 1977. Litho und biostratigraphiche Deutung radiometrischer Altersbestimungen aus dem Jungtertiar der Turkei. *Geologischen Jahrbuch Reihe* B 25, 139–167.

BESANG, C., ECKHARDT, F.C., HARRE, W., KREUZER, H. & MÜLLER, P. 1977. Radiometriche Altersbestimungen an Neogenen Eruptivgesteinen der Turkei. *Geologischen Jahrbuch Reihe* B 25, 3–36.

BLUMENTHAL, M.M. 1951. *Batı Toroslar'da Alanya Ard Ülkesinde Jeolojik İncelemeler [Geological Investigations in Alanya Hinterland, Western Taurides].* Mineral Research and Exploration Institute (MTA) of Turkey Publications no. 5.

BLUMENTHAL, M.M. 1963. Le Systeme structural de Taurus sud Anatolien: Livre a la mem. Prof. P. Fallot: Mem.hs-ser. *Buletin de la Société Gélogique de France* 11, 611–662.

BOLT, B.A. 1993. *Earthquakes*. W.H. Freeman and Company, New York.

BORAY, A., ŞAROĞLU, F. & EMRE, Ö. 1985. Isparta Büklümünün kuzey kesiminde D–B daralma için bazı veriler [Evidence for E–W shortening in the north of the Isparta Angle]. *Chamber of Geological Engineers Bulletin* 23, 9–20 [in Turkish with English abstract].

BOZKURT, E. 2000. Timing of extension on the Büyük Menderes Graben, western Turkey and its tectonic implications. *In*: BOZKURT, E., WINCHESTER, J.A. & PIPER, J.D.A. (eds), *Tectonics and Magmatism in Turkey and the Surrounding Area*. Geological Society, London, Special Publications 173, 385–403.

BOZKURT, E. 2001. Neotectonics of Turkey – a synthesis. *Geodinamica Acta* 14, 3–30

BOZKURT, E. 2002. Discussion on the extensional folding in the Alaşehir (Gediz) Graben, western Turkey. *Journal of the Geological Society, London* 159, 105–109.

BOZKURT, E. & SÖZBİLİR, H. 2004. Tectonic evolution of the Gediz graben: field evidence for an episodic, two-stage extension in western Turkey. *Geological Magazine* 141, 63–79.

BOZKURT, E. & ROJAY, B. 2005. Episodic, two-stage Neogene extension and short-term intervening compression in western Turkey: field evidence from the Kiraz basin and Bozdağ horst. *Geodinamica Acta* 18, 299–316.

COHEN, H.A., DART, C.J., AKYÜZ, H.S. & BARKA, A.A. 1995. Syn-rift sedimentation and structural development of Gediz and Büyük Menderes graben, western Turkey. *Journal of the Geological Society, London* 152, 629–638.

ÇEVİKBAŞ, A., ERCAN, T. & METİN, Ş. 1988. Geology and regional distribution of Neogene volcanics between Afyon and Şuhut. *Journal of Pure and Applied Sciences* 21, 479–499.

DEMİRTAŞLI, E., TURHAN, N., BİLGİN, A.Z. & SELİM, M. 1984. Gelogy of the Bolkar Mountains. *In*: TEKELİ, O & GÖNCÜOĞLU, M.C. (eds), *Geology of the Tauırus Belt.* International Symposium, 26–29 September 1983, Ankara–Turkey, Proceedings, 125–141.

DERCOURT, J., RICOU, R.C. & VRIELYNCK, B. 1993. *Atlas Tethys Paleoenvironmental Maps*. CCGM.

DEWEY, J.F. 1988. Extensional collapse of orogens. *Tectonics* 7, 1123–1139.

DUMONT, J.F. 1976. Isparta kıvrımı ve Antalya naplarının orijini: Torosların Üst Kretase tektojenezi ile oluşmuş yapısal düzeninin büyük bir dekroşman, transtorik arızayla ikiye ayrılması varsayımı [Origin of the Isparta Angle and Antalya Nappes: hypothesis that the Upper Cretaceous tectogenesis-induced stratigraphy of Taurides has been broken-up into two parts by a big decollement (trans-taurique scar)]. *Mineral Research and Exploration Institute (MTA) of Turkey Bulletin* 86, 56–67 [in Turkish with English abstract].

DUMONT, J.F. & KEREY, E. 1975. Eğirdir gölü güneydoğusunun temel jeolojik etüdü [Basic geological study of southeastern part of Lake Eğirdir]. *Geological Society of Turkey Bulletin* 18, 169–174 [in Turkish with English abstract].

DUMONT, J.F., UYSAL, Ş., ŞIMŞEK, Ş., KARAMANDERESİ, İ.H. & LETOUZCY, F. 1979. Güneybatı Anadolu'daki grabenlerin oluşumu [Graben formation in Southwest Anatolia]. *Mineral Research and Exploration Institute (MTA) of Turkey Bulletin* 92, 7–17.

ERCAN, T., DİNÇEL, A., METİN, S., TÜRKECAN, A. & GÜNAY, E. 1978. Uşak yöresindeki Neojen havzaların jeolojisi [Geology of Nogene basins in Uşak area]. *Geological Society of Turkey Bulletin* 21, 97–106 [in Turkish with English abstract].

ERCAN, T., SATIR, M., KREUZER, H., TÜRKECAN, A., GÜNAY, E., ÇEVİKBAŞ, A., ATEŞ, M., & CAN, B. 1985. Batı Anadolu'daki Senozoyik volkaniklere ait yeni kimyasal, izotopik ve radyometrik verilerin yorumu [Interpretation of new chemical, isotopic and radiometric data of the Cenozoic volcanics in west Anatolia]. *Geological Society of Turkey Bulletin* **28**, 121–136 [in Turkish with English abstract].

ERGİN, K., GÜÇLÜ, U. & UZ, Z. 1967. *Türkiye ve Civarının Deprem Kataloğu (Milattan Sonra 11. Yıldan 1964 Sonuna Kadar) [An Earthquake Catalog for Turkey and Its Environs (11 BC to the end of 1964)]*. İstanbul Technical University, Faculty of Mines, Department of Geophysical Engineering Publication [in Turkish].

ERKÜL, F., HELVACI, C. & SÖZBİLİR, H. 2005. Stratigraphy and geochronology of the Early Miocene volcanic units in the Bigadic borate basin, western Turkey. *Turkish Journal of Earth Sciences* **14**, 227–253.

EYİDOĞAN, H. & JACKSON, J. 1985. A seismological study of normal faulting in Demirci, Alaşehir and Gediz earthquakes of 1969–70 in western Turkey: implication for the nature and geometry of deformation in the continental crust. *Geophysical Journal of the Royal Astronomical Society* **81**, 569–607.

FLECKER, R., POISSON, A. & ROBERTSON, A.H.F. 2005. Facies and palaeogeographic evidence for the Miocene evolution of the Isparta Angle in its regional eastern Mediterranean context. *Sedimentary Geology* **173**, 277–314.

GLOVER, C.P. & ROBERTSON, A.H.F. 1998a. Neotectonic intersection of the Aegean and Cyprus tectonic arcs: extensional and strike-slip faulting in the Isparta Angle, SW Turkey. *Tectonophysics* **298**, 103–132.

GLOVER, C.P. & ROBERTSON, A.H.F. 1998b. Role of regional extension and uplift in the Plio–Pleistocene evolution of the Aksu basin, SW Turkey. *Journal of the Geological Society, London* **155**, 365–387.

GUTNIC, M., MONOD, O., POISSON, A. & DUMONT, J.D. 1979. Géologie des Taurides occidentals (Turquie). *Bulletin de la Société Géologique de France* **58** (137), 112 p.

GÜRER, Ö.F., BOZCU, M., YILMAZ, K. & YILMAZ, Y. 2001. Neogene basin development around Söke-Kuşadası (western Anatolia) and its bearing on tectonic development of the Aegean region. *Geodinamica Acta* **14**, 57–69.

HETZEL, R., RING, U., AKAL, C. & TROESCH, M. 1995. Miocene NNE-directed extensional unroofing in the Menderes Massif, southwestern Turkey. *Journal of the Geological Society, London* **152**, 639–654.

JAFFEY, N. & ROBERTSON, A.H.F. 2005. Non-marine sedimentation associated with Oligocene–Recent exhumation and uplift of the Central Taurus Mountains, S Turkey. *Sedimentary Geology* **173**, 53–89.

KALAFAT, D. 1998. Anadolu'nun tektonik yapılarının deprem mekanizması açısından irdelenmesi [Tectonic structures of Anatolia based on earthquakes mechanism]. *Earthquake Research Bulletin* **77**, 1–217 [in Turkish with English abstract].

KARABIYIKOĞLU, M., TUZCU, S., ÇINER, A., DEYNOUX, M., ÖRÇEN, S. & HAKYEMEZ, A. 2005. Facies and environmental setting of the Miocene coral reefs in the late-orogenic fill of the Antalya basin, western Taurides, Turkey: implications for tectonic control and sea-level changes. *Sedimentary Geology* **173**, 319–345.

KARAMAN, E. 1986. Burdur dolayının genel stratigrafisi [General stratigraphy of Burdur area]. *Bulletin of Isparta Engineering Faculty* **2**, 23–36 [in Turkish with English abstract].

KAYA, O. 1972. Tavşanlı yöresi ofiyolit sorununun ana çizgileri [Main lines of the ophiolite problem in Tavşanlı area]. *Geological Society of Turkey Bulletin* **XV**, 26–108 [in Turkish, with English absract].

KAYA, O., ÜNAY, E., SARAÇ, G., EICHHORN, S., HASSENRÜCK, S., KNAPPE, A., PEKDEĞER, A. & MAYDA, S. 2004. Halitpaşa transpressive zone: implications for an Early Pliocene compressional phase in central western Anatolia, Turkey. *Turkish Journal of Earth Sciences* **13**, 1–13.

KELLER, E.A. & PINTER, N. 1996. *Active Tectonics*. Prentice Hall Inc.

KELLING, G., ROBERTSON, A.H.F. & BUCHEM, F.V. 2005. Cenozoic sedimentary basins of southern Turkey: an introduction. *Sedimentary Geology* **173**, 1–13.

KELLER, J. 1974. Quaternary maar volcanism near Karapınar in Central Anatolia. *Bulletin of Volcanology* **38**, 378–396.

KELLER, J. 1983. Potassic lavas in the orogenic volcanism of the Mediterranean area. *Journal of Volcanology and Geothermal Research* **18**, 321–335.

KELLER, J. & VILLARI, L. 1972. Rhyolitic ignimbrites in the region of Afyon (Central Anatolia). *Bulletin of Volcanology* **36**, 342–358.

KISSEL, C. & POISSON, A. 1987. Etude Paléomagnétique des formations Cénozoiques des Beydağları (Tauride occidentals). *Comptes Rendue de l'Academie des Seciences de Paris* **304**, 343–348.

KISSEL, C., AVERBUCH, O., FRIZON DE LAMOTTE, D., MONOD, O. & ALLERTON, S. 1993. First paleomagnetic evidence for a post-Eocene clock-wise rotation of the Western Taurides thrust belt east of Isparta reentrant (southwestern Turkey). *Earth and Planetary Science Letters* **117**, 1–14.

KOCAEFE, S. 1982. Actual tectonics of the western Anatolia. *Bulletin of Institute of Earth Sciences of Hacettepe University* **9**, 149–162.

KOÇYİĞİT, A. 1981. Isparta Büklümünde (Batı Toroslar) Toros karbonat platformunun evrimi [Evolution of the Tauride carbonate platform in the Isparta Angle (Western Turides)]. *Geological Society of Turkey Bulletin* **24**, 15–23 [in Turkish with English abstract].

KOÇYİĞİT, A.1983. Hoyran gölü (Isparta Büklümü) dolayının tektoniği [Tectonics of Lake Hoyran (Isparta Angle) area]. *Geological Society of Turkey Bulletin* **26**, 1–10 [in Turkish with English abstract].

KOÇYİĞİT, A.1984a. Güneybatı Türkiye ve yakın dolayında levha içi yeni tektonik gelişim [A new intracontinental tectonic regime in southwest Turkey and its close vicinity]. *Geological Society of Turkey Bulletin* **27**, 1–16 [in Turkish with English abstract].

KOÇYIĞIT, A.1984b. Tectono-stratigraphic characteristics of Hoyran Lake region (Isparta Bend). *In*: TEKELİ, O & GÖNCÜOĞLU, C. (eds), *Geology of the Taurus Belt.* International Symposium, 26–29 September 1983, Ankara-Turkey, Proceeding, 53–67.

KOÇYIĞIT, A. 1996. Lakes region graben-horst system, SW Turkey: Differential stretching and commencement age of the extensional tectonic regime. *National Marine Geological and Geophysical Programme-Workshop, İstanbul, Proceedings*, 99–103.

KOÇYIĞIT, A. 2000. Güneybatı Türkiye'nin depremselliği [Seismicity of Southwest Turkey]. *Symposium on the Seismicity of Western Turkey (BADSEM) 2000, Abstracts and Proceeding*, 30–38 [in Turkish, with English abstract].

KOÇYIĞIT, A. 2005. Denzili Graben-Horst System and the eastern limit of the West Anatolian continental extension: basin fill, structure, deformational mode, throw amount and episodic evolutionary history, SW Turkey. *Geodinamica Acta* 18, 167–208.

KOÇYIĞIT, A., KAYMAKÇI, N., ROJAY, B., ÖZCAN, E., DIRIK, K. & ÖZÇELIK, Y. 1991. *İnegöl-Bilecik-Bozöyük Arasında Kalan Alanın Jeolojik Etüdü [Geological Investigation of the Area Between İnegöl–Bilecik and Bozöyük]*. ODTÜ-TPAO Joint Project. Project no. 90-03-09-01-05 [in Turkish, unpublished].

KOÇYIĞIT, A., AKYOL, E., BOZKURT, E. & BEYHAN, A. 1997. *Antalya Körfezi Karasal Çöküntüsünün Neotektonik Özellikleri, Depremselliği ve Daha Önceki Denizaltı Çalışmalarıyla Karşılaştırılması [Neotectonic and Seismic Characteristics of the Onshore Section of the Antalya Bay, and its Correlation with Previous Works]*. TUBİTAK–YDABÇAG Project no. 437/G [in Turkish with English abstract, unpublished].

KOÇYIĞIT, A., CIHAN, M. & ÖZACAR, A. 2001. *Seismicity of Dombayova and Sandıklı (Afyon) Grabens and Kinematic Analysis of Their Margin-bounding Faults*. TÜBİTAK Project Report no. 199Y007 [unpublished].

KOÇYIĞIT, A. & ÖZACAR, A. 2003. Extensional neotectonic regime through the NE edge of the outer Isparta Angle, SW Turkey: New field and seismic data. *Turkish Journal of Earth Sciences* 12, 67–90.

KOÇYIĞIT, A., ÖZACAR, A. & CIHAN, M. 2000b. Batı Anadolu horst-graben sisteminin doğu uzantısı ve Isparta Açısı ile ilişkisi nedir? 'Fetiye-Burdur Zonu' olarak bilinen yapının tektonik niteliği [What is the eastern continuation of the west Anatolian horst-graben system and its relationship with Isparta Angle? Tectonic nature of the so-called 'Fethiye-Burdur Zone']. *Fourth Annual Meeting of Active Tectonic Research Group (ATAG), 16–17 Nowember 2000, Eskişehir,* Abstracts, 4–5 [in Turkish].

KOÇYIĞIT, A., ÜNAY, E. & SARAÇ, G. 2000a. Episodic graben formation and extensional neotectonic regime in west Central Anatolia and the Isparta Angle: a key study in the Akşehir-Afyon graben, Turkey. *In*: BOZKURT, E., WINCHESTER, J.A. & PIPER, J.D.A. (eds), *Tectonics and Magmatism in Turkey and the Surrounding Area*. Geological Society, London, Special Publications 173, 405–421.

KOÇYIĞIT, A., YUSUFOĞLU, H. & BOZKURT, E. 1999. Evidence from the Gediz graben for episodic two-stage extension in western Turkey. *Journal of the Geological Society, London* 156, 605–616.

KOERI 2006. *Records of Earthquakes in Period of 1900-2006 in Turkey*. Kandilli Observatory Laboratory and Earthquake Research Institute, İstanbul, Turkey.

LAMOTTE, D.F., POISSIN, A., AUBOURG, C. & TEMIZ, H. 1995. Chevauchements post-tortoniens vers l'ouest puis vers le sud au coeur de l'angle d'Isparta (Taurus, Turquie). Conséquences géodynamiques [Post-Tortonian westward and southward thrusting in the core of the Isparta re-entrant (Taurus, Turkey): Geodynamic implications]. *Bulletin de la Société Géologique de France* 166, 59–67.

MCKENZIE, D.P. 1972. Active tectonics of the Medirerranean Region. *Geophysical Journal of the Royal Astronomical Society* 30, 109–185.

MCKENZIE, D.P. 1978. Active tectonics of the Alpine-Himalayan belt: the Aegean and surrounding regions. *Geophysical Journal of the Royal Astronomical Society* 55, 217–254.

NEMEC, W., KAZANCI, N. & MITCHELL, J.G. 1998. Pleistocene explosions and pyroclastic currents in west-central Anatolia. *Boreas* 27, 311–332.

ÖCAL, N. 1968. *Seismicity and Earthquake Map of Turkey: An Earthquake Catalog From 1950 to 1960*. Kandilli Rasathanesi Publications no. 7, İstanbul.

OKAY, A. 1984. Kuzeybatı Anadolu'da yer alan metamorfik kuşaklar [Metamorphic belts in northwestern Anatolia]. *İhsan Ketin Symposium, 20–21 February 1984, Ankara, Proceedings*, 83–92.

OKAY, A. 1989. Denizli'nin güneyinde Menderes Masifi ve Likya naplarının jeolojisi [Geology of Menderes Massif and Lycian Nappes to the south of Denizli]. *Mineral Research and Exploration Institute (MTA) of Turkey Bulletin* 109, 45–58 [in Turkish, with English abstract].

ÖZALAYBEY, S., ERGIN, M., BIÇMEN, F., TAPIRDAMAZ, C., YÖRÜK, A., TARANCIOĞLU, A., SAATÇILAR, A. & ERGINTAV, S. 2000. *Denizli Bölgesindeki Mikrodeprem Etkinliğinin Araştırılması [Survey of Microearthquakes Activity in Denizli Region]*. TÜBİTAK, Marmara Research Center, Earth Science Research Institute, Project no. 5007102 [in Turkish with English abstract, unpublished].

ÖZGÜL, N. 1976. Torosların bazı temel jeoloji özellikleri [Some basic geological characteristics of Taurides]. *Geological Society of Turkey Bulletin* 19, 65–78 [in Turkish with English abstract].

ÖZTÜRK, A. 1982. Tectonics of Dinar-Sandıklı-Işıklı region. *Communications of Faculty of Science, University of Ankara* 25, 1–58.

PATON, S. 1992. Active normal faulting, drainage patterns and sedimentation in southwestern Turkey. *Journal of the Geological Society, London* 149, 1031–1044.

PINAR, N. & LAHN, E. 1952. *Türkiye Depremleri İzahlı Kataloğu [Explanatory Catalog for Turkish Earthquakes]*. Bayındırlık Bakanlığı, Yapı ve İmar İşleri Reisliği Publications no. 36, Ankara.

PIPER, J.D.A., GÜRSOY, H., TATAR, O., İŞSEVEN, T. & KOÇYIĞIT, A. 2002. Paleomagnetic evidence for the Gondwanic origin of the Taurides and rotation of the Isparta Angle, southern Turkey. *Geological Journal* 37, 317–336.

POISSON, A. 1977. *Recherches géologiques dans les Taurides Occidentales (Turquie)*. These d'etat; Univ. de Paris-Sud, Orsay.

POISSON, A., WERNLI, R., SAGULAR, E.K. & TEMİZ, H. 2003. New data concerning the age of the Aksu Thrust in the south of the Aksu valley, Isparta Angle (SW Turkey): consequences for the Antalya Basin and the Eastern Mediterranean. *Geological Journal* **38**, 311–327.

PRICE, S.P. & SCOTT, B 1991. Pliocene Burdur basin, SW Turkey: tectonics, seismicity and sedimentation. *Journal of the Geological Society, London* **148**, 345–354.

PRICE, S.P. & SCOTT, B. 1994. Fault-block rotations at the edge of a zone of continental extension, southwest Turkey. *Journal of Structural Geology* **16**, 381–392.

RICHARDSON-BUNBURY, J.M. 1996. The Kula volcanic field, western Turkey: the development of a Holocene alkali basalt province and the adjacent normal faulting graben. *Geological Magazine* **133**, 275–283.

ROBERTSON, A.H.F. 2002. Overview of the genesis and emplacement of Mesozoic ophiolites in the Eastern Mediterranean Tethyan region. *Lithos* **65**, 1–67.

ROBERTSON, A.H.F. & COMAS, M. 1998. Collision-related processes in the Mediterranean region – Introduction. *Tectonophysics* **298**, 1–4.

SARAÇ, G. 2003. *Türkiye Omurgalı Fosil Yatakları [Mammalian Fossils Localities in Turkey]*. Mineral Research and Exploration Institute (MTA) of Turkey Report no. **10609** [in Turkish, unpublished].

ŞENEL, M., SELÇUK, H., BILGİN, Z.R., ŞEN, A.M., KARAMAN, T., DINÇER, M.A., DURUKAN, E., ARBAS, A., ÖRÇEN, S. & BILGİ, C. 1989. *Çameli (Denizli)–Yeşilova (Burdur) ve Elmalı (Antalya) Dolayının Jeolojisi [Geology of the Çameli (Denizli)–Yeşilova (Burdur) and Elmalı (Antalya) Area]*. Mineral Research and Exploration Institute (MTA) of Turkey Report no. **9429** [in Turkish, unpublished].

ŞENGÖR, A.M.C.1987. Cross faults and differential stretching of hanging wall in the regions of low-angle normal faulting: examples from western Turkey. *In:* COWARD, M.P., DEWEY, J.F., HANCOCK, P.L. (eds), *Continental Extension Tectonics*. Geological Society, London, Special Publications **28**, 575–589.

ŞENGÖR, A.M.C. & YILMAZ, Y. 1981. Tethyan evolution of Turkey: A plate tectonic approach. *Tectonophysics* **75**, 181–224.

SEYİTOĞLU, G. & SCOTT, B. 1992. The age of the Büyük Menderes Graben (west Turkey) and its tectonic implications. *Geological Magazine* **129**, 239–242.

SEYİTOĞLU, G. & SCOTT, B.C. 1994. Late Cenozoic basin development in west Turkey – Gördes basin tectonics and sedimentation. *Geological Magazine* **131**, 631–637.

SEYİTOĞLU, G. & SCOTT, B.C. & RUNDLE, C.C. 1992. Timing of Cenozoic extensional tectonics in west Turkey. *Journal of the Geological Society, London* **149**, 533–538.

SICKENBERG, O., BECKER-PLATEN, J., BENDA, L., BERG, D., ENGESSER, B., GAZIRY, W., HEISSIG, K., HUNERMANN, K.A., SONDAAR, P.Y., SCHMIDT-KITTLER, N., STAESCHE, U., STEFFENS, P. & TOBIEN, H. 1975. Die Geliederung des höheren Jungtertiars und Altquartars in der Türkei nach Vertebraten und ihre Bedeutung für die internationale Neogene-Stratigraphie. *Geologischen Jahrbuch Reihe* **15**, 167 p.

SOYSAL, H., SIPAHİOĞLU, S., KOLÇAK, D. & ALTINOK, Y. 1981. *Türkiye ve Çevresinin Tarihsel Deprem Kataloğu (M.Ö. 2100 – M.S. 1900) [Historical Earthquakes Catalog of Turkey and its Environment (2100 B.C. to 1900 A.D.)]*. TÜBİTAK Project no. **TBAG–314** [in Turkish, unpublished].

SÖZBİLİR, H. 2002. Revised stratigraphy and facies analysis of Paleocene–Eocene supra-allochthonous sediments (Denizli, SW Turkey) and their tectonic significance. *Turkish Journal of Earth Sciences* **11**, 87–112.

TAYMAZ, T. 1993. The source parameters of the Çubukdağ (W Turkey) earthquake of 1986 October 11. *Geophysical Journal International* **113**, 260–267.

TAYMAZ, T. & PRICE, S. 1992. The 1971 May 12 Burdur earthquake sequence, SW Turkey: a synthesis of seismological and geological observations. *Geolphysical Journal International* **108**, 589–603.

TAYMAZ, T. & TAN, O. 2001. Source parameters of June 6, 2000 Orta-Çankırı (Mw = 6.0) and December 15, 2000 Sultandağ-Akşehir (Mw=6.0) earthquakes obtained from inversion of teleseismic P- and SH body-waveforms. *Symposia on Seismotectonics of the Northwestern Anatolia–Aegean and Recent Turkish Earthquakes, İstanbul, Extended Abstracts*, 96–107.

TAYMAZ, T., TAN, O., ÖZALAYBEY, S. & KARABULUT, H. 2002. Source characteristics of February 3, 2001 Çay-Sultandağ earthquake (Mw=6.5) sequence in SW Turkey: a synthesis of seismological observations of body-waveforms, strong motions, and aftershock seismicity survey data. *1st International Symposium of İstanbul Technical University, the Faculty of Mines on Earth Sciences and Engineering, İstanbul, Turkey, Abstracts*, p. 60.

TEMİZ, H., POISSON, A., ANDREIEUX, A.A. & BARKA, A.A. 1997. Kinematics of the Plio–Quaternary Burdur-Dinar cross-fault system in SW Anatolia (Turkey). *Annales Tectonica* **XI**, 102–113.

TEMİZ, H., GÜRSOY, H. & TATAR, O. 1998. Kinematics of Late Pliocene–Quaternary normal faulting in the southeastern end of the Gediz Graben, Western Anatolia, Turkey. *International Geology Review* **40**, 638–646.

TEKELİ, O., AKSAY, A., ÜRGÜN, B.M. & IŞIK, A. 1984. Geology of the Aladağ Mountains. *In*: TEKELİ, O. & GÜNCÜOĞLU, M.C. (eds), *Geology of the Taurus Belt*. International Symposium, 26–29 September 1983, Ankara, Turkey, Proceedings, 143–158.

TOKÇAER, M., AGOSTINI, S. & SAVAŞÇIN, Y. 2005. Geotectonic setting and origin of the youngest Kula volcanics (Western Anatolia), with a new emplacement model. *Turkish Journal of Earth Sciences* **14**, 145–166.

USGS 2007. United States of Geological Survey–National Earthquake Information Center.

UYSAL, Ş., DUMONT, J.F. & POISSON, A.1980. *Batı Toros Platformları [Western Tauride Platforms]*. Mineral Research and Exploration Institute (MTA) of Turkey Report no. **80/4-13** [in Turkish, unpublished].

WALDRON, J.W.F. 1984. Structural history of the Antalya Complex in the 'Isparta Angle', southwest Turkey. *In*: DIXON, J.E. & ROBERTSON, A.H.F. (eds), *The Geological Evolution of the Eastern Mediterranean*. Geological Society, London, Special Publications **17**, 273–278.

WELLS, D.L. & COPPERSMITH, S.R. 1994. New empirical relationships among magnitude rupture length, rupture width, rupture area and surface displacement. *Bulletin of Seismological Society of America* **84**, 974–1002.

WESTAWAY, R. 1990. Block rotation in western Turkey. 1. Observational evidence. *Journal of Geophysical Research* **95**, 19857–19884.

WRIGHT, T.J., PARSONS, B.E., JACKSON, J.A., HAYNES, M., FIELDING, E.J., ENGLAND, P.C. & CLARKE, P.J. 1999. Source parameters of the 1 October 1995 Dinar (Turkey) earthquake from SAR interferometry and seismic bodywave modeling. *Earth and Planetary Science Letters* **172**, 23–37.

YAĞMURLU, F. 1991. Yalvaç-Yarıkkaya Neojen havzasının stratigrafisi ve depolanma ortamı [Stratigraphy and depositional setting of the Yalvaç-Yarıkkaya Neogene basin]. *Geological Society of Turkey Bulletin* **34**, 9–19 [in Turkish with English abstract].

YETİŞ, C. 1987. Facies and environmental aspects of the fluvio-lacustrine Oligo–Miocene deposits in the Çameli (Niğde) area. *Geological Society of Turkey Bulletin* **30**, 1–8 [in Turkish with English abstract].

YILMAZTÜRK, A. & BURTON, P. W. 1999. Earthquake source parameters as inferred from the body waveform modeling, southern Turkey. *Geodynamics* **27**, 469–499.

YILMAZ, Y., GENÇ, Ş.C., GÜRER, F., BOZCU, M., YILMAZ, K., KARACIK, Z., ALTUNKAYNAK, Ş. & ELMAS, A. 2000. When did the western Anatolian grabens begin to develop? *In*: BOZKURT, E., WINCHESTER, J.A. & PIPER, J.D.A. (eds), *Tectonics and Magmatism in Turkey and the Surrounding Area*. Geological Society, London, Special Publications **173**, 353–384.

YUSUFOĞLU, H.1998. *Paleo- and Neo-tectonic Characteristics of the Gediz and Küçük Menderes Grabens in West Turkey*. PhD Thesis, Middle East Technical University, Geological Engineering Department [unpublished].

ZANCHI, A. & ANGELIER, J. 1993. Seismotectonic of western Anatolia: regional stress orientation from geophysical and geological data. *Tectonophysics* **222**, 259–274.

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