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ORIGINAL PAPER

Layer-parallel shortening and related structures in zones undergoing active regional horizontal extension

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Abstract Folds and thrust faults formed by layer-parallel shortening coaxial with extensional structures such as normal dip-slip faults and ductile necking structures with orthorhombic fabric symmetry are usual, but little-recognised structures formed within normal dip-slip shear zones bounding rifts. They are generated because of the shear distribution in a zone of progressive deformation and may be later extended and disrupted depending on which part of the strain ellipsoid they may be located. We here describe folds and thrust faults from the southern margin of the Alaşehir Rift in western Turkey as an opportunity to discuss the properties of *pure extension–related structures formed by layer-parallel shortening*. Such structures are more commonly generated during the early stages of rifting, when deformation rates are slow and the shear zones broader than those forming later in the life of a rift when strain rates are usually higher. Such structures have commonly been mistaken for witnesses documenting regional episodes of shortening rather than as integral parts of the extensional structures forming rifts. Not all layer-parallel shortening-related structures therefore indicate regional shortening. We plead that hasty statements concerning the meaning of geological structures at all scales be avoided before a thorough understanding of bulk strains that have affected a region are properly understood.

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Introduction

In structural geology, the structures, by themselves, are the least important things. What is important is the *total strain* and the *strain path* that has deformed the rock body in which the structures are located, because the geologist is interested not only in the *present architecture* of rock masses, but also in their *history of construction*. One can read the total strain by considering the structures belonging to a *strain history* in a given *volume* of rock, and it is only from the viewpoint of their role, as a group, in reflecting that strain history, that structures are of any importance. Kinematic indicators, which are individual structures, are equally unimportant, unless they inform us about the strain path of the studied rock body. Inattention to bulk strain history, that is, to *progressive deformation in volume*, has led to many egregious errors not only in the distant past of structural geology (e.g. Stille 1913, 1925), but also more recently (see the critique in Flinn 1994) (Fig. 1).

The purpose of this paper is to describe a series of medial Miocene folds and thrust faults from the central part of the Menderes Massif in western Turkey as an excuse to discuss a certain environment of folding and thrusting resulting from local shortening within extensional tectonic regime. Dewey (2002, 2004) earlier described how transtension may generate shortening structures that mimic orogenic structures and lead to the erection of 'spoof orogenies'. Our purpose here is to show that: (1) not every fold or thrust fault resulting from real shortening necessarily indicates a tectonic regime of general horizontal shortening or even transtension and (2) there are ways to

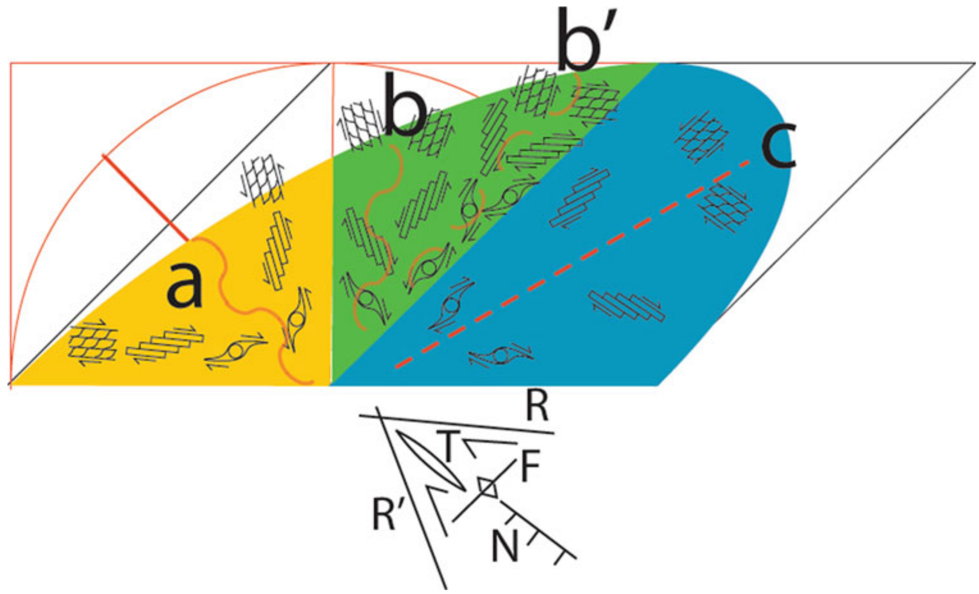


Fig. 1 Strain ellipse (only the upper half is shown) in a dextral shear zone (Ramsay 1967) showing not only the strain, but some possible structures that may indicate localised strains within the deforming body of rock. Such localised zones may form and die as the rock body continues its deformation. The *red circle* and the square represent the undeformed state of the rock body illustrated. As the shear deformation begins (here represented as plane strain for graphic simplicity), the strain ellipse indicates how certain volumes of rock will behave. In the *yellow* region, all lines connecting the periphery of the ellipse with the former centre of the *circle* will deform by shortening at all times. This shortening may be taken up by a variety of structures. Here, we chose to indicate R (=Riedel) and R′ (=anti-Riedel) shears only. Along such shear zones, a variety of kinematic indicators may develop depending on the mineralogy and the rheology of the rocks. As the R′ shears develop earlier (Tchalenko 1970), if a shear zone becomes fossilised at that stage, a geologist mapping only the kinematic indicators forming along the R′ shears may get a completely incorrect view of the orientation and the slip along the major shear zone. In the *green* sectors, all *lines* will first shorten and then lengthen. Here, the kinematic indicators will give the impression of having reversed their sense, possibly giving the impression of a major tectonic regime change in the area, although all the geologist sees would in reality be the continuation of the same regime. In the *blue* zone again, one may be misled if only one set of kinematic indicators happen to predominate. The kinematic indicators are useful if and only if the geologist is aware of their misleading roles unless a total strain picture of the rocks being mapped is carefully considered in a sufficiently large area. This is often not the case (Flinn 1994)

distinguish folds formed by shortening in overall horizontally shortening regimes from the folds formed by shortening in overall purely horizontally extensional regimes, although their morphologies may be identical.

In order to forestall any misunderstanding, we point out that earlier in the literature folds have been described from many environments of extension, such as folding perpendicular to extension direction (e.g. Şengör 1987; Chauvet and Séranne 1994; Lévy and Jaupart 2011), folding over fault ramps (e.g. Howard and John 1997; Faulds et al. 2002; Khalil and McClay 2002; Resor 2008), folding at the toes of large-scale extensional structures along Atlantic-type continental margins (for an excellent review, see Rowan et al. 2004), such as the Perdido Fold belt in the Gulf of Mexico at the toe of the creeping supra-Louann secession (e.g. Fiduk et al. 1999; Trudgill et al. 1999) or the Niger Delta deep-water fold and thrust belt (Bilotti and Shaw 2005; Maloney et al. 2010), gravity collapse-related folds (Harrison and Falcon 1934) or folding owing to unequal isostatic motions (e.g. Martinez-Martinez 2002). However, none of these are a necessary consequence of extension-related simple shear. Earlier, folds and thrusts from a number of rifts (Willis 1923, 1936; also see Carey 1958) led to the idea of ramp valleys (Willis 1923), although the structures reported in these earlier publications have been later discredited and/or explained by dip-slip normal fault-related folding.

Folds formed by shortening in a purely extensional regime

In regions of tectonic stretching, extension is taken up by a variety of means (Fig. 2a). If the rocks undergoing extension possess a high degree of ductility, the extension may be taken up by fairly homogeneous thinning and stretching, forming an orthorhombic fabric in the rocks. With decreasing ductility, discrete shear zones may take up the extension, creating tabular and/or scalloped zones of monoclinic fabric. As the rocks become brittle, faults begin to serve as the extensional structure, and finally, where there is inconsiderable confining pressure, simple tension joints and tension fissures (*gjá* fissures of the Icelandic literature) may be dominant. In the same temperature/pressure/fluid circulation environment, sudden change of strain rate (as during earthquakes) may break the rocks otherwise deforming in a ductile manner along broader simple shear zones. All of the structures named above are in reality formed from a family of structures on different scales (Fig. 2a). We confine our attention to discrete shear zones, as they are by far the most widespread forms of structure families encountered in zones of deformation, and particularly in zones of extension (e.g. see Fig. 2b). It has to be borne in mind, however, that the thickness of such shear zones may range from centimetre-scale to kilometre-scale.

Figure 3a and a′ illustrates a case of *simple shear* with a shear angle of 45° clockwise applied to a square with inscribed circle. Figure 3b shows a 50 % shortening, that is, *pure shear*, in a vertical direction of the same initial square + circle and then Fig. 3b′ shows the same shortened object sheared again by 45° by *simple shear*. This is equivalent to a single simple + pure shear deformation ('transpression' in zones of strike-slip). In Fig. 3c, the square of Fig. 3a is also vertically shortened, but not allowed to extend in either of the other dimensions, resulting in net volume loss (that may be accomplished by a variety of means such as stylolite or solution cleavage formation). Figure 3c′ shows what happens when the object of Fig. 3c is sheared by 45° clockwise. The result is different from the strain depicted in Fig. 3a′, b′. In Fig. 3d, the object shown in Fig. 3a is first sheared clockwise by 45° and then vertically shortened by pure shear 50 %. Note here too that the resulting strain is different from the ones earlier depicted. Finally in Fig. 3e, the object is again sheared clockwise by 45° and then shortened vertically 50 % but not allowed to extend in either of the other directions. The result is again volume loss, but the resulting strain is again different from that shown in Fig. 3c. *These differences demonstrate the great importance of strain history* (see also Schmidt 1932, chapter IV, Means 1976, Chap. 24). These are well-known concepts ever since their introduction into geology by Bruno Sander (1930, 1948) and Walter Schmidt (1932) and popularisation and extension by Turner and Weiss (1963) and Ramsay (1967), but the implications of strain history are seldom considered in regional geological work, because strain history is difficult to establish. Consequently, its importance in interpreting large-scale tectonics is commonly ignored.

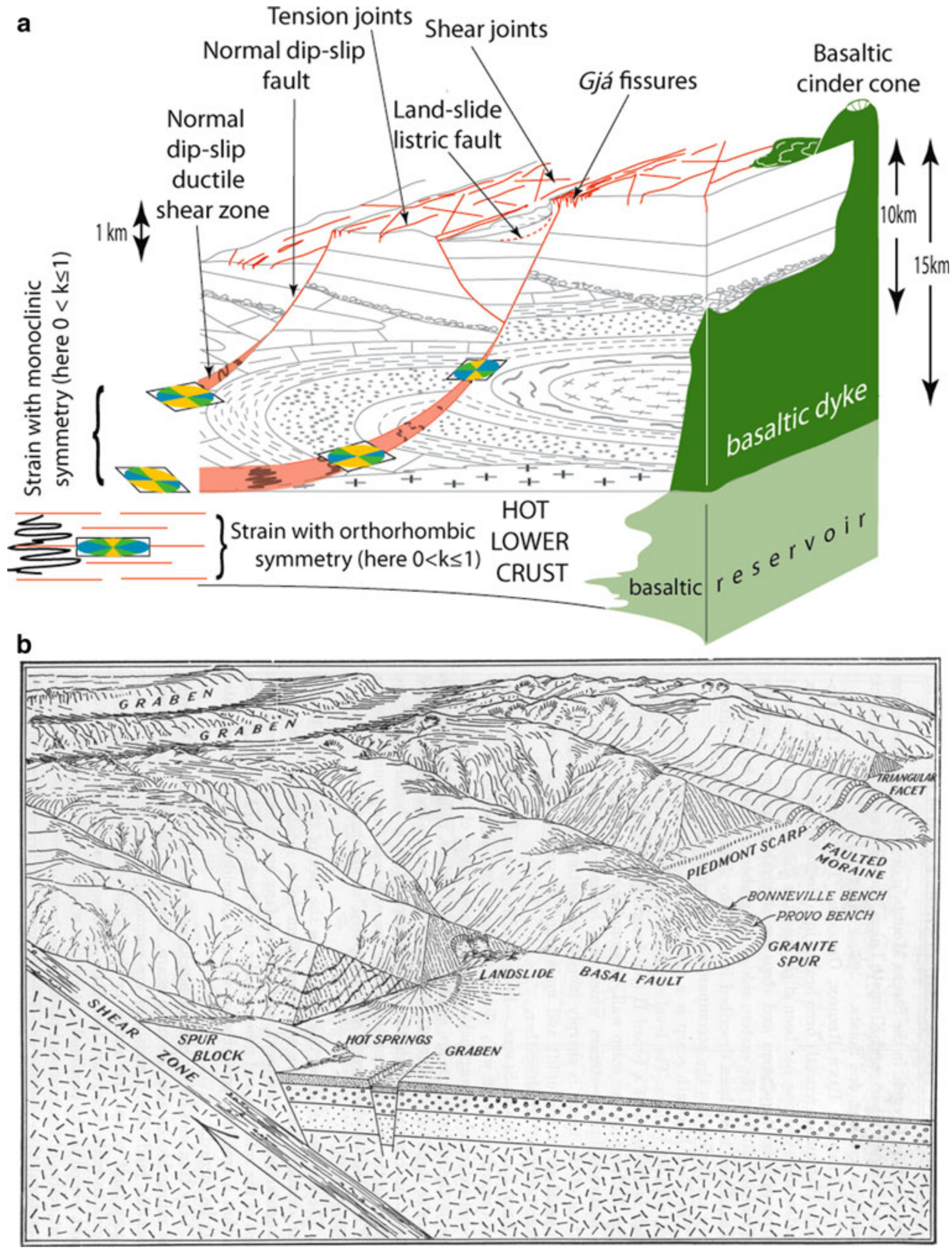
In zones of stretching, the shear zones taking up extension are zones either of simple shear or simple + shortening pure shear because of the role of gravity; that is, in such regions, one faces either case illustrated in Fig. 3a′, c′ (remember that the structural implications of volume loss are neglected in the following discussion), because the overburden commonly leads to fluid loss or even rock solution that diminishes volume. We now treat the cases shown in Fig. 3a, a′, b, b′ because they are the commonest cases met by the mapping geologist.

Under these conditions, Fig. 4 illustrates the strain ellipse that develops in the case of Fig. 3a′ as first pointed out by Ramsay (1967). Figure 5 shows three cases, where shear zones have hades of 30°, 45° and 80°. The following discussion is summarised from Flinn (1962, corrections in Flinn 1965) and Ramsay (1967).

Figure 5a illustrates the 30°-right-hading shear zone with a number of 'competent' layers through it. Note that the grey competent layers may be folded into the black undulating layers, of which we show only two examples. In fact, any competent layer in the yellow zones will undergo more or less folding or other sorts of layer-parallel structure generation depending on their orientation with respect to the strain ellipse. Pink layers, whose orientations place them into the green regions of the strain ellipse, will first be folded and then stretched creating the red 'folded boudins' or thrusts that later act as, or sliced by, normal dip-slip faults. If encountered in an isolated state, such boudins in a region of pure extension may be misinterpreted to have disrupted a previous phase of general shortening in the area, whereas both the initial folds and the subsequent boudins will be simple products of the very same extensional regime. By contrast, the yellow lines, falling into the blue regions of the strain ellipse, will undergo only extension and possible boudining or creating pinch-and-swell structure, depending on the ductility contrast between the layers.

Figure 5b, c show that layers of very different orientations in space may undergo folding in diverse orientations, depending on the hade of the shear zone (remember the plane strain assumption!). In some of simple shear normal dip-slip zones, for layers to be folding, their dips must be more than 45°, but in fairly shallowly hading shear zones (<25°), quite shallowly dipping layers may also be folded throughout the history of the normal dip-slip shear zone.

Figure 6 shows the case of 'squashed' ('transpressional') normal dip-slip shear zones (simple + pure shear + volume loss, i.e. the case in Fig. 3b′!). The colour coding is the same as in Fig. 4 (remember that we neglect volume loss!). Note here that zone 3, in which all layers experience continuous layer-parallel shortening during the shearing, has increased at the expense of other two zones. This means that as the vertical shortening increases (presumably with depth, assuming geotherm behaviour being usual), more layer-parallel shortening of planar features is likely to take place. *In other words, as one descends down the hade*



of a shear zone and in the direction of increasing burial metamorphism, one is likely to see more folding of planar features in zones of pure regional extension. This has some interesting implications for the structural interpretation of high-grade metamorphic terrains. Dewey (2002, 2004) has shown that not all 'orogenic-looking' regions are really orogenic, but results of extensional keirogeny. Our considerations extend his conclusions into areas even of pure taphrogeny.

If volume loss is not allowed, the strain ellipse in Fig. 6 may also show extension in the dimension perpendicular to the extension direction. In such a situation, chocolate-boudinage-like structures may develop. Figure 7 shows three cases, where shear zones have hades of 25°, 45° and 60° as in Fig. 5.

Extension-related folding in the Alaşehir Rift

The Alaşehir Rift is one of the most spectacular structural elements of the N–S extension in western Turkey and forms the boundary between the northern and central parts

◀**Fig. 2** **a** The main family of structures forming in a purely extensional regime in the crust. Near the surface, where confining pressure is unimportant, tension and shear joints may form and some link with one another dividing the rock into blocks of various sizes. In certain cases of intense and rapid extension (as along mid-oceanic spreading centres), some tension joints may open up to form large tension gashes called *gjá* fissures in Iceland, which may extend for kilometres along their strike. In some cases, lava may fill such fissures creating dykes. It is now becoming increasingly more widely recognised that intruding magma helps in the formation of the fissures by hydrolic fracturing into which the dykes get rapidly emplaced. As confining pressure increases, inclined normal dip-slip faults form. The major, crustal-extension-controlling faults may normally extend down to some 5–7 km in continental regions of active and rapid extension (e.g. the Aegean region in western Turkey and Greece, the Basin-and-Range region in the USA), although they slip along single planes at 8–10 km presumably only during very rapid strains, such as develop during earthquakes. Normally, already above such depths, broader and more ductile shear zones begin to take over the displacement along discrete faults farther up. The shear zones must be listric at depth as the rotation around horizontal axes of major crustal blocks indicate (domino style faulting probably does not extend very deep into the crust; even if it does, at least one major listric fault is necessary to cause rotations, be it active all at once, or in segments and if one can form, so can many). Below the mid-crust orthorhombic homogeneous stretching is probably the normal mode of extension, although what is mid-crust changes from region to region depending on the crustal thickness and the thermal history of the area. In East Africa for instance, seismic, planar faulting extends down to 30 km, whereas in western Turkey, it goes down rarely below 10. This figure also shows the orientation and character of the strain ellipse along the shear zones that follow down from the major normal dip-slip faults. A variety of structures will form along such shear zones depending on the pre-existing anisotropy in rock texture. Accordingly, local structures ranging from purely extensional to purely shortening-related may be seen along the very same shear zone. The size of such structures will be commensurate with the width of the shear zone in which they form. **b** Lobeck's (1939) block diagram of the Wasatch Front, showing the morphology to be controlled by a broad shear zone rather than a single normal fault

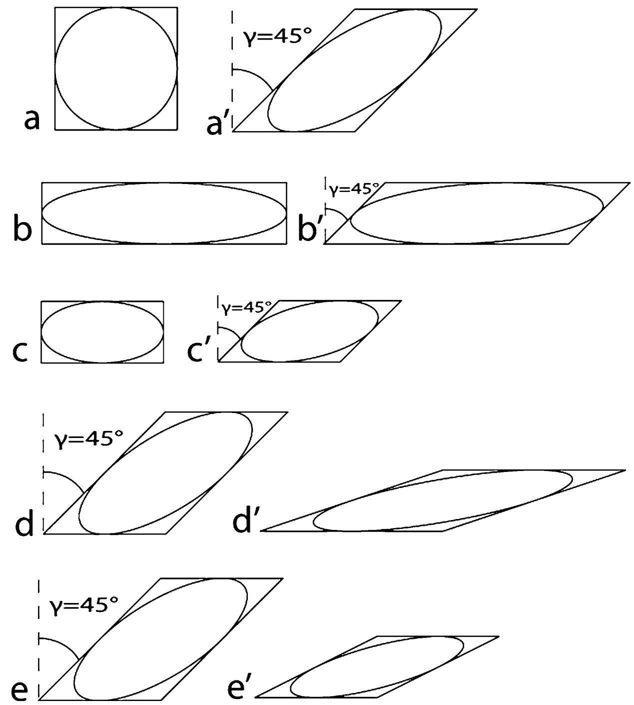


Fig. 3 Simple and simple + pure ('transpressional') shear with and without volume loss. **a** shows a square with an inscribed *circle* representing an unstrained body of rock. **a′** is the result of simple shear with $\gamma = 45°$ affecting the same body of rock. In **b**, the body of rock in **a** is first shortened vertically by plane strain by 50 % with respect to its original thickness. In **b′**, the shortened body is now deformed by simple shear with $\gamma = 45°$. In **c**, the same body as shown in **a** is again shortened vertically, but no stretching in either of the other dimensions is allowed resulting in net 50 % volume loss. It is then sheared by 45° in **c′**. Notice that the resulting strains in **b′** and **c′** are not identical. In **d**, the *square* shown in **a** is first sheared 45°. The resulting strain is again different from the previous one. Finally, the sheared object shown in **d** is vertically shortened by 50 %, with no allowance of extension in either of the other dimensions (**e′**). Again we obtain a totally different strain from the previous strains! Therefore, the sequence of deformations, that is, strain history, is important in generating the final strain

of the Menderes Massif (Fig. 8). It commenced as a north-facing[1] half-rift with an active southern margin during the early to possibly mid-Miocene[2]; then, it evolved into an asymmetric rift as a result of younger subordinate post-Miocene faulting at its northern margin, thus maintaining its northerly facing. The early history of the Alaşehir Rift involved the exhumation of the northernmost edge of the central part of the Menderes Massif in the footwall of a now inactive, low-angle, north-hading normal fault (a lag or an *extensional* detachment fault, called the Gediz Detachment) at the southern margin of the rift, while the hanging wall formed the site of Miocene lacustrine and fluvial deposition. High-angle faults chopped up the extensional detachment and caused back-tilting of the resulting blocks. They form a block staircase descending down to the rift with northward younging of the faults (Fig. 9).

The rift fill (>3,000 m: Çiftçi and Bozkurt 2010) is predominantly exposed along the southern margin of the rift and comprises Miocene to Recent continental clastic sedimentary rocks and very weakly consolidated sediments. It consists of two groups of rocks, separated by an angular unconformity: (1) Miocene formations (Alaşehir, Çaltılık and Gediz) exposed at the southern margin only and (2) post-Miocene formations (Kaletepe and Bintepeler at both margins) and overlying Quaternary alluvium (Fig. 9). The angular unconformity separating the two groups marks a change in the rift architecture from a

[1] For the concept of 'rift facing', see Şengör (2011a).

[2] The age of inception of the rift depends on the dating of the Alaşehir Formation (see below). The evidence rests entirely on palynomorph fossils, and the best bracket that can now be given is early to possibly medial Miocene (İztan and Yazman 1991; Ediger et al. 1996).

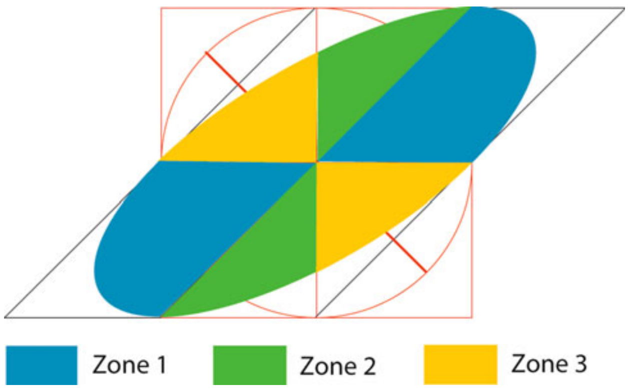


Fig. 4 The strain ellipse showing the case illustrated in Fig. 3a′. *Zone 1* represents a region in which all lines connecting the periphery of the ellipse to the original centre of the reference *circle* lengthen at all times during the deformation. *Zone 2* is the region in which all such lines first shorten and then lengthen. Finally, *Zone 3* is a zone in which all *lines* between the periphery of the ellipse and the centre of the reference *circle* shorten at all times during the deformation

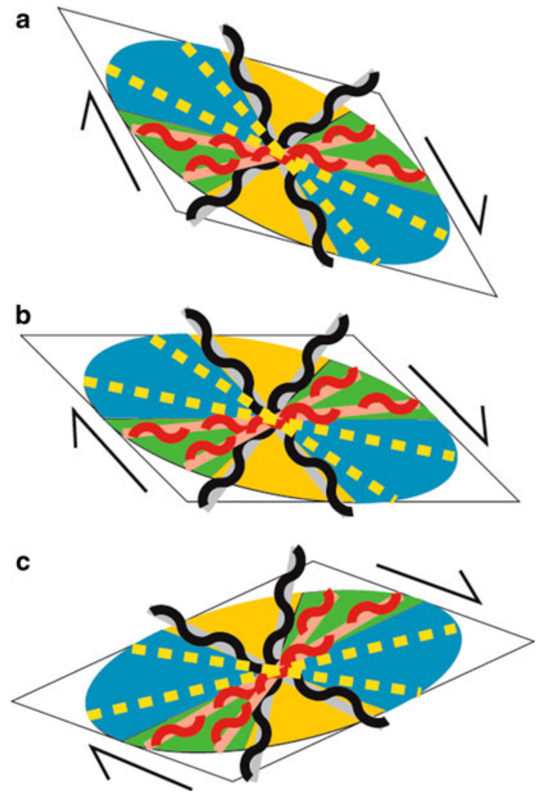


Fig. 5 Three *parallelograms* representing shear zone segments with inscribed strain ellipses and planar features (such as beds, dykes or foliations) in various orientations. **a** represents a shear zone hading 30° to the *right*. In it, almost all steeply dipping beds (>45°) (or other planar features capable of buckling) experience layer-parallel shortening and will fold (as shown in the figure) or shorten otherwise (thrusting, thickening, layer-perpendicular stylolite or solution cleavage formation, etc.) throughout the regional extension. Layers that are *horizontal* or near will first shorten and then stretch creating boudined fold trains (as shown in the figure) or thrusts that later act as normal faults, etc. Those that have intermediate dips to the right will stretch throughout their history. The shear zone in **b** hades 45° to the *right*. In it, beds (or other planar features) dipping more steeply than 45° in either direction will experience shortening. Planar features dipping about 45° to about 0° gently to the *left* will first shorten and then lengthen. Those dipping less steeply to (up to 45°) the *right* will stretch throughout the deformation. Finally in **c**, the shear zone hades some 75° to the *left*. In it, planar features dipping about 75° or more steeply to the *left* and those dipping about 45° or more to the right will experience shortening throughout the regional stretching. Planar features with dips less than about 35° will stretch, and those dipping between 35 and 75° to the left will be first shortened and then stretched. This figure also shows the critical importance of the pre-existing fabric *and* the dip of the shear zones affecting the area in any extensional deformation in creating diverse and mutually incompatible structures that may lead to much confusion in deciding their origin if not properly considered

Miocene north-facing half-rift configuration to post-Miocene, still north-facing asymmetric rift (Çiftçi and Bozkurt 2009a, 2010). The stratigraphy of the rift has been variously interpreted (e.g. Cohen et al. 1995; Emre 1996; Seyitoğlu and Scott 1996; Koçyiğit et al. 1999; Sarıca 2000; Yılmaz et al. 2000; Gökten et al. 2001; Seyitoğlu et al. 2002; Sözbilir 2002; Bozkurt and Sözbilir 2004; Öner and Dilek 2011). The scheme we present here is the work of one of us with his research student (Çiftçi and Bozkurt 2008, 2009a, b), which developed in part in criticism of older works.

The Alaşehir Miocene sedimentary sequence commences with the Alaşehir Formation made up of a fining-upward succession, with polygenetic cobble/pebble conglomerates intercalated with pebbly sandstones at the base (İztan and Yazman 1991). The unit continues with a sand-rich facies and grades into bituminous paper shales and alternating thin sandstone and siltstone beds. The coarser sediments are interpreted as fault scree and alluvial fan deposits, the fine-grained detritals as lacustrine facies. This unit is early to mid-Miocene in age (İztan and Yazman 1991; Ediger et al. 1996). The poorly sorted sandstones, interbedded with conglomerates, siltstones and the mudstones of the Çaltılık Formation conformably overlie the Alaşehir Formation. Greenish grey colour and red staining of the outcrop surfaces are characteristic. In areas where the Alaşehir Formation does not exist, the Çaltılık Formation lies structurally above the metamorphic rocks of the Menderes Massif along a low-angle normal fault plane. The succeeding Gediz Formation is typically composed of red-coloured, massive and current-bedded polymictic conglomerates and sandstones with some siltstone and mudstone. Two northward-dipping normal faults juxtapose the Gediz Formation with the older formations and the Menderes Massif basement to the south and the Quaternary rift floor sediments to the north. The southern fault is

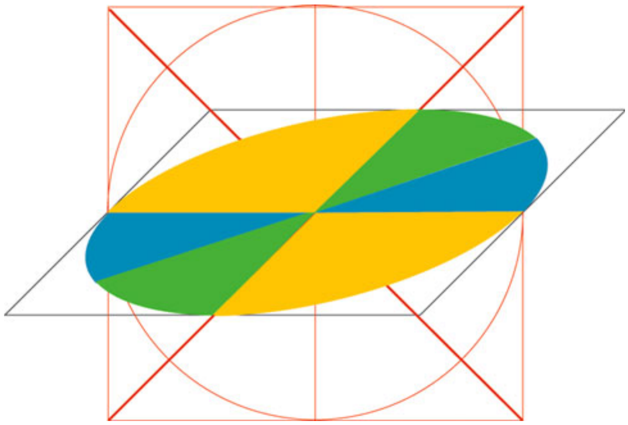


Fig. 6 Strain ellipse corresponding with the strain shown in Fig. 3c′. The colour convention for the strain zones is the same as that in Fig. 4

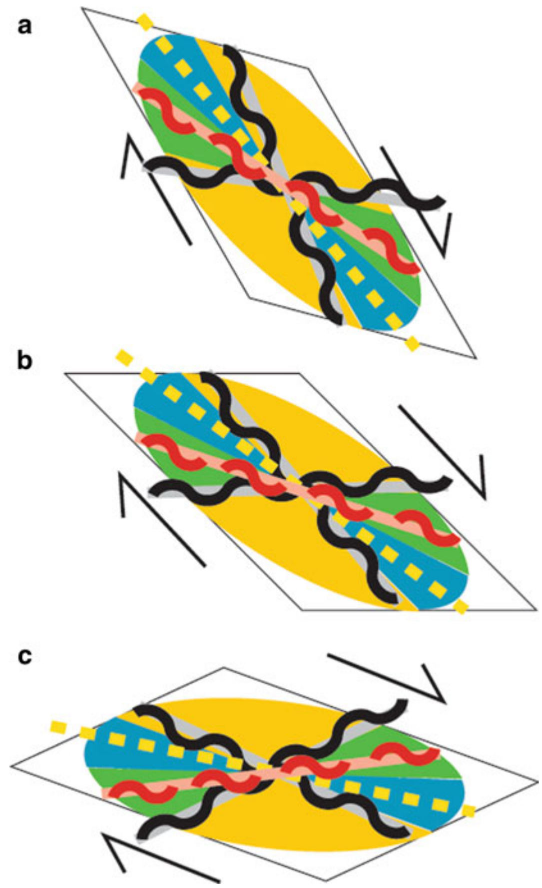


Fig. 7 This figure illustrates by means of three shear zone segments and the inscribed strain ellipses what would happen to variously dipping planar elements in these zones in cases of shear zone hades of 25°, 45° and 60° as in Fig. 5, but for the case of flattening + simple shear as in the case of Fig. 3c′. Notice the greater variability in dip of the layers that experience either continuous layer-parallel shortening or first shortening and then extension. In the following section, we describe examples of folding resulting from layer-parallel shortening during nearly pure north–south extension from the Alaşehir Rift in western Turkey, separating the northern and the central lobes of the Menderes Massif (e.g. Şengör et al. 1985; Şengör 1987; Seyitoğlu and Scott 1996; Bozkurt and Park 1997; Bozkurt 2003; Bozkurt and Sözbilir 2006)

interpreted as the master rift-bounding fault and forms the most prominent morphological feature of the Alaşehir Rift. The Miocene sedimentary rock units are generally tilted to the south (Fig. 9).

Unconformably above the Miocene units lie poorly bedded, poorly lithified polymictic conglomerates containing relatively minor sandstone and mudstone intercalations. They constitute the Kaletepe ('Fortress Hill') Formation along the southern horst block, and the Bintepeler ('Thousand Hills') Formation along the northern margin, of the Alaşehir Rift. The main difference between the two units is the predominance of marble fragments in the Bintepeler Formation (Fig. 9). A Pliocene or Plio–Pleistocene age for these rocks is assigned in the literature, and as their names imply, they form a highly differentiated, hilly topography.

Folds are common structural features in the Lower to Middle Miocene Alaşehir Formation from mesoscopic to macroscopic scales (Figs. 10, 11, 12, 13, 14, 15). The best and the most accessible outcrops exhibiting folded strata occur at two localities: (1) area to the SW of Osmaniye and (2) Kara Kirse villages (Fig. 9).

The best and relatively large-scale folded strata are found along a road cut (about 1.5 km in length) to the SW of Osmaniye (Fig. 10). There, numerous folds with various vergence and different degrees of closure, low- and (mostly) high-angle thrust faults and normal faults deform the thinly bedded and laminated bituminous paper shales and alternating thin sandstone and siltstone beds of the Alaşehir Formation (Figs. 11, 12). This section was studied by Çiftçi and Bozkurt (2008), and the information summarised below is taken from their paper. The folds are closely associated with southerly hading (commonly less than 45°) north-vergent thrust faults (Figs. 11a–d, 12). Seventeen measurements of striations from the fault planes (see Çiftçi and Bozkurt 2008, Fig. 10b; what the figure shows is of course shortening and *not* contraction as it is incorrectly stated there) are consistent with a predominantly dip-slip motion (at least for their latest motion), suggesting an approximately NNE–SSW-orientated shortening direction, very close to the direction of stretching deduced from the normal faults in the Alaşehir Rift. The open to tight folds show inconsistent vergence (but northerly overturning dominates) with angular to broadly curved, open hinges and gently to strongly inclined axial planes (Fig. 11e–f). They are intensely imbricated by many small-scale thrust faults (Fig. 13). Fold axes are generally

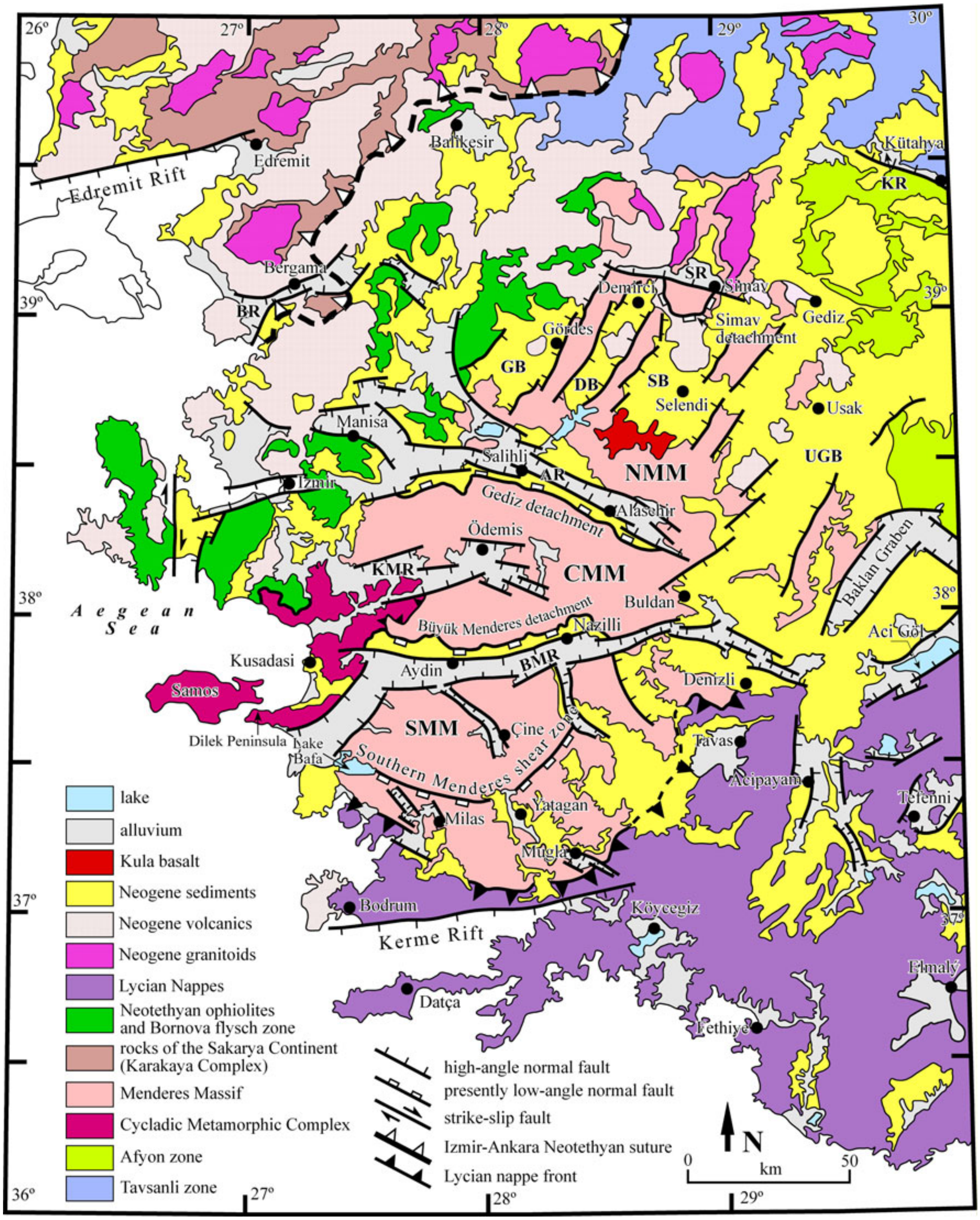


Fig. 8 Geological map of western Turkey showing the Menderes Massif and its subdivision (from Bozkurt 2007). *AR* Alaşehir Rift, *BR* Bakırçay Rift, *DB* Demirci Basin, *GB* Gördes Basin, *KG* Kütahya Rift, *SB* Selendi Basin, *SG* Simav Rift, *BMG* Büyük Menderes Rift, *CMM* Central Menderes Massif, *KMG* Küçük Menderes Rift, *NMM* Northern Menderes Massif, *SMM* Southern Menderes Massif, *UGB* Uşak-Güre Basin

orientated E–W and plunge mainly westward at angles <20° (see Çiftçi and Bozkurt 2008, Fig. 11). Several high-angle normal faults cut these folds (Figs. 10, 14), showing that normal faulting clearly postdated folding. On a centimetre- to decimetre-scale, normal faults parallel the larger ones (Fig. 14b). This relationship further suggests that the

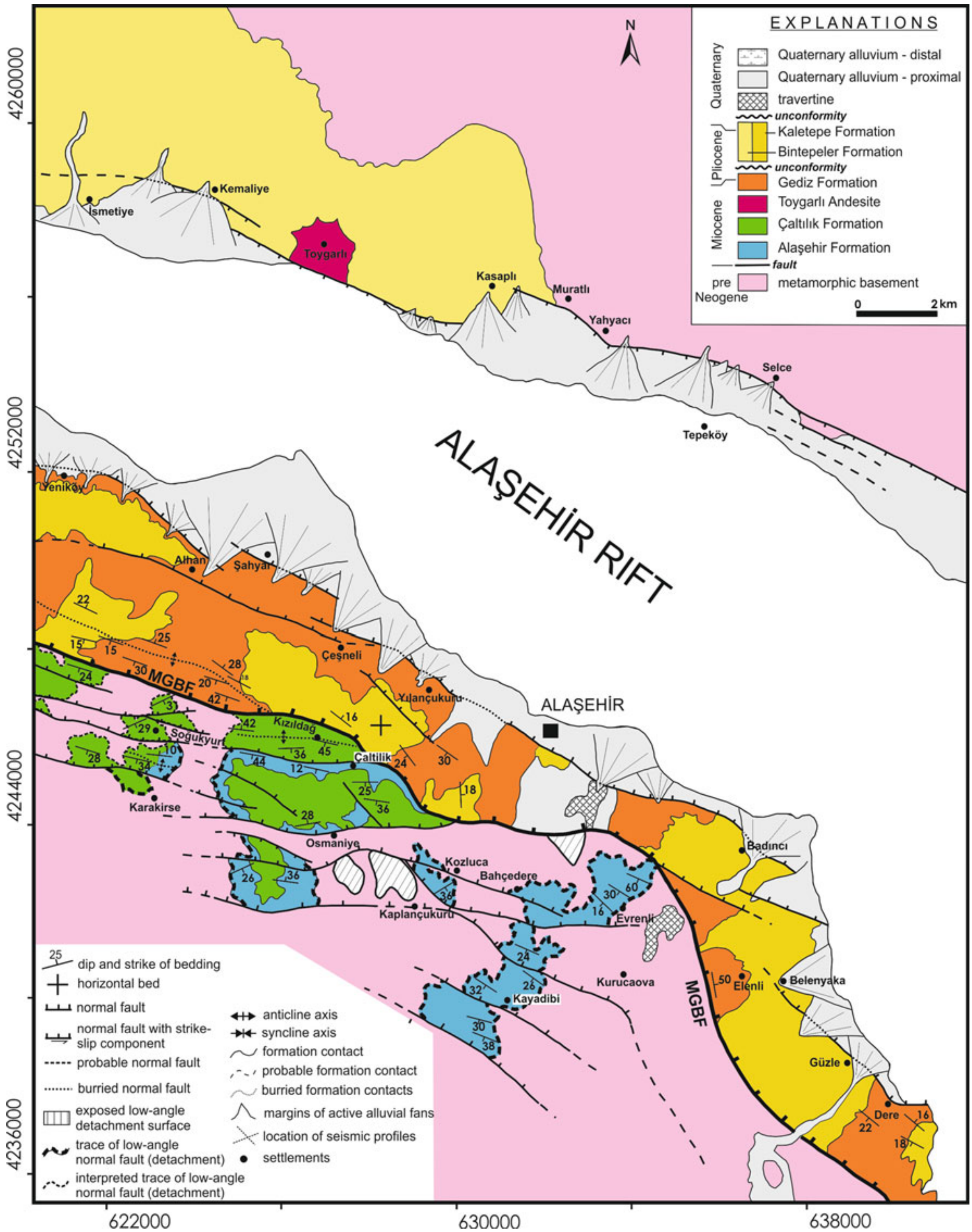


Fig. 9 Geological map of the Alaşehir Rift (from Çiftçi and Bozkurt 2008)

beds of the Alaşehir Formation were first shortened and then lengthened. In some beds, the field evidence (as described in this section; see also Çiftçi and Bozkurt 2008) permits inferring that the early thrust faults were reactivated as normal faults during the later stages of progressive deformation (Fig. 14a), although the later stretching does not nearly compensate the shortening achieved by folding and thrusting.

A word is perhaps necessary to state why we think the deformation was progressive; it has two reasons: (1) the

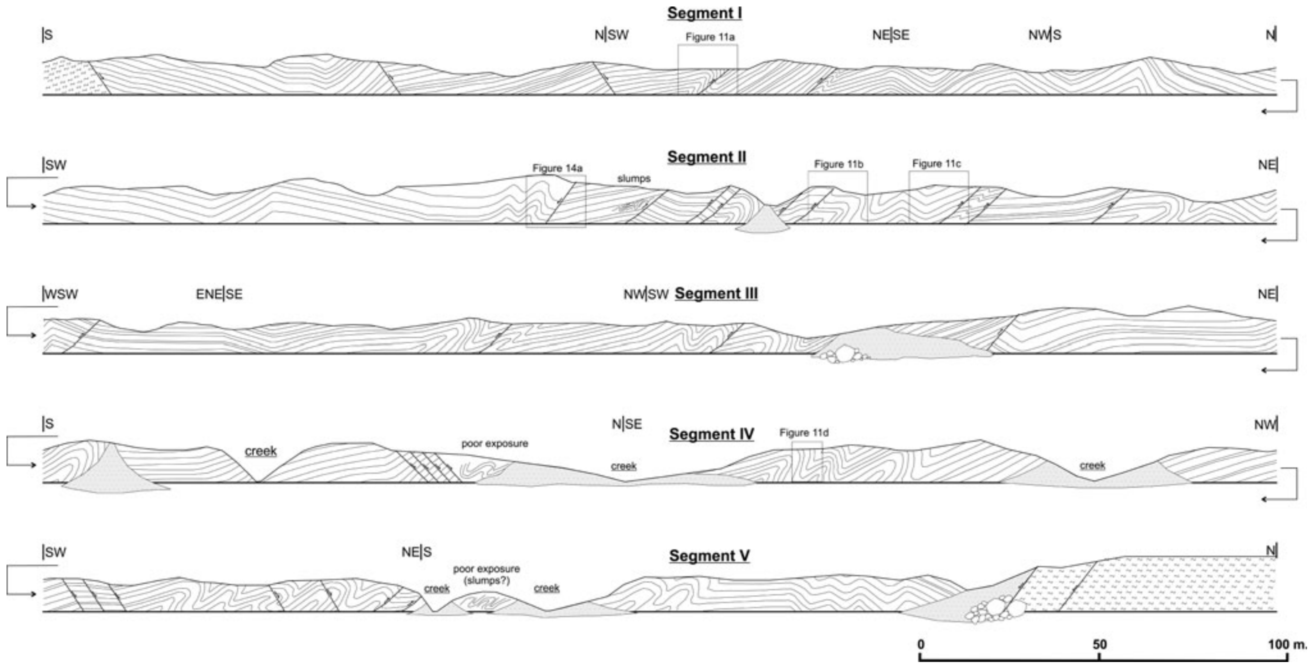


Fig. 10 Road cut cross section illustrating macroscopic folds and thrust faults deforming the Alaşehir Formation to the SW of Osmaniye (from Çiftçi and Bozkurt 2008)

structures of various generations are compatible with a single evolving strain history and they take up the same strain; (2) nowhere is there an episode of erosion between the superposed structures.

The second locality where folds occur is in Kara Kirse (*kara* is black; we think that *kirse* may be a bastardised form of *kilise*, i.e. from the Greek *εκκλησια*, church) village, where a low-angle normal fault separates the metamorphic rocks of the Menderes Massif below and the sediments of the Alaşehir Formation above (Fig. 15). The fault plane hades to the north by about 80°, whereas sediments dip by 30° to the south. Several, small-scale north-vergent asymmetric to overturned folds develop in the sediments immediately above the fault. They were later cut by small-scale normal faults. The metamorphics in the footwall are intensely sheared and crushed.

Discussion

Folds in the Alaşehir Rift had been noticed earlier and ignited a debate as to their origin. Koçyiğit et al. (1999) were the first ones to mention folding in the Miocene sedimentary rocks of the Alaşehir Rift on the basis of the location of opposing dips (see their Fig. 3). They attributed the inferred folds to north–south shortening resulting from a putative 'short-lived north–south compression' (Koçyiğit et al. 1999, p. 613). Later, Seyitoğlu et al. (2000) attributed parts of the same folds to motion of the Miocene sedimentary section above listric normal faults, thus making the folds rollover structures (their Figs. 3, 4). Bozkurt (2002) acknowledged the possibility of Seyitoğlu et al.'s (2000) view, but stressed that a conclusive evidence for the presence of fault-related structures had not been presented. Sözbilir (2002) later interpreted some of the broad folds near Ahmetli as having formed over possible flat-ramp structures of low-angle normal faults. Bozkurt and Sözbilir (2004) later found folds with all the hallmarks of rollover anticlines south of Kara Kirse (see their Figs. 2, 7).

However, the folds and thrust faults discovered by Çiftçi and Bozkurt (2008) have provided clear evidence for layer-parallel shortening creating structures indistinguishable from those that form in environments of general shortening (orogenic belts, compressional bends along keirogens and zones of transtension). But Çiftçi and Bozkurt (2008) could not find a satisfactory explanation for how these structures might have formed in their present setting at a time when everywhere else in western Turkey nearly pure north–south extension was underway (Şengör 1987). They distinguished the earlier reported fault-related folds from those that actually formed by layer-parallel shortening. They mentioned that a short-lived phase of north–south shortening had been invoked by Koçyiğit et al. (1999) to explain all the folds seen in the rift fill, but they rightly thought this would not do for two reasons: first, some of the folds were shown to be related to fault-related bending folds of

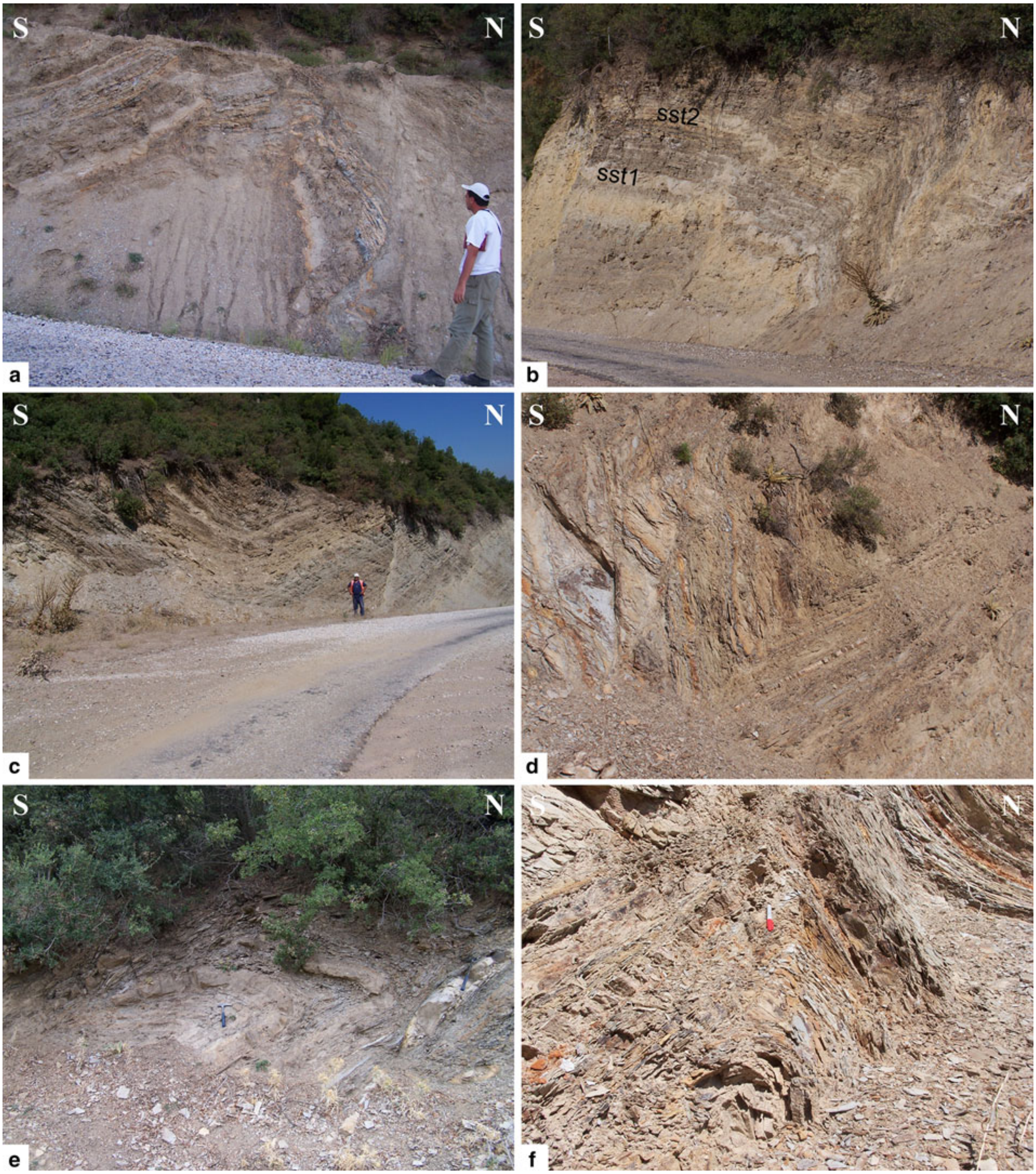


Fig. 11 Outcrop photographs illustrating mesoscale folds deforming the Alaşehir Formation. For their locations see Fig. 10. Fig. 11a–d are from Çiftçi and Bozkurt (2008). **a** A north-vergent asymmetric close fold displaying numerous small thrust faults on its northern limb and a large normal fault on its southern limb. Note that the crestal region is also cut by thrusts, so that no extrados extension can be seen. The normal fault is probably not a part of the folding process, but it originated later cutting the entire structure. Note also that the extension caused by the normal fault is very much smaller than the shortening that formed the fold itself. Bozkurt Çiftçi stands for *scale*. **b** A faulted, north-facing monocline. The cream-coloured sandstone bed (marked as sst 1) is repeated by small thrust faults near the monoclinal axis. By contrast, the sandstone bed marked as sst2 is cut by both normal and thrust faults. Here again, the extension caused by normal faulting is considerably less than the shortening accomplished by the thrust faults. In both cases, the normal faults are later than the thrust faults. **c** A nearly upright (with only a slight northerly inclined axial plane) open fold that formed immediately behind a steep, north-vergent thrust fault. Erdin Bozkurt stands for scale. **d** A steeply north-verging fold. Small thrusts with only a small normal fault cut the northern steeper limb. Here, the relative ages of the two kinds of faults are indeterminate. **e** Tight hinge of north-vergent fold. Note that the extrados extension joints in the competent layers. **f** A beautifully exposed gently north-vergent fold hinge trending approximately east–west

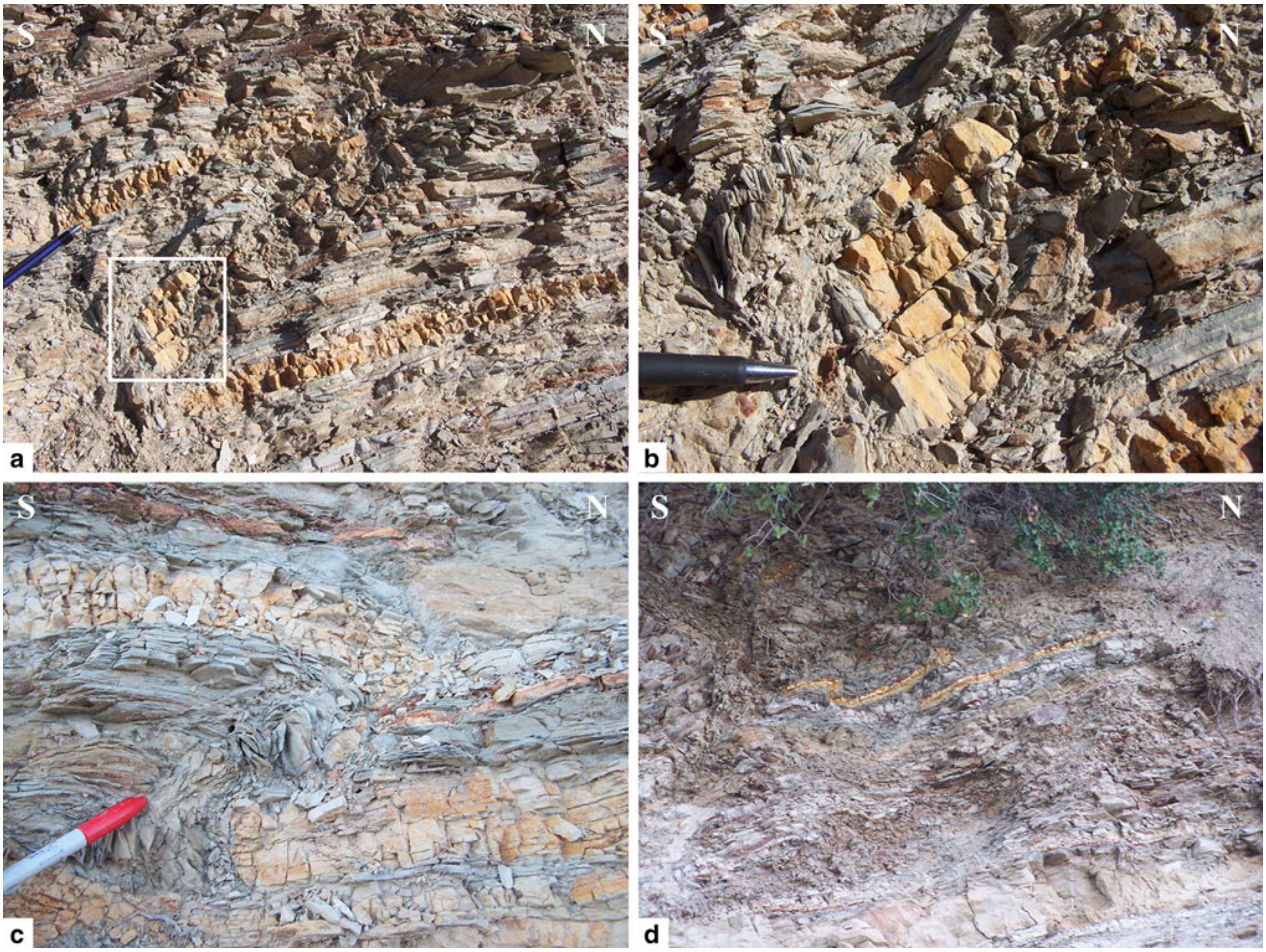


Fig. 12 Mesoscopic folds and associated structures. **a** A thrust fault that grew out of an asymmetric fold. **b** Note the small thrust faults offsetting the *pink* sandstone bed showing the nearly pervasive nature of the strain. **c** A similar asymmetric fold with a thinned and stretched intermediate limb. **d** Thrust fault completely disrupting an asymmetric fold within an asymmetrically developed fold train

extensional origin; secondly, the folds and thrust faults they themselves had discovered were seen in a relatively restricted area of broad shearing and only in the Alaşehir Formation that had formed when the rift had first originated. It seemed unlikely that a phase of regional shortening, however, short-lived, could be manifested in so limited a space.

We here show that the folds Çiftçi and Bozkurt (2008) discovered are only to be expected in a broad dip-slip shear zone with a normal throw if one considers the progressive shear evolution of the bulk volumes of the affected rocks. Bozkurt (2001) estimated the initial fault hade that had controlled the deposition of the Alaşehir Formation to have been about 40°. Gessner et al., in their 2001 paper, where they republished Şengör's (1987, Fig. 1b) bivergent horst model, argued for a 30° to 50° hade for the same structure, thus agreeing with Bozkurt's estimate. With such hades, the Alaşehir sedimentary rocks may have been rotated to dips up to 40° already during their deposition as the bounding faults remained active. It is likely that the bounding faults were delimiting blocks that were also undergoing internal strain. This is shown by the thickness changes of the overlying sediments as documented on the basis of seismic reflexion data showing that the rift floor to rift shoulder transition is a broad shear zone (Figs. 2, 16). This shear zone obviously strained the rocks affected, and the strain ellipse appropriate for the Alaşehir Formation rocks should be similar to that in Fig. 7c. As the strain ellipse evolved, the beds would first shorten. Those with very flat dips would then extend. As the rift evolved and the hades became flattened and dips became steepened, they rotated within the strain ellipse and entered into the *domain of continuous shortening during*

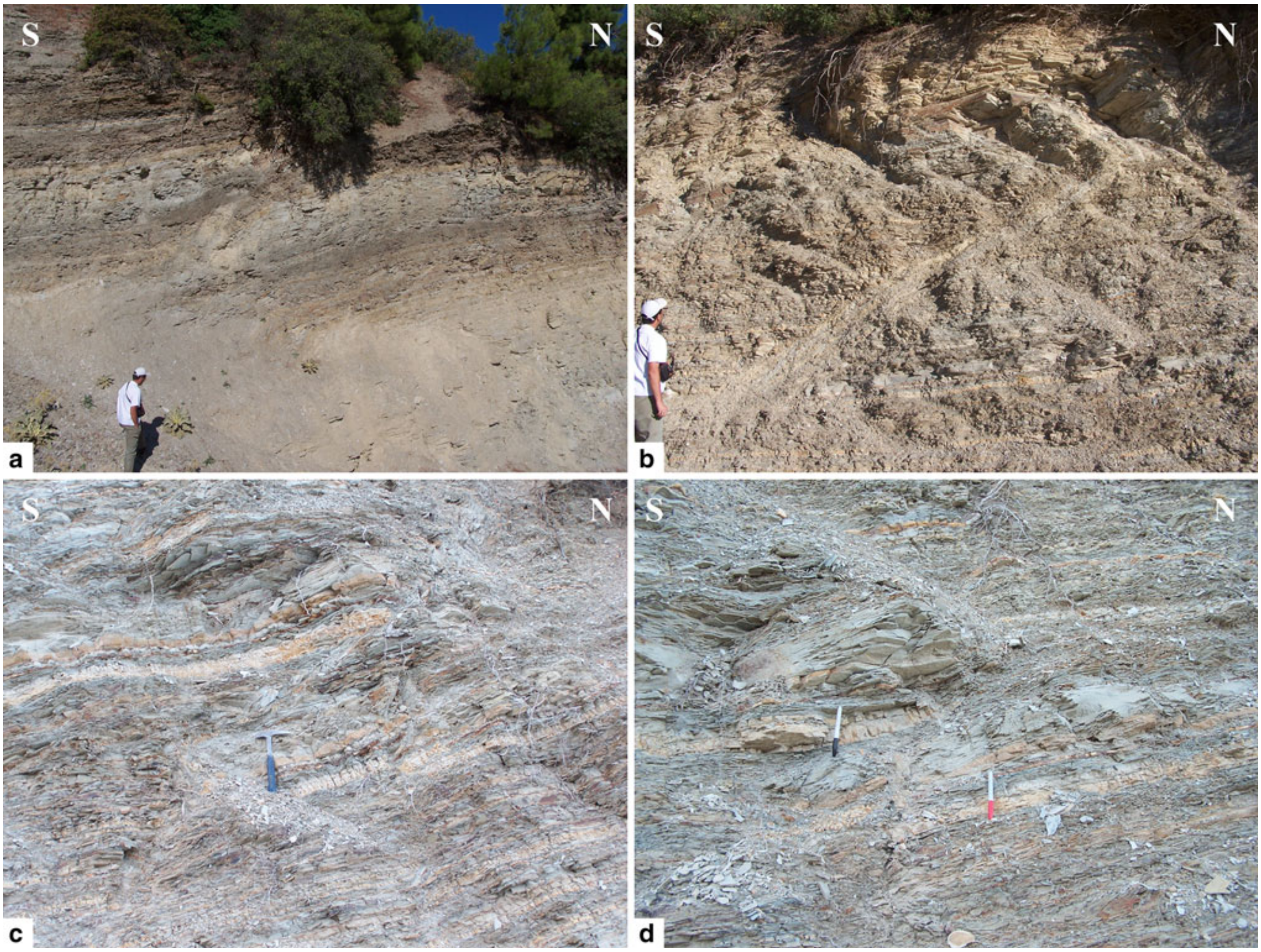


Fig. 13 a–d Macro- to mesoscopic thrust faults. Note that some developed out of earlier folds and some have cut the strata without a previous folding

the extension! In such cases, the strain ellipses themselves would also rotate. Because the faults and the beds would rotate at different rates, the resulting strain in rocks would not today reflect the geometry and attitude of a single strain ellipse. What one has here is the progressive deformation not only within a strain ellipse, but also the progressive reorientation of the strain ellipse itself.

The question that presents itself is why the broad shear zone was confined to Alaşehir Formation time. Let us remember that this Formation is the earliest fill of a young rift. At that time, rifting was possibly slower than it is today; only the smallest and thus the most insignificant of the faults in the shear zone would have been localised immediately, and therefore, it was these faults that have taken up the extension. At slow strain rates, the deformation was evidently (and expectedly) more ductile or ductile-like (because more distributed) and thus the generated shear zone broader. As rifting progressed, the rate of deformation increased to its present rate of at least 0.6 cm/a in western Turkey (Şengör 2011b). This rate is obtained by simply distributing the annual rate of extension of some 3 cm/a to five master rift zones in western Turkey, namely, from north to south, Bakırçay/Simav, Alaşehir, Küçük Menderes, Büyük Menderes and Kerme. Jean-Pierre Burg (pers. comm. 2011) suggested that the same rate of extension may have been more widely distributed and its localisation to the five mast rifts may have increased the strain rate along those rifts. This is a more specific way of saying what we stated above and is very likely to be true. The present seismicity indicates that the Alaşehir and the Büyük Menderes rifts are the most active ones and they probably extend with the highest rates. That is why we think the rate of extension along them is *at least* 0.6 cm/a. This rate is almost twice as fast as the Ethiopian Rift System in East Africa.

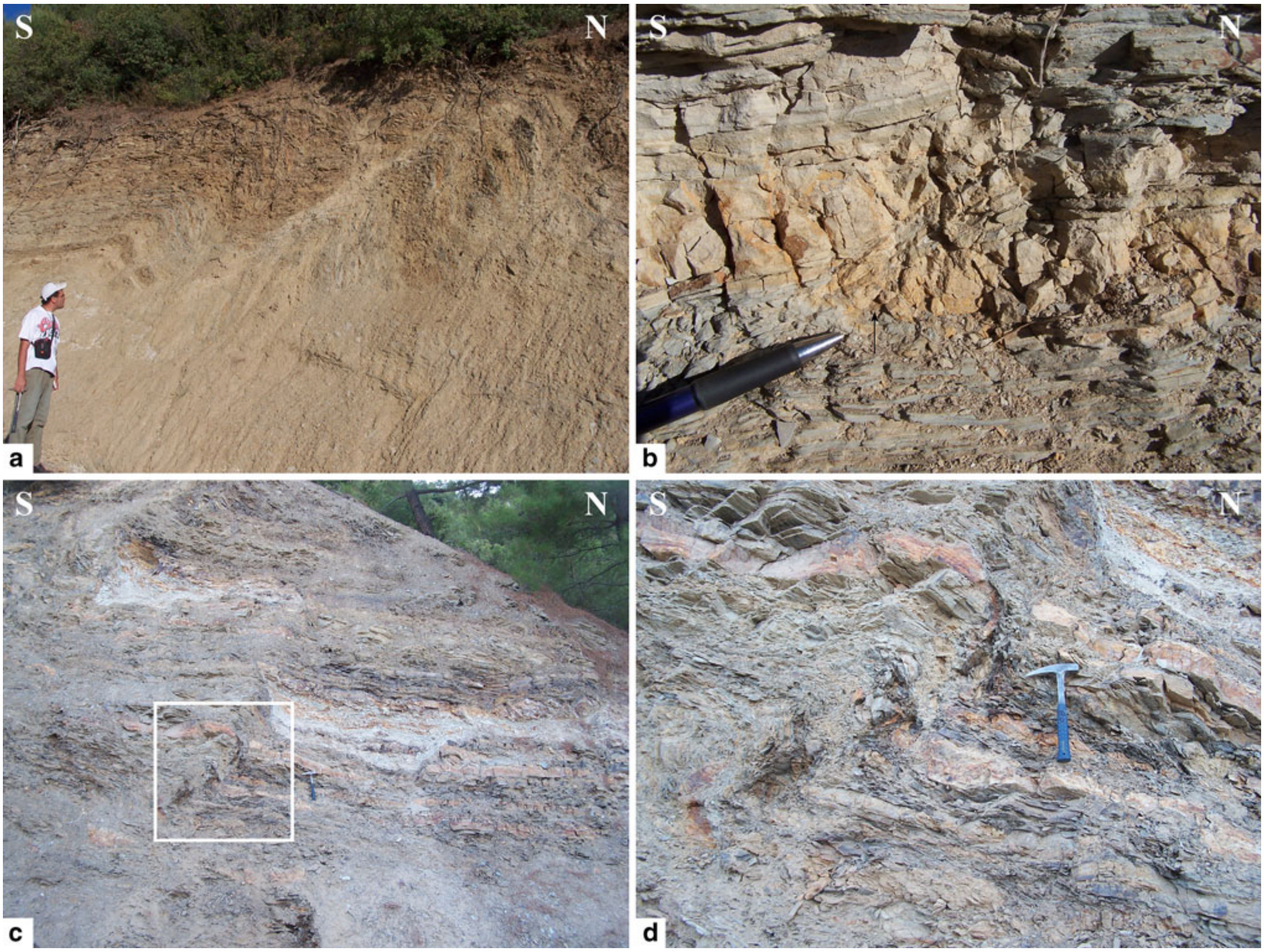


Fig. 14 Normal faults disrupting previously formed folds in a manner incompatible with the folding. **a** Shows a south-hading normal fault disrupting a previously formed fold with an angular hinge. **b** A small mesoscopic normal fault (indicated with the *pencil*) disrupting a previously formed fold-related thrust fault. Note that the displacement of the normal fault is incompatible with the strain implied by the thrust fault. **c** A small graben whose bounding faults cut earlier formed thrusts and folds (**d**). Note that the graben faults do not penetrate the entire section and their displacement is probably accommodated by dissipating the strain by intra-layer deformations (observe the rapid layer-thickness changes). **d** Detail of the thrust fault truncated by the later normal fault. Note that the displacement of the normal fault decreases downwards

At present, rifting is expressed by motion along narrow fault zones mainly along the southern margin of the Alaşehir Rift, although subordinate faulting started deforming its northern margin after the Miocene. As the faults thus formed are as narrow as one to a few metres at the surface, the shear zones they represent are not broad enough to form large folds resulting from layer-parallel shortening.

It is a different matter as to what may be happening at depth. As confining pressure and temperature increase, the fault zones seen at the surface normally pass into broader shear zones and folding similar to that we see in the Alaşehir Formation may be taking place at depth. There we expect the geometry of the resulting folds to be much more complicated for the reasons illustrated in Fig. 2a. At depth, the faults continue into the basement of the Menderes Massif that has had a complex history of deformation since the latest Precambrian. There, many layers of diverse orientations must be cut by shear zones broadening and flattening with depth during the latest episode of rifting. Depending on the orientation of such layered elements, structure geometries and histories of progressive deformation ranging from continuously shortening folds through folds that first form and then are disrupted by boudins to finally layers that experience only continuous extension are likely to be encountered.

However, at such deep levels, it is hard to distinguish folds that formed during orogenies from those that formed during the subsequent taphrogeny.

The very nature of extensional deformation provides, however, a guide to distinguish extensional folds that formed by layer-parallel shortening as explained above from those that form through regional orogenic shortening. Inherently, extension reduces crustal thickness and always cools the crust if the sedimentation rate and local magmatism do not raise the geotherms. Every crustal element is continuously cooled during extension, provided that we ignore sedimentation in the resulting rift trough. Extensional folds must therefore form in an environment of retrograde metamorphism. By contrast, orogeny thickens the crust, and during orogeny crustal elements are heated, if we ignore erosion. Thus, orogenic folds generally form in an environment of prograde metamorphism. This simple picture is regrettably marred by sedimentary deposition and erosion. Thick sedimentary fills filling rift troughs increase the temperature at rift bottoms and may cause some prograde metamorphism, although as Smith (1976) pointed out, for example, even at the bottom of the about 16-km-thick Gulf of Mexico Coast sediment pile that the temperature may be just about 320 °C under 4 kb pressure (but this ignores the radioactive heating of the pile). Therefore, it is unlikely that heating through sedimentary deposition will much affect the metamorphic grade of the deep metamorphic rocks under rifts (see also Şengör 2011a).

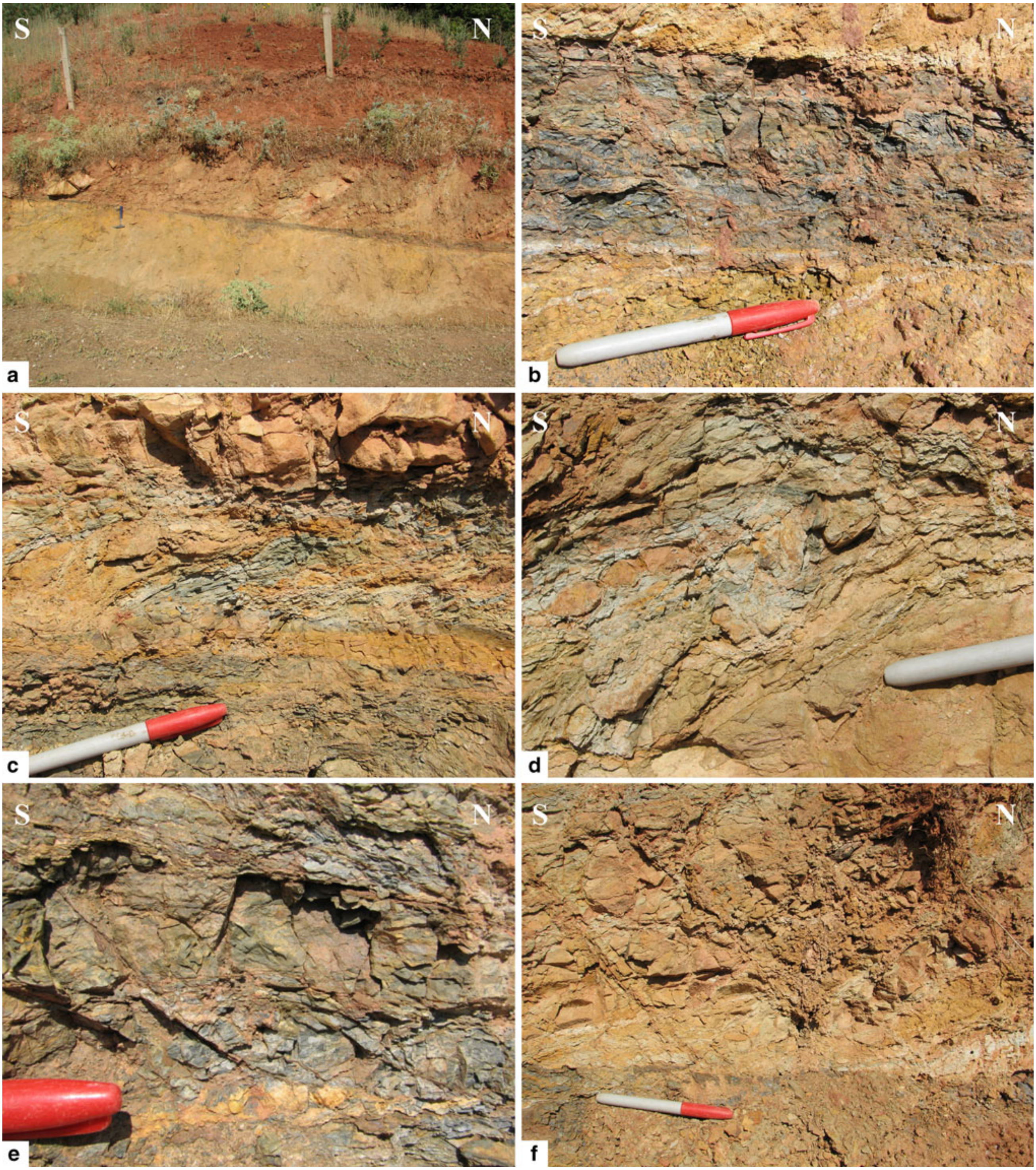
It is unfortunate that in non-metamorphic supracrustals, in which layer-parallel shortening creates folds during pure extension, no such helpful criteria may be found within the micro- to mesofabric of the folded rocks to distinguish their tectonic ecology. But in the case of supracrustals, the original tectonic environment itself is generally preserved because of little erosion.

Conclusions

Our observations in the Alaşehir Rift have revealed that rifting creates structures and structural associations betraying a multifarious strain picture and history, which, if not properly understood, are likely to lead to confusion. Hardly, any rift-bounding structure is a single fault (Fig. 2b). Normally, a set of faults bound rifts, and their spacing may be so close as to approximate a strain continuum, in which the rift appears to be bounded by a broad fault zone. At shallow structural depths, individual faults govern the deformation, but deeper even these individual faults normally pass into broader ductile shear zones. The depths at which this transition occurs are variable, depending on the thermal regime of the region. Where the thermal gradient is steep, the transition occurs within the first 5–7 km or even less. During rapid displacement events such as earthquakes, brittle failure extends farther down. In western Turkey, the normally ductilely slipping shear zones break brittley during earthquakes (e.g. Eyidoğan and Jackson 1985). This explains why some think that the normal faults extend down to great depths as planar structures. They do, but at greater depths must bend into listric geometries as broader shear zones. Only during earthquakes a part of the ductile shear zone fails along the extension of the planar fault, giving the misleading impression of planar faults that always slip along an uncurved surface. Otherwise, it would be impossible to explain the transfer of displacement from faults hading some 60°–30° at rift margins to almost horizontal slip planes below the rift.

Along the shear zones, folds and thrust faults of diverse type and orientation may form depending on the shape and attitude of the shear ellipsoid, the presence and attitudes of local inhomogeneities such as bedding and foliations in the rocks and the ambient thermal and pressure regime during the deformation. If the geologist does not attend to the detailed strain represented by the totality of the structures observed, he may come to entirely wrong conclusions. Particularly, misleading would be hasty observations on the so-called shear or kinematic indicators. Such indicators say even less than the individual folds and/or thrusts, because in many cases they only represent spatially limited sectors of such structures (see also the pertinent, but generally ignored criticism in Flinn 1994). Unless the total strain responsible for the structures in which such indicators may form is understood, it is unlikely that their message can be correctly interpreted.

Some have interpreted the folds in rift-bounding shear zones as ‘drag structures’ (e.g. Shackelford 1989). Apart from the fact that the term ‘drag’ is inappropriate, such structures are not even ‘first fold, then break’ type structures along faults as illustrated by the inconsistent vergence that folds display within the same shear zone (e.g. Fig. 10) giving a ‘wrong sense of drag’. Without a thorough structural mapping in an area of well-established stratigraphy, it is extremely hazardous to make conjectures about the origin of the structures observed in reconnaissance campaigns. Regrettably, structural geology has become, as have many branches of our science, fashion-driven. This paper is a small contribution to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◀ **Fig. 15** Outcrop photographs of the Gediz Detachment showing layer-parallel folding associated with extensional deformation. **a** The Gediz Detachment at outcrop near Kara Kirse. The layers nosing down onto the Detachment with dips of about 30° southwards belong to the Alaşehir Formation. The Detachment itself here hades about 80° northwards. Below the Detachment are the metamorphic rocks of the Menderes Massif. **b** The character of the intense shear deformation along the Detachment fault rocks [mostly intensely folded and sheared clay-rich rocks with only local mylonite (We here use mylonite in the sense originally defined by Lapworth (1885), in which all the characteristics of both brittle and ductile deformation are clearly spelled out. The recent tendency of calling any shear-generated brecciated and foliated rock a mylonite has only caused unnecessary confusion. We here reproduce Lapworth's definition that has long seemed forgotten: 'The old planes of schistosity become obliterated, and new ones are developed; the original crystals are crushed and spread out, and new secondary minerals, mica and quartz, are developed. The most intense mechanical metamorphism occurs along the grand dislocation (thrust) planes, where the gneisses and pegmatites resting on those planes are crushed, dragged; and ground out into a finely-laminated schist (*Mylonite*, Gr. *mylon*, a mill) composed of shattered fragments of the original crystals of the rock set in a cement of secondary quartz, the lamination being defined by minute inosculating lines (fluxion lines) of kaolin or chloritic material and secondary crystals of mica. Whatever rock rests immediately upon the thrust-plane, whether Archæan, igneous, or Palæozoic, &c., is similarly treated, the resulting mylonite varying in colour and composition according to the material from which it is formed. The variegated schists which form the transitional zones between the Arnaboll gneiss and Sutherland mica schists are all essentially mylonites in origin and structure, and appear to have been formed along many dislocation planes, some of which still show between them patches of recognisable Archæan and Palæozoic rocks. These variegated schists (Phyllites or Mylonites) differ locally in composition according to the material from which they have been derived, and in petrological character according to the special physical accidents to which they have been subjected since their date of origin—forming frilled schists, veined schists, glazed schists, &c., &c. The more higly crystalline flaggy mica schists, &c., which lie generally to the east of the zones of the variegated schists, appear to have been made out of similar materials to those of the variegated schists, but to have been formed under somewhat different conditions. They show the fluxion structure of the mylonites; but the differential motion of the particles seems to have been less, while the chemical changes much greater. In some of these crystalline schists (the augen-schists) the larger crystals of the original rock from which the schist was formed, are still individually recognisable, while the new matrix containing them is a secondary matrix of quartz and mica arranged in the fluxion planes. While the *mylonites* may be described as microscopic pressure breccias with fluxion structure, in which the interstitial dusty, siliceous, and kaolinitic paste has only crystallised in part; the *augen-schists* are pressure breccias, with fluxion structure, in which the whole of the interstitial paste has crystallised out. The *mylonites* were formed along the thrust planes, where the two superposed rock systems moved over each other as solid masses; the *augen-schists* were probably formed in the more central parts of the moving system, where the all-surrounding weight and pressure forced the rock to yield somewhat like a plastic body. Between these augen-schists there appears to be every gradation, on the one hand to the mylonites, and on the other to the typical mica schists composed of quartz and mica'. (Lapworth 1885, pp. 558–559, italics Lapworth's)) occurrences]. Note the distinct Riedel shears and tension gashes. These are probably the manifestations of a late-stage, mostly brittle slip along the Detachment. **c** A small-scale nearly recumbent fold formed from an antiform–synform pair resembling a tiny nappe of the first genre (see Termier 1911, p. 5) with north vergence. **d** A similar fold but somewhat stretched and boudined. **e** A late possibly anti-Riedel (R′) shear soling into the main shear plane. **f** A north-dropping normal fault that cuts and offsets the shear plane parallel with the main shear zone. This brittle structure is most likely younger than the one shown in Fig. 15e

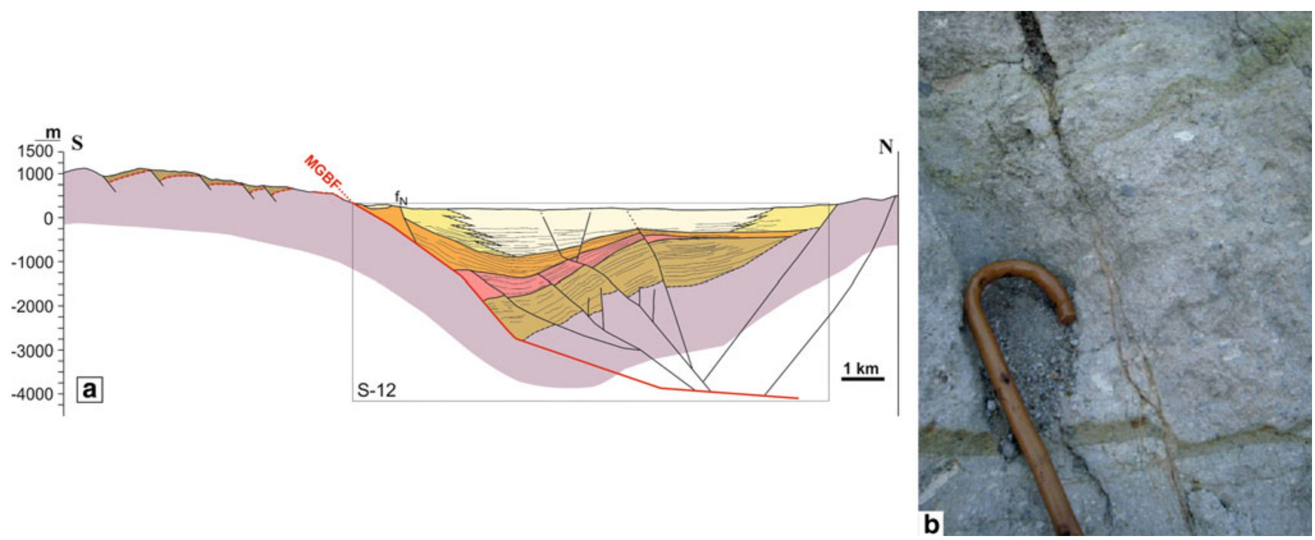

Fig. 16 Rift margins are usually broad shear zones and not single normal faults or even widely spaced normal faults. They are commonly present as such, because smaller, intervening faults and evidence for ductile-like deformation are ignored. **a** The floor geometry of the Alaşehir Rift seen in a N–S cross section (Çiftçi and Bozkurt 2010, Fig. 14a). Note that the sedimentary section is bent into a concave upwards geometry because of the numerous faults that cut it. Only a few of the faults could be seen on the seismic reflexion profiling. Those that cannot be seen most likely have much smaller offsets than the ones shown here. The overall picture one sees here is the presence of a broad normal dip-slip shear zone from rift shoulder to rift floor. **b** A small normal fault family in the Miocene volcanic rocks on the northern shoulder of the Edremit Rift in western Turkey. Note the small *green* layer that is offset by the fault family. The geometry of that layer indicates that where the fault family is, there is a broad shear zone that cannot be seen where such a layer is not present. Even at such shallow depths, normal faults are rarely single surfaces, but consist of a family of faults that function as a 'ductile-looking' shear zone broader than any single fault. This is our justification for treating the rift margins as broad dip-slip shear zones

alerting geologists not to follow fashions unless their meticulous mapping can justify it.

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