Roles of strike-slip faults during continental deformation: examples from the active Arabia-Eurasia collision

Mark B. Allen

Department of Earth Sciences, University of Durham, Durham, DH1 3LE, UK

Abstract: This paper concerns the kinematics of active strike-slip faults in the Arabia-Eurasia collision zone, and how they accommodate plate convergence. Several roles are discernible. *1) Collision zone boundaries* – the left-lateral Dead Sea Fault System and right-lateral faults in eastern Iran form the western and eastern boundaries of the collision zone. *2) Tectonic escape structures* – the North and East Anatolian faults transport intervening crust westwards, out of the path of the Arabia. *3) Strain partitioning* – right-lateral slip on the Zagros Main Recent Fault and NW-SE striking thrusts to its SW produce north-south convergence, parallel to the plate vector. Left-lateral slip along the Alborz range and thrusts across it produce oblique left-lateral shortening. *4) Shortening arrays* – arrays of strike-slip faults (e.g. Kopeh Dagh and eastern Iran) rotate about vertical axes, producing north-south shortening without crustal thickening. *5) Transfer zones* – fold trends and earthquake slip vectors change orientation across strike-slip faults in the Zagros, suggesting that these faults allow for changes in thrust transport along strike in the orogen. These different roles emphasise the complex behaviour of continental crust, and the advantages of studying active tectonics rather than ancient examples.

Introduction

This paper reviews active strike-slip faults from the Arabia-Eurasia collision zone (Figures 1, 2 & 3), to summarise the different ways such faults help achieve plate convergence during continent-continent collision. This is an important issue for two reasons. The first is that it is part of the more general problem of how faults in the upper crust collectively produce the velocity fields required by plate motions. The second is that strike-slip faults are common features in the geological record of the continents, but it is not always easy to determine why such faulting took place. Active tectonics provides data and constraints not available in ancient settings, principally through studies of decadal to millennial slip vectors (via GPS and seismicity studies) and through use of the landscape to deduce deformation patterns. The approach is to use case studies from different regions to make general conclusions about the way the strike-slip faults in the upper crust behave during continental deformation. It is not intended to be a systematic account of every active strike-slip fault in SW Asia, nor does it dwell on the many other aspects of the collision. Other papers (e.g. Mann, 2007) synthesise the structures associated with continental strike-slip faults, regardless of their origins: such material is not repeated here.

Active slip rates and finite offsets are known for many of the strike-slip faults, and in some cases there are data for the timing of onset. Therefore it is possible to compare the patterns of short-term and long-term deformation in the collision zone, and by implication in continental crust in general. Strike-slip faults are easier to work with in this respect than thrusts or normal faults, where the overall shortening or extension may be poorly constrained through lack of sub-surface data.

Continental collision zones are excellent places in which to study continental deformation processes in general because of the widespread and highly variable nature of the deformation that takes place. Although collision by definition implies plate convergence, this can be accommodated in a tremendous variety of ways by combinations of compressional, strike-slip and even extensional structures (Dewey et al., 1986). Faulting is the main way in which strain is accomplished within the brittle upper crust, therefore the kinematics of fault zones are revealing about overall strain. However, there are few active continental collision zones in the world, compared with active subduction zone boundaries for example. One is the Arabia-Eurasia collision, part way along the network of Cenozoic orogenic belts between the Pyrenees and SE Asia known collectively as the Alpine-Himalayan system. Following a geological overview of the collision, later sections focus on individual faults and groups of faults, to show how their kinematics fit in to the overall plate convergence. Figure 3 is a summary map of the main active strike-slip faults in the Arabia-Eurasia collision, but also highlights the generic roles outlined in this paper, namely: collision zone boundaries (Dead Sea Fault system, eastern Iranian faults); tectonic escape structures (North and East Anatolian faults); strain partitioning elements (Main Recent Fault of the Zagros; Mosha Fault in the Alborz); shortening arrays (Kopeh Dagh); transfer and tear faults (Sangavar Fault).

Geological background

Collision between Arabia and Eurasia initially took place along the Bitlis-Zagros suture, which curves through SE Turkey before running NW-SE through southern Iran (Figure 1). The plate boundaries were a passive continental margin on the northern side of the Arabian plate and an active continental margin along southern Eurasia (Şengör et al., 1988; Beydoun et al., 1992).

The plate scale present-day convergence between Arabia and Eurasia is well-understood: GPS studies show that roughly 18±2 mm/yr north-south convergence takes place between the stable interiors of Arabia and Eurasia at longitude 48^{o}E (Figure 1; McClusky et al., 2000). Convergence velocities increase and azimuths swing anti-clockwise west to east along the collision zone, with a rotation pole in the northeast Africa/eastern Mediterranean region (McClusky et al., 2003) and velocities ~10 mm/yr higher in eastern Iran than the western side of the collision. GPS and seismicity studies together show that deformation is concentrated between the Persian Gulf and the north side of the Greater Caucasus and Kopeh Dagh ranges – there is a good correlation between the limits of seismicity and topographic fronts (Figure 2). But deformation is not distributed evenly within these northern and southern limits. Seismogenic thrusting, and hence plate convergence achieved by crustal shortening *and* thickening, is presently concentrated in areas below the 1 km topographic contour (Talebian and Jackson, 2004). This is mainly within the lower parts of the Zagros and Alborz/Caucasus regions at the southern and northern sides of the collision respectively (Figure 3). The intervening region has lower relief, elevations commonly over 1.5 km and is known as the Turkish-Iranian plateau. GPS data from within the collision zone reveal that little active internal shortening takes place within this plateau (~2 mm/yr or less; Vernant et al., 2004a), and large areas are aseismic. It is not totally quiescent: Late Cenozoic volcanics occur in discrete fields across it (Pearce et al., 1990; Kheirkhah et al., 2009), and strike-slip faults are locally associated with historical earthquakes, indicating at least some tectonic activity (Copley and Jackson, 2006). Another area of low internal deformation at present is the South Caspian Basin, north of the Alborz (Figure 3). Para-oceanic basement to this basin is in the early stages of subducting under the northern and possibly western basin margins (Mangino and Priestley, 1998; Jackson et al., 2002). This basement is detached from folds within the thick sedimentary cover: these folds are not typically associated with major seismicity, indicating that the basement behaves as a rigid block, presumably because of unusually strong basement.

The western margin of the collision zone is sharply defined along the Dead Sea Fault System, which allows the largely stable interior of Arabia to move northwards with respect to the eastern Mediterranean. This basement to the latter area is not well known as it is buried beneath a thick sedimentary cover, including salt. It is probably underlain by highly thinned continental or even oceanic crust (de Voogd et al., 1992). West of a triple junction at the northern end of the Dead Sea Fault System, subduction of eastern Mediterranean basement takes places along the Cypriot and Hellenic arcs. Collision has not yet taken place in these regions, and north of the Hellenic arc the Aegean crust is rapidly extending. This extensional province merges eastwards in onshore Turkey, in to the crust of Anatolia. Here there is little active internal deformation, but wholesale westwards transport between the North and East Anatolian faults (McKenzie, 1972). The eastern side of the collision roughly coincides with the political boundary of Iran and Afghanistan; the latter is part of stable Eurasia, in the context of the active deformation field. There is active subduction of Indian plate oceanic lithosphere under the Makran (Regard et al., 2005).

Less is known about the earlier evolution of the collision zone. Even the onset of collision is debated, with recent estimates ranging from Late Eocene (~35 Ma) to mid-late Miocene (12-10 Ma) (McQuarrie et al., 2003; Vincent et al., 2005; Guest et al., 2006a; Verdel et al., 2007). Allen & Armstrong (2008) proposed that there was evidence from many localities both sides of the original suture for Late Eocene (~35 Ma) deformation, uplift or changing sedimentation patterns, and that this was the true time of initial collision. This debate on the collision timing highlights how difficult it can be to interpret geological data from ancient settings. It arises in part because we can never have an overview for past times across the entire orogen, in the way that remote sensing, seismicity and GPS all provide for the active tectonics. Therefore data from one region for initial rock uplift, say, can get treated as though it is representative of the entire collision zone. This is misguided, given how the present day tectonics show the wide variety of deformation, and quiescence, that takes place at any one time.

As a general point, there is no systematic difference in the depths of the strike-slip and thrust earthquakes in the various regions of the collision zone, such as the Alborz and Zagros ranges (Figure 3). They are typically up to ~15-20 km, i.e. within the crystalline basement of the crust (Jackson et al., 2002; Talebian & Jackson, 2004). This indicates that the strike-slip deformation described in this paper is "thick-skinned" in structural geology terms.

Collision zone boundaries

Reduced to its simplest, the Arabia-Eurasia collision represents ~north-south convergence between a promontory (Arabia) and a much broader continental mass

(Eurasia). Figure 4 is a cartoon that highlights the main elements of the collision, and illustrates the role of strike-slip faults and the boundaries of deformation. The real locations of these structures are shown on Figure 3. The pre-collision position of the Arabian and Eurasian plate margins is not precisely known, but the north-south convergence vector requires hundreds of kilometres of northwards motion of the stable interior of Arabia with respect to stable Eurasia, over tens of millions of years (McQuarrie et al., 2003; Allen & Armstrong, 2008). Therefore it is unsurprising that the northern and southern limits to deformation are marked by thrusting (Figure 2) – allowing for plate convergence via crustal thickening, whereas the western and eastern limits are strike-slip fault zones – allowing the Arabian plate to move past adjacent crust. A crucial difference between the strike-slip faults on the western and eastern margins of the collision is that the former, the Dead Sea Fault System, decouples Arabia from the eastern Mediterranean, but both regions were part of the combined African-Arabian plate before collision. In the case of the east Iranian faults, the great majority of the region involved was part of Eurasia before the initial collision.

Deformation is sharply focused along the ~1000 km long, left-lateral Dead Sea Fault System (Garfunkel, 1981; Figure 3), except for local splays at releasing and restraining bends such as the Dead Sea pull-apart basin (Manspeizer, 1985) and the Mount Lebanon range. The southern end of the fault links in to the active extension within the Red Sea: debate continues as to the interaction of extension in this region and initial collision on the northern side of the Arabian plate (Jolivet and Faccenna, 2000; McQuarrie et al., 2003). The northern end links in to the folds and thrust belts in southeastern Anatolia and the Zagros. Total offset across the fault is ~105 km south of the Dead Sea (Quennell, 1958), and this is fully observed in an offset dyke swarm dated at 22-18 Ma (Eyal et al., 1981). Active and late Quaternary slip rate estimates are variable, at 2-8 mm/yr (e.g. Klinger et al., 2000), although more recent studies are producing values of ~5 mm/yr (Ferry et al., 2007; Gomez et al., 2007). This velocity would need ~20 million years to achieve the full offset, consistent with the age of the offset dykes, but inconsistent with the fault having operated at this slip rate since the proposed Late Eocene start of collision.

North-south right-lateral faulting in eastern Iran form the eastern boundary to the collsion zone (Figures 1-3). Oceanic subduction takes place under the Makran region, such that right-lateral faults in the extreme southeast of Iran juxtapose the easternmost Zagros (originating on the Arabian passive margin) with the accretionary prism to the east (Regard et al., 2005; Bayer et al., 2006). Further north, right-lateral faults to the east (Neh and Zahedan) and west (Nayband and Gowk) of the inert Dasht-e-Lut have a total offset estimated by Walker and Jackson (2004) as ~80 km. A difference between the eastern and western margins to the collision zone is that in the west there is only one, whereas in eastern Iran there are at least two active, parallel fault systems, and possibly several more. There is little doubt that the Nayband and Gowk faults and the Neh and Zahedan faults take up most of the slip between Iran and Afghanistan (Walker and Jackson, 2004; Walker et al., 2009), but the Deh Shir, Anar and Kuh Bahnan faults are also active (Meyer, 2006; Meyer and LeDortz, 2007), plausibly slip at 1-2 mm/yr in the Holocene, and so may contribute part of the overall shear. A more fundamental problem is why deformation is so focused at the western collision margin, and distributed in the east. The reason may be the distinct contrast in crustal type at the western side, where the Arabian crust was juxtaposed with para-oceanic basement to the eastern Mediterranean long before initial collision, when both regions formed part of the passive margin at the northern side of the African-Arabian plate. In eastern Iran and Afghanistan there is a mosaic of similar Gondwana-derived basement blocks (Şengör et al., 1988). Those blocks east of the Arabian indentor are not being deformed by the Arabia-Eurasia collision, but there is no sharp contrast within this crust as there is in the west.

The GPS derived right-lateral shear between eastern Iran and Afghanistan is ~16 mm/yr (Vernant et al., 2004a). This only requires 5 million years to achieve the total observed offset along the Neh/Zahedan and Nayband/Gowk faults. Given that all estimates of the initial collision put it much earlier than 5 Ma, something else accomplished right-lateral shear at the eastern side of the collision. The obvious explanation is that the region must contain faults that are now inactive, or only weakly active. The Deh Shir, Anar and Kuh-e Bahnan faults may have contributed relatively more to the boundary shear in the past, regardless of their precise present contribution. There may be further structures within the deserts of eastern Iran as yet unquantified or unrecognised.

Tectonic escape structures

The Arabia-Eurasia collision zone contains the first recognised example of so-called escape tectonics, in the case of Anatolian crust between the North and east Anatolian faults (McKenzie, 1972). These are active right- and left-lateral faults respectively, and act to transport intervening crust westwards, largely without internal deformation (Figure 1). Figure 4 reduces the kinematics to their simplest. The left-lateral East Anatolian Fault is the boundary between Arabia and Anatolia (Figure 3), and runs for

~400 km southwest of its intersection with the North Anatolian Fault at Karliova, at approximately 39.5^o N 41^o E. There are several strands to the fault zone, with localized pull-apart basins and push-up zones (Lyberis et al., 1992; Westaway, 1994). The GPS-derived slip rate is 9±1 mm yr-1 (McClusky et al., 2000) only needs to operate for ~3 million years to achieve the geological offset of 27-33 km (Westaway and Arger, 1996; Westaway et al., 2006), constrained by offset geological markers. This is in good agreement with the age of initial offset as late Pliocene (~3 Ma) or younger (Şaroğlu et al., 1992; Westaway and Arger, 2001), based on the offset of volcanics of this age.

The right-lateral North Anatolian Fault (NAF) achieves the slip between Eurasian and Anatolian crust for >1200 km (Figures 1 and 2), at a GPS-derived slip rate of 24±1 mm yr-1 (McClusky et al., 2000). The western end of the fault splits where it enters the north Aegean and passes in to the extensional deformation in that region. Roughly 80-85 km is emerging as a consensus figure for the total offset of most of the length of the fault zone, based on combinations of geological and drainage offsets (Armijo et al., 1999; Westaway, 1994; Seymen, 1975). Distributed strike-slip and/or extension took place in the mid or late Miocene, before the establishment of the present fault trace in some regions (e.g. Barka and Hancock, 1984; Tüysüz et al., 1998; Coskun, 2000; Şengör et al., 2005). There is no consensus on a precise age for the start of motion on the NAF, despite several estimates of ~5 Ma (see Bozkurt, 2001). The GPS-derived slip rate (24±1 mm/yr) achieves the total offset of 80-85 km in only ~3.5 million years, less than most geological estimates for the fault age. It seems that: i) the slip-rate is higher now than in the past (but this is uncertain), and ii) the fault has not been active since the start of collision (this is more definite).

Like the Dead Sea Fault System, the narrowness of both the NAF and EAF and the sharp velocity contrasts across them resemble plate boundaries, as utilised as long ago as McKenzie (1972) in his vector calculations. But this is a nearly instantaneous picture, and it is striking that both faults are young with respect to the overall collision zone, and need only a few million years at their present slip rates to achieve their total offset. In the case of the EAF, other faults may have played similar kinematic roles in the past. Other (inactive?) left-lateral faults have been identified in eastern Turkey, such as the Malatya-Ovacik Fault (Westaway and Arger, 2001), with ~29 km offset between 3-5 Ma, and the Ecemiş Fault (Jaffey and Robertson, 2001), with ~60 km offset, mainly between the Late Eocene and Miocene. Activity on the Central Anatolian Fault (Kocyigit and Beyhan, 1998) is disputed (Westaway, 1999). However, as the triple junction at the eastern end of NAF and EAF should migrate west with time, it is difficult to see how any of these inactive left-lateral faults in eastern Anatolia were the precise equivalent of the modern EAF.

Elements in strain partitioning

Plate boundaries are rarely orthogonal to plate vectors (Woodcock, 1986). This fact underlies the origins of many continental strike-slip faults, not only in collision zones. Accommodation of north-south convergence by east-west trending faults would be likely in idealised, isotropic crust, but has not happened in the heterogeneous crust of both Arabia and Eurasia. The suture zone trends NW-SE for much of its length (mainly within Iran), at roughly 45^{o} to the plate convergence vector. Pre-collision structural fabrics commonly lie parallel to the suture within both plates (e.g. Sarkarinejad et al., 2008). The pattern of active faulting in the Zagros strongly suggests pre-collision normal faults in the Arabian passive margin are now active as thrusts. Conclusive evidence for individual fault reactivation is rarely available, largely because of a thick sediment carapace over blind thrusts, but most folds and thrusts in the northwest Zagros trend NW-SE, parallel to both the suture and the trend of pre-collision sediment isopachs (Beydoun, 1992). The resultant NE-SW shortening is therefore oblique to the north-south plate convergence, and cannot achieve it on its own. The answer is the combination of this thrusting with adjacent strike-slip faulting, in an example of so-called strain partitioning (Figure 4).

Along the northeast side of the Zagros, loosely along the line of the original suture, there is a right-lateral strike-slip fault, the Main Recent Fault (MRF) (Talebian and Jackson, 2002; Figure 3). Offset along the MRF is ~50 km (Talebian and Jackson, 2002). Shortening across the widest structural unit in the Zagros, the Simple Folded Zone, is similar in magnitude (Blanc et al., 2003; McQuarrie, 2004). Combining the two estimates suggests ~70 km of north-south convergence across the Zagros, by applying Pythagoras' rule (Figure 5A). This is only valid if the strains took place at the same time. It is clear that shortening across the Zagros is active, and focused on lower elevations (<1 km) in the Simple Folded Zone. Vernant et al. (2004a) estimated 6.5±2 mm/yr north-south convergence at longitude ~51^{o}E, in their GPS survey of Iran. Likewise, both seismicity and GPS data indicate right-lateral slip along the Main Recent Fault, and the difference in slip vector azimuths between the Main Recent Fault and the Simple Folded Zone emphasise the effectiveness of partitioning. But the active slip rates do not fit a Pythagorean triangle as neatly as the total displacements, because GPS-derived slip along the MRF is only 3±2 mm/yr (Vernant et al., 2004a). This is less than the expected ≥10 mm/yr, if the onset of slip was ≤5 Ma (Talebian and

Jackson, 2004). A further complication is that the Simple Folded Zone deformation may have begun earlier than 5 Ma, as suggested by syn-fold deposition at ~8 Ma near the Zagros foreland (Homke et al., 2004).

Another example of strain partitioning in the active collision zone is from the Alborz mountains of northern Iran (Jackson et al., 2002; Allen et al., 2003; Guest et al., 2006b). This range lies between the Turkish-Iranian plateau to the south and the South Caspian Basin to the north (Figure 3). It is actively thrusting to both the north and south, and cut by range-parallel left-lateral strike-slip faults with offsets in the order of several tens of kilometres (Mosha, Astaneh; Figure 6) (Allen et al., 2003; Ritz et al., 2006; Hollingsworth et al., 2008). These are apparently segmented along strike, and at least locally more than one parallel fault segment is active – such as the Damghan Fault south of the longer Astaneh Fault. The resultant oblique motion across the range allows for westward motion of the rigid South Caspian basement with respect to Iran. Like the Zagros, the variation in earthquake slip vector azimuths helps make the case for effective strain partitioning (Jackson et al., 2002). Thus in contrast to the Zagros example, the strike-slip component of oblique shortening takes place predominantly within the thrust belt (Figures 5B and 6). Vernant et al. (2004b) determined the north-south shortening rate across the Alborz as 5±2 mm/yr and the left-lateral shear as 4±2 mm/yr, from a GPS study. Ritz et al. (2006) identified an extensional component on some of the left-lateral faults, which they suggested represented a Quaternary re-organisation of the deformation.

There is evidence for older, but probably late Cenozoic, right-lateral faulting along parts of the range (Axen et al., 2001; Allen et al., 2003; Guest et al., 2006b; Zanchi et al., 2006). Thus at least part of the Alborz strike-slip system shows evidence of rapid reversal of its sense of motion, possibly within the last few million years. Given that the folding within the South Caspian cover succession is only a few million years old at most (Devlin et al., 1999), the overall westward motion of the South Caspian basement is very young (Jackson et al., 2002), the present fault configuration may be as recent as the Quaternary (Ritz et al., 2006). In contrast, Hollingsworth et al. (2008) showed that present slip rates in the eastern Alborz require ~10 million years to achieve the total offset, suggesting that the present kinematics go back further in time.

The combination of left-lateral faulting along the Alborz and right-lateral along the Zagros has attracted repeated interest over the years, promoting the idea of eastwards escape of Iranian crust out of the collision zone, in an apparent mirror image to the westwards transport of Anatolian crust (McKenzie, 1972; Axen et al., 2001; Bachmanov et al., 2004). Both seismicity data (Jackson et al., 1995) and the GPS-derived velocity field (Vernant et al., 2004a) show that this is not the case (Figure 1), and that the strike-slip faults parallel to each range help accommodate oblique convergence across them (Allen et al., 2006). It the case of the Zagros, the resultant convergence is parallel to the regional plate vector. The Alborz strike-slip relates to the South Caspian basement moving as a rigid block within the collision zone, at a high angle to the overall plate convergence vector. This case study is a warning for all interpretations of escape tectonics in ancient orogens, where seismicity data and GPS-velocity fields are not feasible and the regional plate kinematics are not known: it is possible that such settings represent the strike-slip component of strain partitioning as outlined here. It should be feasible to distinguish between real and illusory escape tectonics, given that an essential component of strain partitioning is an adjacent zone of contemporary thrusting. In Anatolia, the neotectonic strike-slip faulting postdates previous thrusting and thickening.

Jackson (1992) noted that pure dip slip thrusting in the Greater Caucasus took place on slip vectors oriented clockwise of the overall convergence vector at this longitude. The overall convergence vector is achieved by combining this shortening in the Greater Caucasus with right-lateral strike-slip faulting to the south, within the Lesser Caucasus and the interior of the Turkish-Iranian plateau. This is most active in a WNW-ENE trending swarm of right-lateral faults including Van (Figure 3). Copley and Jackson (2006) also found that an array of NW-SE right-lateral strike-slip faults accommodate a NW-SE velocity gradient of NE directed velocity; these faults are located between the Van and Sevan faults. An aspect of this right-lateral shear within the Turkish-Iranian plateau (south of the Greater Caucasus) is that it is distributed across many faults, rather than focused on one main structure, which is the case to the west and SE in the NAF and Main Recent Fault respectively. In part this may be because of the presence of linear pre-Cenozoic sutures in the latter areas, available for reactivation. But it also relates to the way strain is partitioned across a much wider area than either the Zagros or Alborz, with the shortening component in the Greater Caucasus located north of the strike-slip faults (Jackson, 1992). The strike-slip fault system is constantly transported northwards by the shortening in the Greater Caucasus, in a way that does not happen in either the Alborz or Zagros.

Shortening arrays

Escape tectonics is one scenario where continental shortening takes place without crustal thickening. Strike-slip faults can achieve crustal shortening in another way, via arrays of en echelon faults rotating about vertical axes as they slip (Figure 4). The situation has parallels with the behaviour of normal faults in rift zones; in the latter case the faults rotate about horizontal axes as they slip and thin and extend the crust. In the strike-slip setting the net result is shortening across the fault zone and lengthening along it. Such fault arrays have recently been recognised in several places within the Arabia-Eurasia collision zone, mainly by James Jackson and colleagues.

The Kopeh Dagh in northeastern Iran lies on the northern side of the collision, between the Turkish-Iranian plateau to the south and the undeformed crust of the Turan platform to the north (Figure 3). Its structure is dominated by arcuate but broadly NW-SE trending folds and thrusts, which deform and expose Mesozoic and Lower Tertiary strata at current exposure levels. The right-lateral and range-parallel Ashkabad Fault lies along the northeastern margin of the range, trending WNW-ESE, such that the combination of slip along this fault and shortening/thickening across the range is another example of strain partitioning in the collision zone (Lyberis and Manby, 1999). But the folds and thrusts are offset by an en echelon array of right-lateral faults that strike NNW-SSE or NW-SE (Hollingsworth et al., 2006), such as the Quchan Fault. Palaeomagnetic data are not available to quantify tectonic rotations, but the folds of Mesozoic strata can be traced across the fault zones and the rotations thereby quantified. Knowing the rotations and the present dimensions of the fault arrays allows the total north-south shortening achieved by these faults to be estimated as ~60 km (Hollingsworth et al., 2006). The geometry of such a fault array is shown schematically on Figure 7. GPS data (Vernant et al. 2004a) put the total north-south convergence across the Kopeh Dagh as ~7 mm/yr. As there are no detailed estimates for crustal shortening via thrusting and thickening, it is difficult to compare geodetic and long-term deformation rates across the range.

A similar fault array exists south of the Kopeh Dagh (Figure 3), at the northern end of the north-south right-lateral structures within eastern Iran, where these faults die out and are replaced by left-lateral faults that appear to be rotating clockwise about vertical axes (Dasht-e Bayaz and Doruneh; Jackson and McKenzie, 1984; Walker and Jackson, 2004). The slip along the Deh Shir, Anar and Kuh Bahnan faults further south again (Figure 3) may be another example of this behaviour (Walker and Jackson, 2004) and not simply related to the eastern margin of the collision zone (Meyer and Le Dortz, 2007). This explanation has the advantage that such faults are well within the interior of Iran, and so seem poorly located to contribute to shear resulting from the contrast with Afghanistan beyond the collision zone. At the far northwest of Iran and in easternmost Turkey a similar right-lateral fault array is active and allows for shortening within the tip of the Arabian promontory (Copley and Jackson, 2006). Other right-lateral faults trend NNE-SSW or NW-SE across central Iran (e.g. Kashan, Indes). There is limited seismicity on some of these (Figure 2), but little indication that they contribute much to the overall strain pattern at present.

Deformation in the Greater Caucasus represents the northern component of the collision zone at present. Initial uplift in the range may be as old as Late Eocene (Vincent et al., 2007), such that this range carries a longer record of compressional deformation than most parts of the collision zone. Attention has focused on range-parallel thrusts, held responsible for a present-day convergence rate of ~10 mm/yr across it (Reilinger et al. 2006). However, there are oblique features within or close to the Greater Caucasus that look like fault zones at high angles to the overall structural trend. In particular, several folds terminate along NW-SE lines, just inland of the Caspian shoreline (Figure 3). Other structural breaks have the same orientation in the same region. No offsets are identifiable in the exposed geology, so that it is uncertain what these trends mean.

Transfer zones and tear faults

A textbook explanation for strike-slip faults within zones of compressional deformation is that they link along-strike sections of the thrusts, either where the latter die out laterally and strain needs to be relayed to another structure, or because it would be mechanically unfeasible to move the thrust sheets if they were too long. Such strike-slip faults are known as tear faults, or transfer faults. They have not been highlighted within the active fold and thrust belts of the Arabia-Eurasia collision. In part this may relate to the blind nature of many thrusts within the Zagros, Alborz, Caucasus and Kopeh Dagh: thrust earthquakes do not typically rupture to the surface through the thick sedimentary cover of these ranges. (This is in contrast to many of the longer strike-slip faults, where earthquake magnitudes can be higher, and surface ruptures are common for the larger events).

Transfer zones are present on larger scales, although there is potential overlap with some of the other kinematic roles defined in this paper (Figure 4). The Zagros Simple Folded Zone is cut by NNW-SSE or NE-SE trending right-lateral faults such as Kazerun and Sabz Pushan (Figure 8). These have offsets of a few to a few tens of kilometres. Higher estimates, based on range-wide structural and geomorphic correlations (Berberian, 1995) are not confirmed by local studies (Authemayou et al., 2006). Talebian and Jackson (2004) related these faults to the strike-slip deformation present along the MRF, and the need for lengthening along the Simple Folded Zone as a result of this slip. This is the same style of behaviour as the rotating fault arrays described in the previous section. However, predicted anti-clockwise rotations have not been detected palaeomagnetically (Aubourg et al., 2008). Blanc et al. (2003) noted that the strain partitioning in the NW Zagros does not occur in the east, where folds and thrusts are aligned roughly east-west, orthogonal to the convergence vector, with no strike-slip equivalent to the motion of the MRF. The strike-slip faults within the Simple Folded Zone act to link the zones of strain partitioning and no strain partitioning; individual folds cut by the strike-slip faults also change orientation across them, becoming more east-west further east.

Another scale of transfer behaviour occurs at the western side of the Alborz, where the north-south right-lateral Sangavar Fault (Berberian and Yeats, 1999) links the Alborz to the folds and thrusts in the Talesh (Talysh) range to the north (Figure 9). The arcuate and highly three dimensional nature of the structure in this part of the collision zone relates to the rigid basement of the South Caspian Basin, which underthrusts the Talesh to its west on very gently-dipping thrusts (Jackson et al., 2002). This is superimposed on a component of the regional north-south convergence, such that the overall kinematics appear highly variable in this region (Masson et al., 2006), despite the remarkable consistency in the velocity field with respect to Eurasia (Figure 9).. Deformation at the southeast corner of the collision zone is similarly complex, where the eastern Zagros abuts the Makran accretionary prism (Regard et al., 2005; Bayer et al., 2006).

Discussion

The examples described above demonstrate the different roles that strike-slip faults can play in one timeframe of one collision zone. Some generalities are possible. Strike-slip faults form the boundaries of major deformation zones, where these involve translation rather than convergence or extension. Strain partitioning involves strike-slip faults acting in concert with adjacent, parallel thrusts to achieve the overall convergence vector required by far field conditions. “Far field” mainly means the overall plate convergence zone, but can be rigid blocks moving within it, such as the South Caspian basement. Such partitioning produces the potential for the mis-interpretation of strike-slip faults as tectonic escape structures. Tectonic escape is the valid interpretation for the NAF and EAF, where independent estimates of the regional velocity field confirm the westwards transport of Anatolia with respect to both Arabia and Eurasia. This is not the case for central Iran, where strike-slip faults along the Alborz and Zagros ranges work with parallel thrusts to produce oblique convergence across each range. Geoscientists typically think of thrusts as the predominant structures in orogens, with mountain building as the result. En echelon right-lateral strike-slip faults within Iran show the potential for rotating arrays to achieve plate convergence, without crustal thickening. Such arrays are found both within areas of active thickening (Zagros, Kopeh Dagh, and, possibly, the Greater Caucasus), but also within the Turkish-Iranian plateau, where crustal thickening has ceased. In the latter case, the strike-slip mechanism for convergence has the advantage that it does not require work against gravity, which is important in areas of thickened and/or elevated crust where buoyancy forces oppose crustal thickening. A textbook explanation for strike-slip faults within fold and thrust belts is that they link individual thrusts, and ensure the continuity of strain across large regions. Such features have not been emphasised to date within the Arabia-Eurasia collision zone, but this may be because many thrusts in actively thickening areas are blind. Larger transfer zones exist, linking entire fold and thrust belts such as the western Alborz and southern Talesh (Figure 9).

In Woodcock's (1986) review of strike-slip faults at plate boundaries, all of the faults described in this paper would fall in the type "Indent-linked strike-slip fault", with the exception of the collision zone boundary faults which partly equate to the "Boundary transform" type. The kinematics of the faults within the Arabia-Eurasia collision, and interpretations on the roles they play in plate convergence, permit a more specific analysis. The five categories listed here (collision zone boundaries, tectonic escape structures, strain partitioning elements, shortening arrays and transfer zones; Figure 3) are not meant to be rigid. No doubt future studies will allow further refinement. The different kinematic roles are not necessarily mutually exclusive. Strike-slip faults in the Zagros link the western and eastern parts of this fold and thrust belt, but also contribute a small amount of shortening across the range (Figure 8).

In recent years there has been a debate as to whether continental deformation is best described by continuum models (where the emphasis is on the smoothness of the velocity field; England and Molnar, 2005), or a rigid block model (where the role of individual fault zones is paramount, and a quasi plate tectonic approach to the kinematics is valid; Thatcher, 2007). The Arabia-Eurasia collision has been involved in this debate, because of the availability of GPS- and seismicity data on its deformation. Reilinger et al. (2006) modelled the behaviour of the collision zone as a series of blocks, which collectively satisfied the overall velocity field. This approach involved reducing regions as broad and complex as the Zagros (200-300 km width) to a single boundary. Liu and Bird (2008) performed a finite element analysis of active deformation between eastern Anatolia and Burma, modelling geodetic data, geological fault slip rates and seismic moment tensor orientations. They showed that throughout the entire collision zone deformation was distributed, with only a few embedded rigid blocks, such as the South Caspian and Black Sea basins. These have para-oceanic basement distinct from the surrounding continental crust. The derived anelastic strain rate (0.7% per Ma) across the collision zone, apart from these rare blocks, is inconsistent with a rigid microplate model.

The two approaches outlined above produce radically different results. Each is correct in the technical sense that the data are properly handled in the framework of the model parameters. As Thatcher (2007) noted, the transition between the two end member behaviours is blurred: as fault number increases, block size decreases. The important question is, which is the more realistic model of continental behaviour, given the way faulting is distributed across the continental crust in the active examples we have available for study? In this context it is not only the number of fault zones within the Arabia-Eurasia collision that is notable, but their ability to rotate, reverse, accelerate or die within geologically short length- and timescales. Such mobility indicates a distributed model is the more useful way of understanding the deformation, rather than reduction to a small number of rigid microplates. Most of this review has focused on active or at least late Quaternary deformation, because of the wealth of data available for fault slip rates on these timescales. But a satisfactory description of how deformation occurs within the continents may only appear when we have enough data on the pre-neotectonic kinematics. To apply the phrase Brian

Windley has made famous, their behaviour cannot be summarised by a snapshot, the key lies in how the continents evolve.

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Figures

Figure 1. GPS-derived velocity field of the Arabia-Eurasia collision, with respect to stable Eurasia. The dashed line is the Bitlis-Zagros suture. Compiled from McClusky et al. (2000) and Vernant et al. (2004a).

Figure 2. Seismicity of the Arabia-Eurasia collision zone (from Allen et al., 2006). Small dots are epicentres from the catalogue of Engdahl et al. (1998). Focal mechanisms are from the following sources. Black: Waveform modelled, from Jackson (2001) and references therein, with additional events from Talebian et al. (2004) and Walker et al. (2005). Dark gray: Best-double-couple CMT solutions from the Harvard catalogue (http://www.seismology/harvard.edu/CMTsearch.html) for earthquakes with depth ≤ 35 km, Mw ≥ 5.5 and double-couple component ≥ 70%, in the interval 1977-2002. Light Gray: First motion solutions from Jackson and McKenzie (1984). Earthquakes deeper than 35 km associated with the subduction zones in the Makran, South Caspian and Hellenic Trench have been omitted.

Figure 3. Major active strike-slip fault zones within the Arabia-Eurasia collision zone. Derived from Allen et al. (2006) (Iran), Copley and Jackson (2006) (NW Iran), Allen et al. (2003) (N Iran), Bozkurt (2001) (central and NW Turkey), Kocyigit et al. (2001) (eastern Turkey). Activity on strike-slip faults in much of Anatolia is debated (e.g. Kocyigit and Beyhan, 1998, and Westaway, 1999), so that the Eskisehir and Central Anatolian faults are marked by dashed lines, and others shown by Bozkurt (2001) and Kocyigit et al. (2001) are not shown at all. The Salanda Fault is in the vicinity of a strike-slip earthquake of 1938 (Jackson and McKenzie, 1984), and so is more confidently assigned as active. Barbed lines show active thrust fronts, schematically.

Thrust zones are typically harder to map as precisely, because many of the active thrusts are blind. White barbs are subduction zones at the margins of the South Caspian Basin and Makran and along the Cypriot and Helenic arcs. Red Sea oceanic spreading is shown schematically by the double line.

Figure 4. Schematised kinematics of a continent-continent collision between plates X and Y, modelled after the Arabia-Eurasia collision and showing westward tectonic escape of block Z (i.e. Anatolia) and lateral strike-slip faults at the western and eastern boundary zones. Solid triangles indicate thrusts at the margins of the collision zone; open triangles indicate adjacent subduction zones. Thick black arrows indicate velocities with respect to the stable interior of block Y, with length proportional to velocity. The five roles of strike-slip faults described in this paper are highlighted as follows: (1) Collision zone boundaries – either diffuse or focussed (2) Tectonic escape structures (3) Strain partitioning elements (4) Shortening arrays with vertical axis rotations (5) Transfer zones.

Figure 5. The concept of strain partitioning: A) combined slip on the strike-slip fault and shortening across the adjacent thrust belt produces net convergence oblique to the fault trends – northwards motion of block X with respect to Y. This scenario is similar to the northwest Zagros Simple Folded Zone. B) Strain partitioning where the strike-slip fault system lies within the interior of the thrust zone. This geometry is similar to the Alborz mountains.

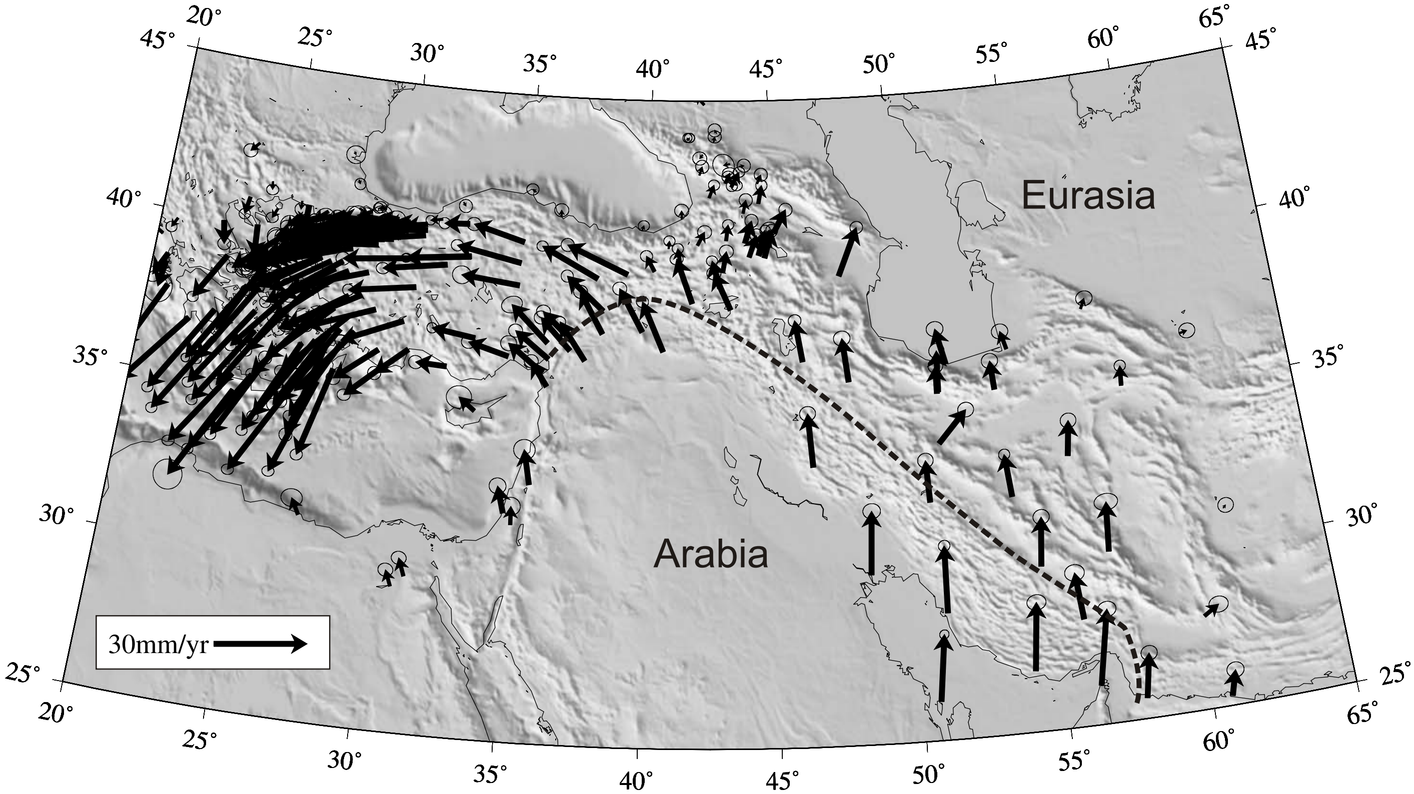
Figure 6. Active faults in the Alborz between 51° and 55° E. Left-lateral faulting occurs within the range interior, principally on the Taleghan, Mosha, Firuzkuh, and

Astaneh faults, which collectively form a segmented fault system. Thrusting takes place on inward-dipping faults at both the northern and southern margins of the range. The continuity of the Khazar Fault may be an artefact of Caspian lake highstands bevelling southwards against the bedrock of the range: the thrust is blind. Map derived from Allen et al. (2003), Ritz et al. (2006), Hollingsworth et al. (2008) and analysis of SRTM digital topography; focal mechanisms from Jackson et al. (2002) and Tatar et al. (2007).

Figure 7. Rotating strike-slip arrays acting to produce shortening and along-strike elongation (from Hollingsworth et al., 2006), as seen in the Kopeh Dagh. A) Fault blocks have initial width d and angle θ_0 with the deformation zone boundary, across a zone of width W_0. Grey bands represent fold trends, which act as strain markers as the faults and fault blocks are offset and rotated. B) Offset and fault block rotation produces new boundary length D, and angle θ_1, across a width W_1. C) If all fault block rotations are of the same amount, the geometry simplifies to a single triangle with lengths ΣD, Σd and Σs. Measurement of ΣD, Σs, θ_0 and θ_1 allows the original length of the deforming boundary (Σd) to be calculated using the cosine rule.

Figure 8. Active strike-slip faults in the Central Zagros. Several segmented right-lateral faults fan out from the southeastern end of the Main Recent Fault. Fault locations derived from Authemayou et al. (2006) and analysis of SRTM imagery. Focal mechanisms for thrust and strike-slip events in the region are from the Harvard and USGS catalogues (http://neic.usgs.gov/neis/sopar/) for earthquakes with Mw $\geq$ 5 and double-couple component $\geq$ 70%, in the interval 1986-2005.

Figure 9. Active faulting in the Talesh and western Alborz mountains, illustrating the role of the right-lateral Sangavar Fault as a transfer fault between the regions. Focal mechanisms from Jackson et al. (2002), with three additional events from the Harvard and USGS catalogues (http://neic.usgs.gov/neis/sopar/) for earthquakes with Mw ≥ 5 and double-couple component ≥ 70%, in the interval 2002-2007. Arrows show GPS-derived velocities with respect to Eurasia, from Masson et al. (2006). These do not change markedly across the region, despite the wide variation in fault strikes and focal mechanisms. The inset is a schematic transfer zone between two thrust belts, modelled on the junction of the Talesh and Alborz ranges. Deformation not only wraps around the rigid basement of block X, but has to accommodate its motion independent of the north-south convergence of larger regions Y and Z. This produces highly arcuate and complex fault geometries, which are unlikely to be stable over long periods.



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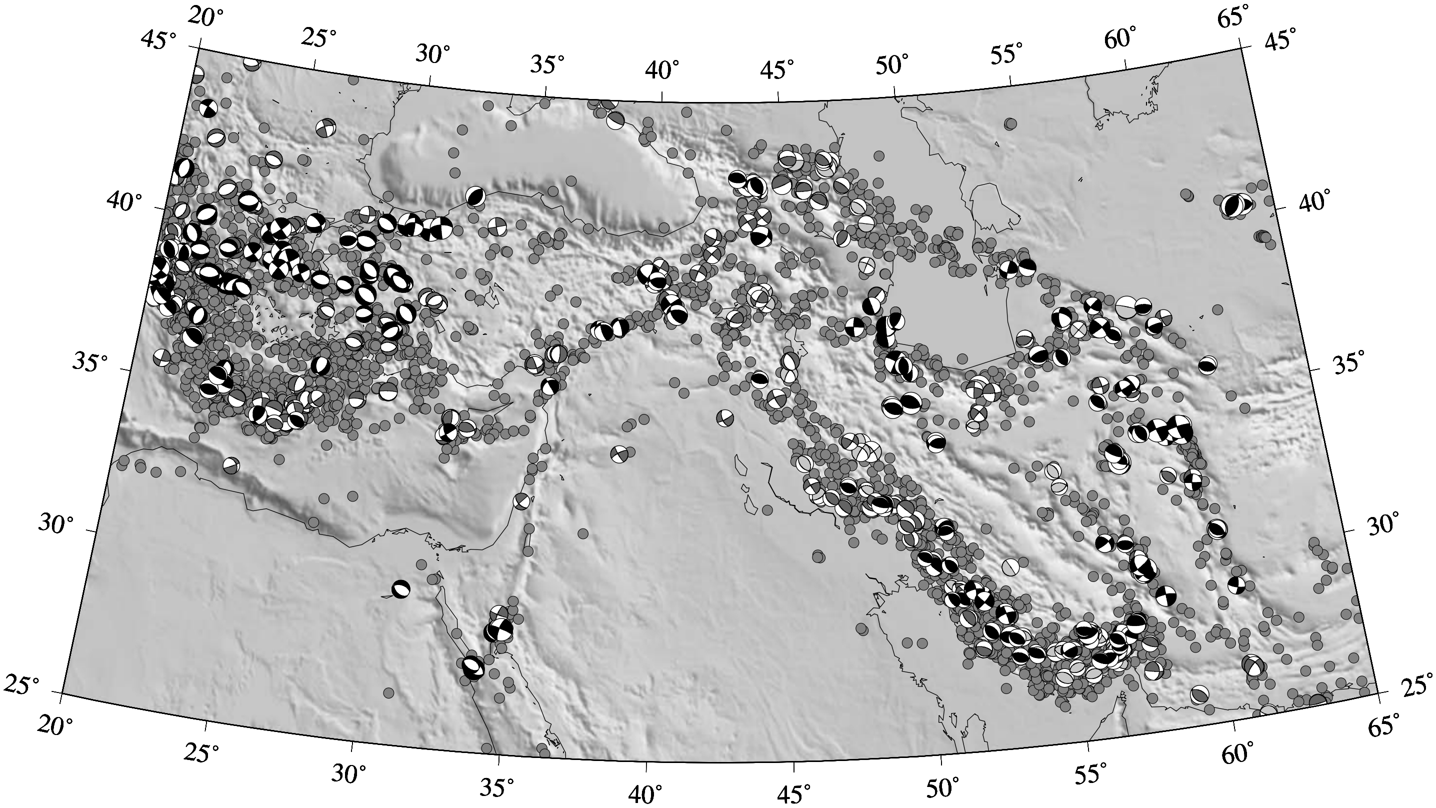

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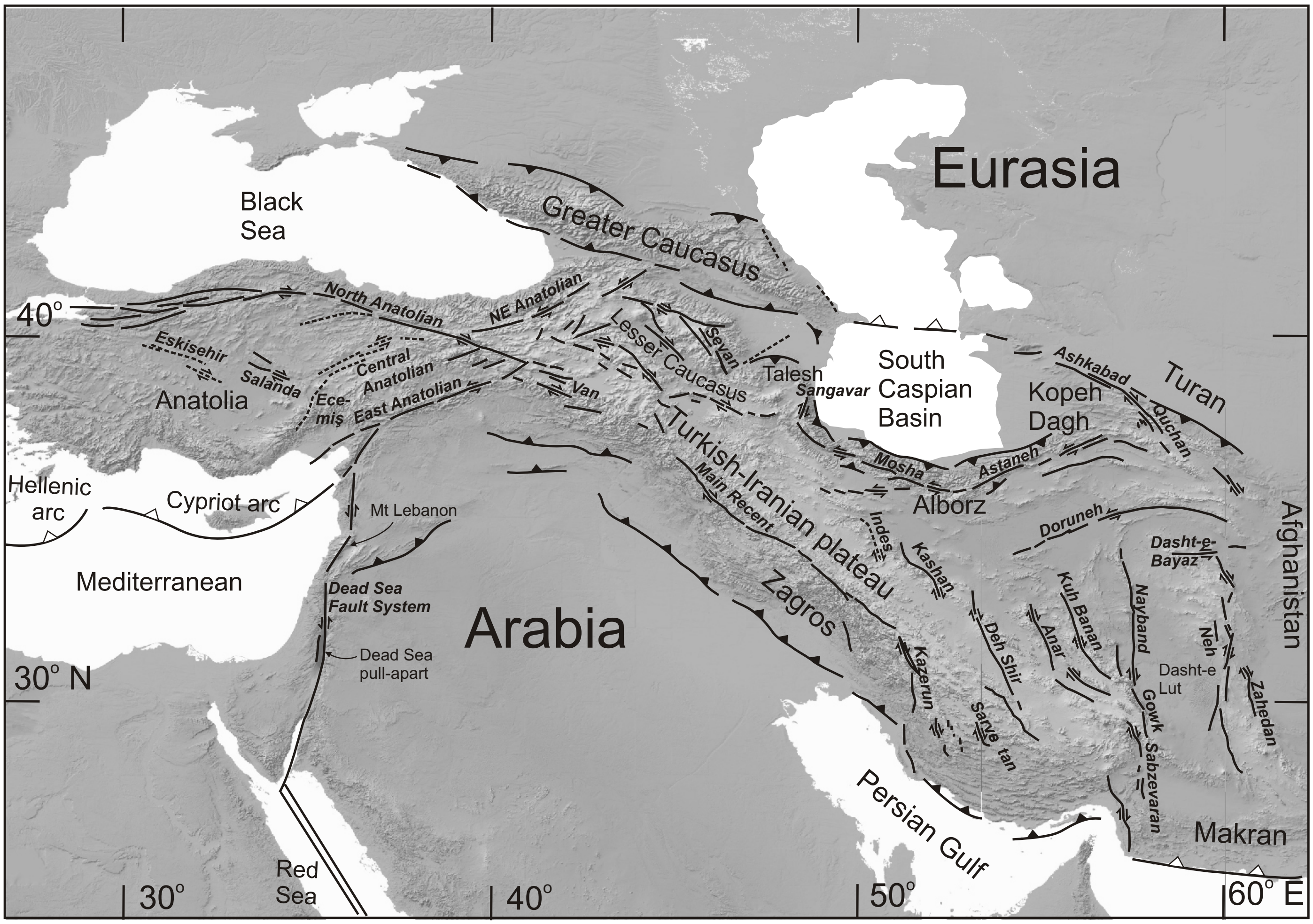
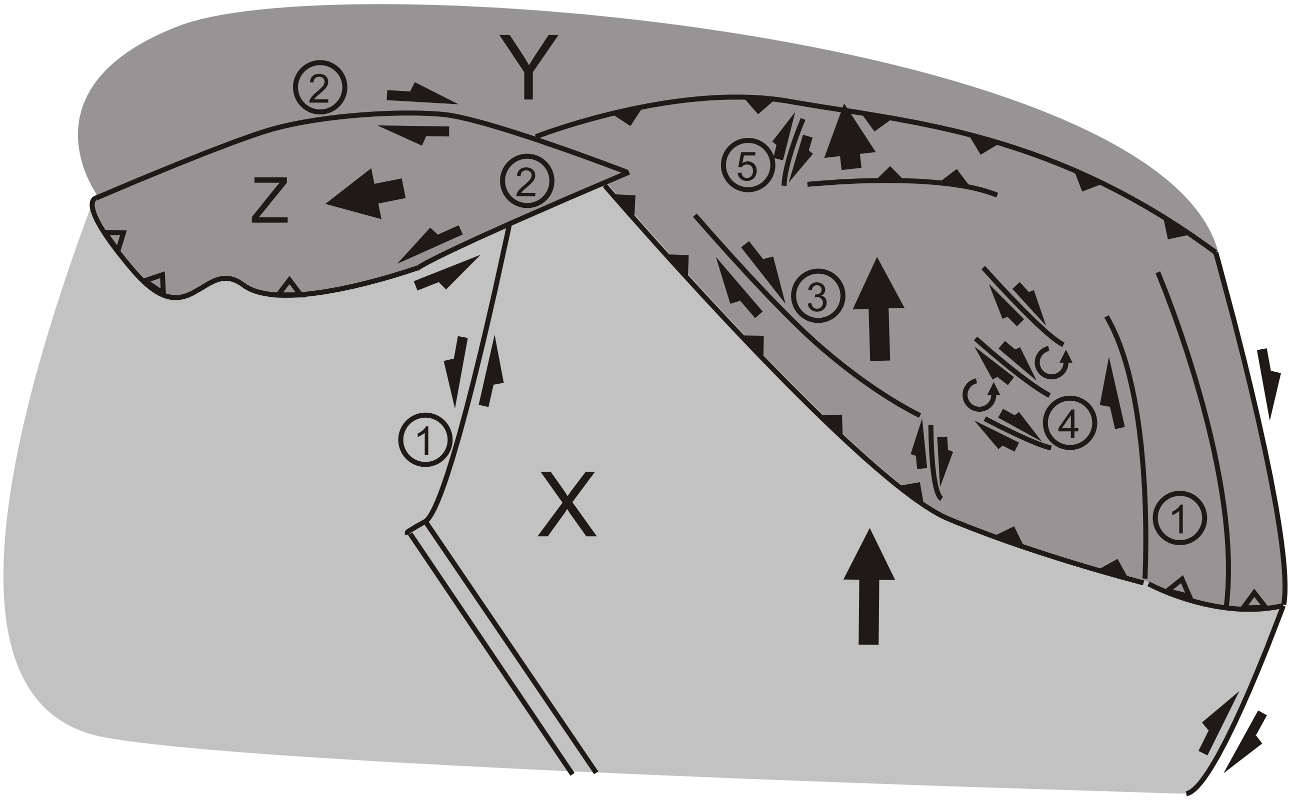

Eurasia
Black Sea
Greater Caucasus
North Anatolian
NE Anatolian
Lesser Caucasus
Sevan
Talesh
Sangavar
South Caspian Basin
Kopeh Dagh
Turan
Ashkabad
Quchan
Eskisehir
Salanda
Central Anatolian
Anatolia
Ece-miş
East Anatolian
Van
Turkish-Iranian plateau
Main Recent
Mosha
Astaneh
Alborz
Hellenic arc
Cypriot arc
Mt Lebanon
Mediterranean
Dead Sea Fault System
Dead Sea pull-apart
Arabia
Zagros
Indes
Kashan
Doruneh
Dasht-e-Bayaz
Afghanistan
Kuh Banan
Naybând
Neh
Deh Shir
Anar
Kazerun
Sarvestan
Dasht-e Lut
Gowk
Zahedan
Sabzevaran
Persian Gulf
Makran
Red Sea
40°
30° N
30°
40°
50°
60° E

3.



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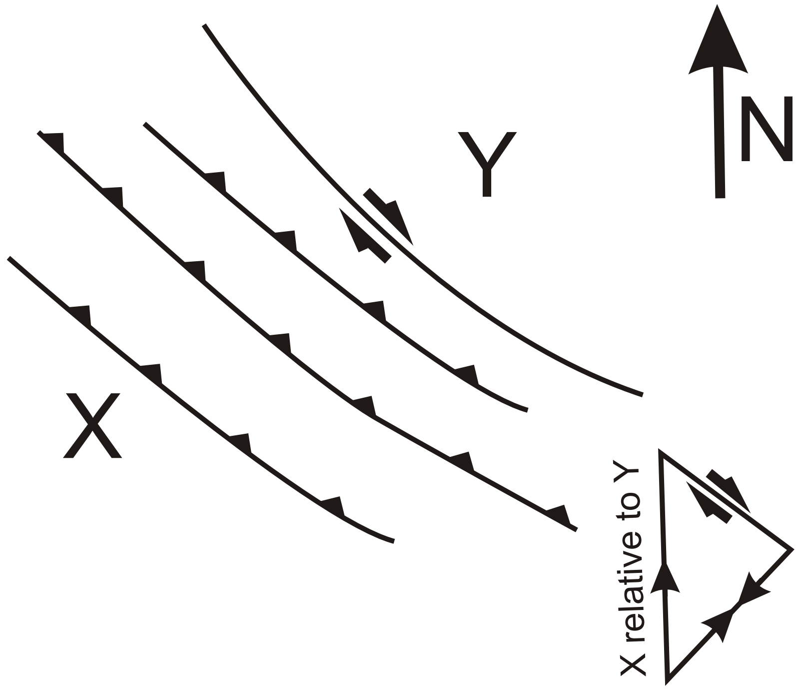
A)



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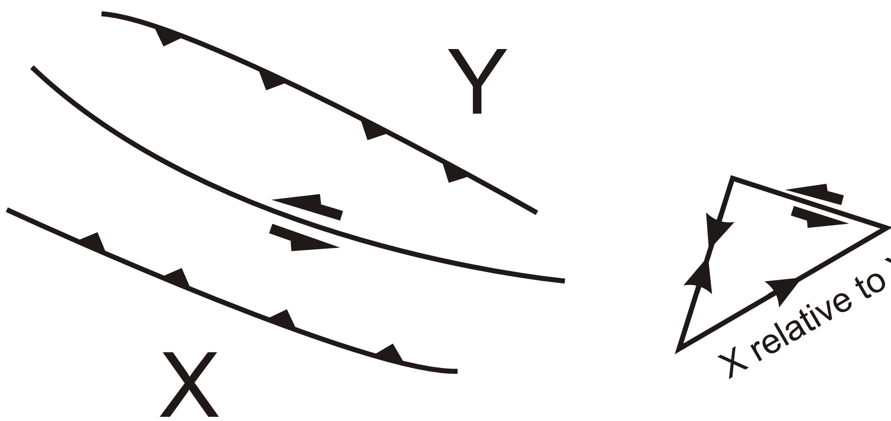

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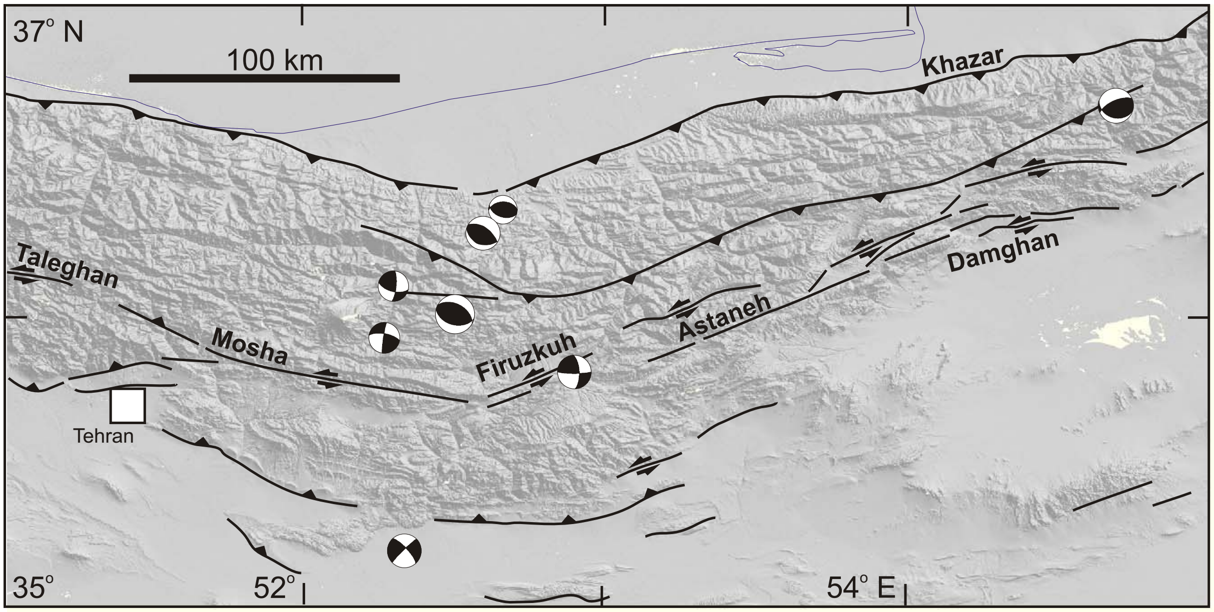
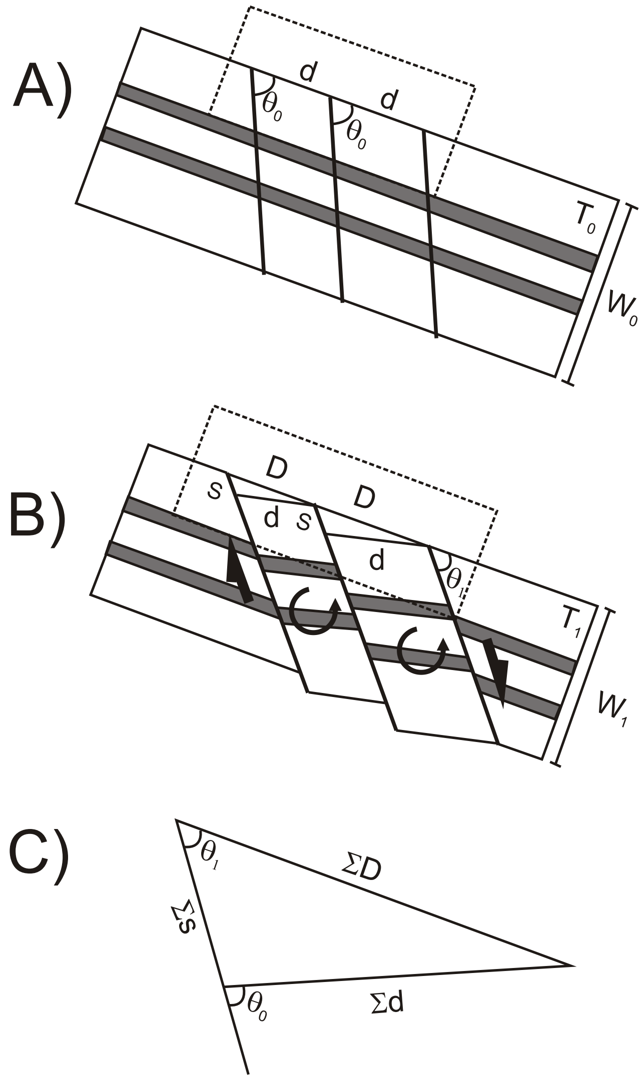
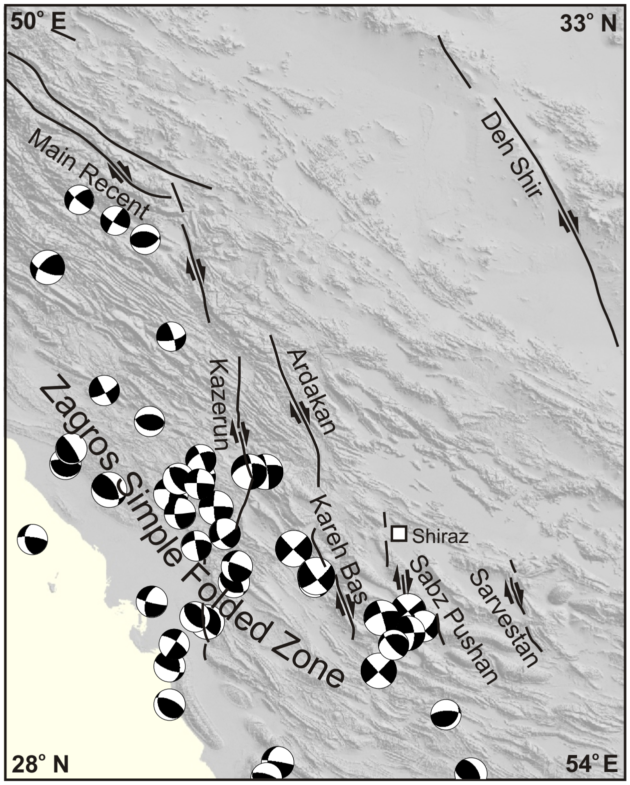

37° N
100 km
Khazar
Taleghan
Mosha
Firuzkuh
Astaneh
Damghan
Tehran
35°
52°
54° E

6.

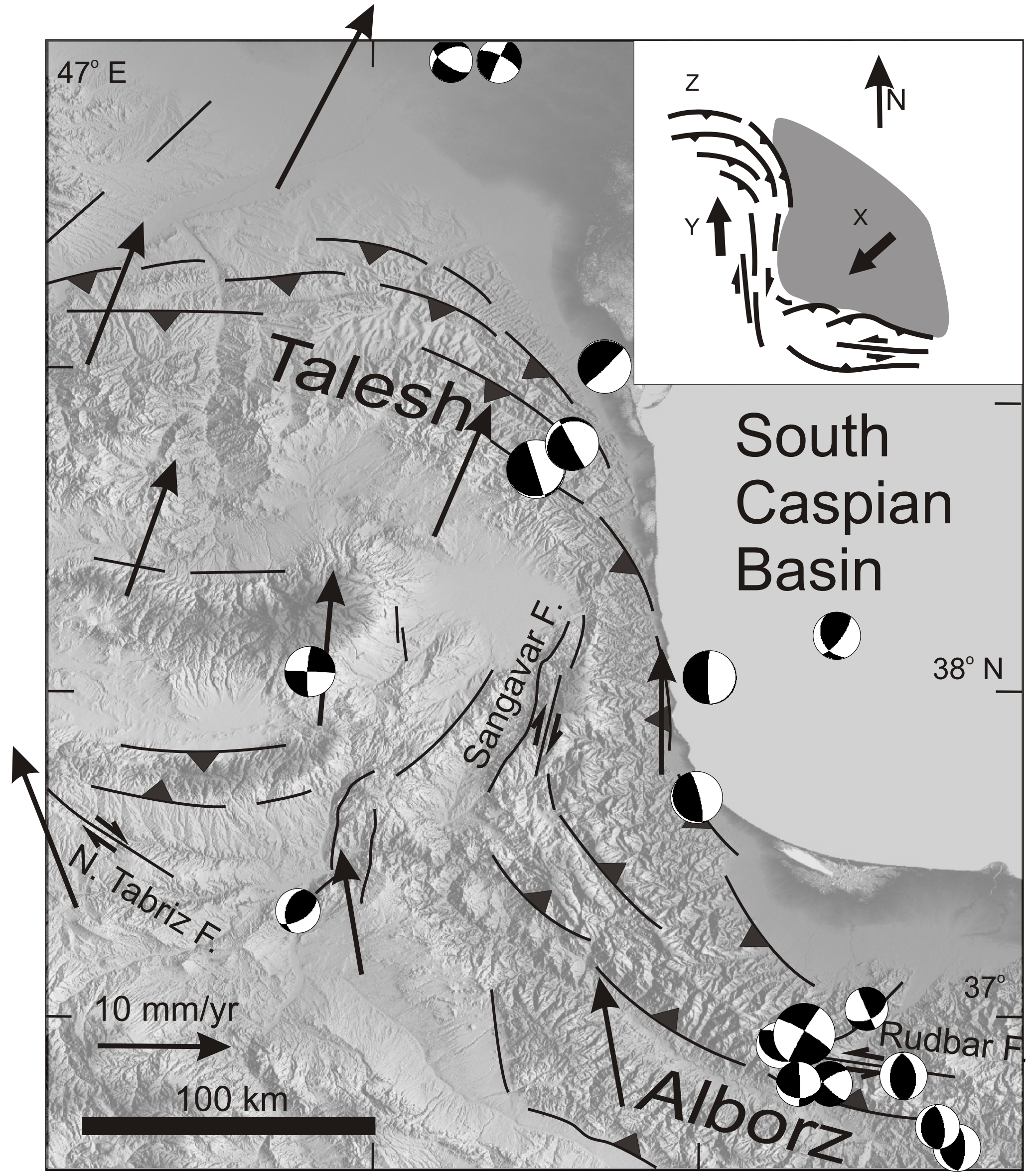


A)
d
d
θ_0
θ_0
T_0
W_0
B)
D
D
s
d
s
d
θ_1
T_1
W_1
C)
θ_1
ΣD
Σs
θ_0
Σd

7.


50° E
33° N
28° N
54° E
Main Recent
Deh Shir
Kazerun
Ardakan
Zagros Simple Folded Zone
Kareh Bas
Shiraz
Sabz Pushan
Sarvestan

8.



47° E
Z
N
Y
X
Talesh
South
Caspian
Basin
Sangavar F.
38° N
N. Tabriz F.
37°
10 mm/yr
Rudbar F.
100 km
Alborz