Comparisons of the kinematics and deep structures of the Zagros and Himalaya and of the Iranian and Tibetan plateaus and geodynamic implications

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COMPARISONS OF THE KINEMATICS AND DEEP STRUCTURES OF THE ZAGROS AND HIMALAYA AND OF THE IRANIAN AND TIBETAN PLATEAUS AND GEODYNAMIC IMPLICATIONS

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[1] We compare the geologic histories, the deep structures, and the present-day kinematics of deformation of the Himalaya and the adjacent Tibetan Plateau with those of the Zagros and Iranian Plateau to test geodynamic processes of continental collision. Shortly after India and Arabia collided with Eurasia, horizontal shortening manifested itself by folding and thrust faulting of sedimentary rock detached from India's and Arabia's underlying crystalline basement. Subsequently, slip on thrust faults stacked slices of India's basement to build the Himalaya on India's northern margin. Such faulting has not yet developed in the Zagros, where collision is more recent and Arabia penetrates into Eurasia more slowly than India does, so that postcollision convergence with Eurasia is less. The greater elevation, thicker crust, and more marked heterogeneity of the upper mantle beneath the Tibetan than beneath the Iranian Plateau also reflect a more advanced stage of development. Moreover, while thrust or reverse faulting and crustal shortening continue on the margins of both plateaus, normal faulting, suggesting horizontal extension and crustal thinning, occurs within Tibet but not in Iran. Hence, the balance of forces that built the high Tibetan Plateau must have changed, apparently some time since ~15 Ma. Removal of Tibetan mantle lithosphere could have altered that balance. If mantle lithosphere beneath the Iranian Plateau has been removed, however, the change in force balance has been too small to initiate normal faulting. Low seismic wave speeds in the uppermost mantle just beneath the Moho of both plateaus suggest (to us) that lithosphere beneath both is thin, consistent with late Cenozoic removal of it, but alternative explanations might account for these low speeds. Despite its apparently thin, and hence presumably weak, mantle lithosphere, much of central Iran undergoes little deformation. It illustrates how a crustal block can behave rigidly not necessarily because it is strong but because deviatoric stresses can be small. Whereas differences between the two regions clearly depend on the amount that Arabia and India have penetrated into Eurasia, which scales with both the dates of collision and rates of convergence, we see no differences in the operative processes that depend on the present-day rates of convergence.

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1. INTRODUCTION

[2] The two most prominent mountain belts built by continental collision, the Zagros and the Himalaya, illustrate many of the same processes but at different rates, with different amounts of total convergence, and apparently at differing stages of development. Thus, comparisons of their present-day structures and of the kinematics of their ongoing deformation offer both illustrations of processes common to collisions in general and differences that reveal aspects of collisional zones as they evolve.

[3] Both the Zagros and the Himalaya have been built by the collision of what appears to be strong lithosphere (beneath Arabia and India) with seemingly weaker material that included segments of Andean-type margins along the southern edge of the Eurasian plate. For the Zagros, the Arabian platform, which has been stable since Precambrian time and lies adjacent to the Arabian shield, plunges beneath the crust of central Iran, which has progressively become part of Eurasia. For the Himalaya the Indian shield has been underthrust beneath southern Tibet. Collision in the Zagros

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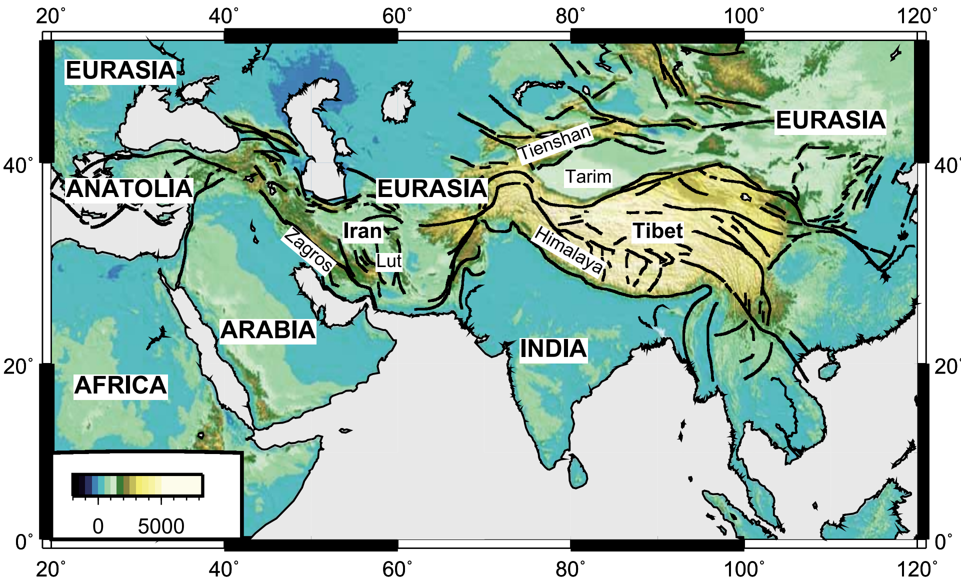


Figure 1. Map of Asia showing topography and major faults. The two belts, Zagros and Himalaya, lie adjacent to stable regions of Arabia and India, where elevations are low, and to higher terrain in the Iranian Plateau, northeast of the Zagros, and in the Tibetan Plateau north and east of the Himalaya. Both plateaus are bounded on their other sides by mountain belts. Tibet is notably higher than Iran, and high terrain extends much farther north and northeast of Tibet than it does from Iran.

seems to have begun sometime between the end of the Eocene epoch and before the Miocene epoch, between ~35 and ~23 Ma. Most believe that India's collision with Eurasia occurred between 55 and 45 Ma (though see section 2.2 for a discussion of controversy here). The mountain belts are manifestations of sustained convergence as Arabia and India have penetrated into the Eurasian continent. In both cases, that convergence has resulted not only in the building of mountain ranges over the northeastern edges of the Arabian and Indian plates but also in widespread deformation 1000–3000 km from the collision zones. Both belts are now bounded by high plateaus, high surfaces with modest relief compared with their mean elevations, except on their margins (Figure 1). Because the collision between India and Eurasia occurred before that between Arabia and Eurasia and because the average convergence rates between the former are higher (50–60 mm/yr since 45 Ma) than the latter (18–25 mm/yr since 25 Ma), the scales of the two mountain belts and associated widespread deformation differ.

[4] We discuss both the similarities of these two regions and their differences with the goal of advancing our understanding of basic, large-scale processes of collision. Some differences reflect different stages in the penetration of the Arabian and Indian plates into Eurasia [e.g., *Barazangi*, 1989], but others offer tests of how different properties of the lithosphere or different kinematics in the two regions manifest themselves in the resulting widespread deformation. We exploit these differences to address how both crustal and mantle dynamics may evolve during the collision and penetration of one continent into another. Readers will be quick to note that aspects of both regions remain poorly known or disputed, and we try to make controversies apparent.

2. BACKGROUND

2.1. Present-Day Topography and Geographical Differences

[5] The Tibetan and Iranian plateaus dominate topographic maps not only of Asia but also of the entire Alpide belt of Mesozoic and Cenozoic mountain building that stretches from the Iberian peninsula and Morocco eastward to Indonesia (Figure 1). Both plateaus are bounded on the south and southwest by important mountain ranges, the Zagros and Himalaya, which have formed in Cenozoic time. To the north and northeast of the plateaus, additional high terrain reflects continuing crustal deformation, which includes crustal thickening and continued mountain building.

[6] Both regions of high terrain are the loci of major earthquakes that claim lives at rates among the highest on Earth and that attest to continuing tectonic activity (Figure 2). Most great intracontinental earthquakes have occurred in eastern Asia and not in Iran (Figure 2). Presumably, this results, at least in part, from more rapid ongoing deformation in eastern Asia than Iran but perhaps also from a longer history and greater amount of deformation that have created more mature faults with longer homogeneous segments in eastern Asia.

[7] In terms of deaths, however, when scaled to areal extent, the populations of Iran and eastern Asia have suffered comparable losses, in part because so much of eastern Asia is remote and unpopulated. Moreover, people living in

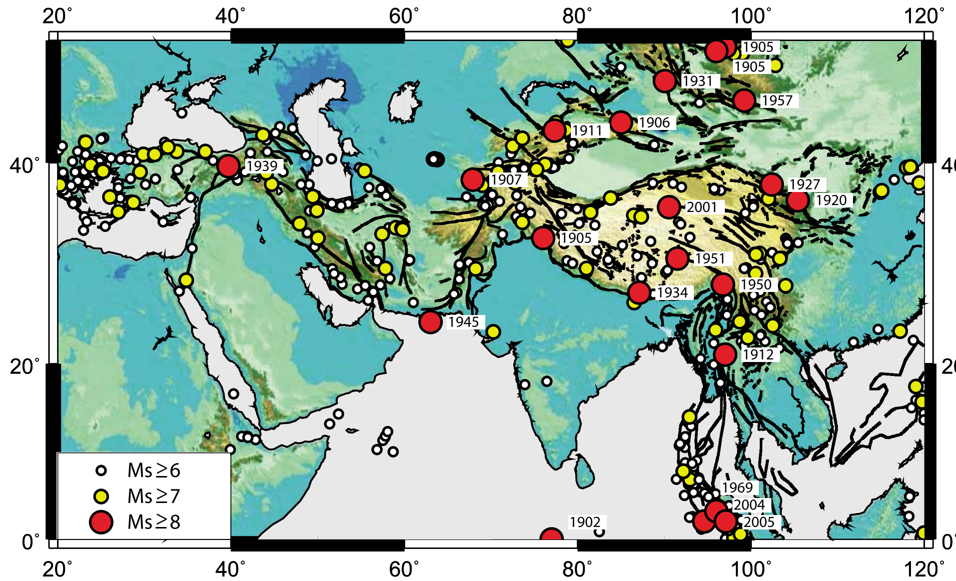


Figure 2. Map of Asia showing significant earthquakes between 1900 and 2005 and those between 1973 and 2005 with $M \geq 6$ from compilations by the U.S. Geological Survey (http://neic.usgs.gov/neis/epic/epic_global.html) but with modifications from *Molnar and Deng* [1984] for eastern Asia and from *Ambraseys and Jackson* [1998] for the eastern Mediterranean. For events with $M \geq 8$, we have included a couple with $M = 7.9$ (e.g., 1931 and 2001). Note that the largest earthquakes tend to occur in eastern Asia.

arid climates like in Iran build houses in areas where there its water, which in turn commonly lie at the edges of mountain ranges, not only where rain falls but also where slip on active faults, during earthquakes, maintains those topographic differences [*Jackson*, 2006].

[8] Not all of the high topography in these regions has developed since the collisions of Arabia and India with Eurasia, but most surely has. Although the simple geographical concept of a plateau applies to both regions, mountain ranges of varying heights and widths surround both plateaus, and these ranges are loci of active deformation. Currently, internal drainage occurs in both plateaus, and hence, erosion of elevated terrain manifests itself as a reduction in relief as debris is deposited in basins adjacent to the eroding terrain.

[9] The two plateaus also differ, most obviously in Tibet's mean elevation being nearly 5000 m but with virtually all of Iran lying at lower elevation and the mean elevation for the Iranian Plateau being only ~1000–1500 m (see Table 1 for a summary of quantitative comparisons). Similarly, although the widths of the Zagros and Himalaya are comparable (~200–300 km), the Himalaya stands much higher. Moreover, the Himalaya has much greater access to moist air (from the Arabian Sea and Bay of Bengal) than the Zagros, which lie downwind of the Arabian desert. With orographic rainout of that moisture, erosion of the Himalaya seems to be much faster than that of the Zagros. Glaciers carve high terrain in the Himalaya but have had little impact on the Zagros. Canyons in the Himalaya are deeper. Correspondingly, sediment accumulation is modest in the minor foredeep southwest of the Zagros (including the Persian Gulf with only ~50 m of water and a maximum depth of 90 m); by contrast, a relatively deep basin south and southwest of the Himalaya, the Ganga Basin, is full, and most sediment today passes over it and continues to the Indus and Bengal fans in the Arabian Sea and Bay of Bengal.

TABLE 1. Quantitative Comparison of the Two Mountain Belts and Plateaus

	Himalaya-Tibet	Zagros-Iran
Width of mountain belt	200–300 km	200–300 km
Crustal thickness beneath plateaus	~70 km (over a wide area)	<50 km (except in narrow belts)
Mean elevation	5000 m	1000–2000 m
Date of collision	45–55 Ma	23–35 Ma
Postcollision tectonic event	~15 Ma	~12 Ma
Onset of normal faulting	13–8 Ma	in the future?
Precollision rate	110 mm/yr	31 mm/yr
Postcollision rate	50–60 mm/yr	25 mm/yr
Posttectonic rate	32–44 mm/yr	20 mm/yr
Himalaya/Zagros shortening rates	18–20 mm/yr	6–10 mm/yr
Postcollision plate convergence	2500–3500 km	500–800 km
Himalaya/Zagros shortening	300–900 km	50–150 km

2.2. Precollision Geologic History and the Timing of Collisions

[10] Since Paleozoic time, the Asian continent has grown larger with the collisions of fragments broken from Gondwana in late Paleozoic time and successively added to the southern margin in early Mesozoic time [e.g., *Şengör*, 1979a, 1984; *Şengör et al.*, 1988; *Stöcklin*, 1968], a result amply confirmed by paleomagnetic studies [e.g., *Besse et al.*, 1998; *McElhinny et al.*, 1981; *Muttoni et al.*, 2009; *Soffel and Förster*, 1980; *Soffel et al.*, 1975, 1996; *Wensink*, 1979, 1982, 1983; *Wensink et al.*, 1978; *Wensink and Varekamp*, 1980]. Some fragments, like the Tarim Basin north of Tibet or the Lut Block within the Iranian Plateau (Figure 1) [e.g., *Stöcklin*, 1968, 1974, 1977], define relatively intact blocks of crust apparently sufficiently strong to have resisted deformation since they joined Eurasia. Some fragments, however, are so small that they must consist of only upper crust, detached from the underlying, presumably strong, mantle lithosphere, and others comprise deformed belts that mark ancient sutures and subduction zones, such as those trending east-west across Tibet [e.g., *Allégre et al.*, 1984; *Chang and Cheng*, 1973; *Chang et al.*, 1986; *Matte et al.*, 1996], in northern Pakistan [e.g., *Tahirkheli et al.*, 1979; *Tapponnier et al.*, 1981] and the Pamir [e.g., *Burtman and Molnar*, 1993], or within Iran [*Stöcklin*, 1968, 1974, 1977]. Thus, throughout Mesozoic time, southern Asia was in a state of continual growth and modification as fragments accreted. Moreover, the collisions of Arabia and of India with Eurasia were preceded by tens of millions of years of subduction of oceanic lithosphere, which built volcanic belts, thinned mantle lithosphere, and also may have added crust to Asia.

[11] As a result of these late Paleozoic and Mesozoic additions to southern Eurasia and of Cenozoic subduction of oceanic lithosphere beneath the southern margin of Asia, southern Eurasia appears to have been underlain by weaker lithosphere than that beneath the Precambrian shields and platforms that underlie India, Arabia, and most of Eurasia north of Iran or Tibet [e.g., *Molnar and Tapponnier*, 1981]. Here, however, there seems to be an important structural difference between precollisional Iran and Tibet. Idealized cross sections across the Sanandaj-Sirjan Zone of Cretaceous and early Cenozoic arc-like magmatic rock in Iran show a normal crustal thickness [e.g., *Ahmadian et al.*, 2009; *Omrani et al.*, 2008]. Cretaceous and early Cenozoic marine sedimentary rock crops out over much of Iran, and although unconformities and lacunae attest to emergent regions, there seems to be little doubt that most of this region lay near sea level until approximately Oligocene time [e.g., *Berberian and King*, 1981; *Davoudzadeh et al.*, 1997; *Morley et al.*, 2009; *Reuter et al.*, 2009; *Schuster and Wielandt*, 1999; *Stöcklin*, 1968, 1974, 1977]. Moreover, *Verdel et al.* [2007] inferred intense, east-west crustal extension, normal faulting, and therefore presumably crustal thinning during Eocene time (40–44 Ma). By contrast, much evidence suggests that southern Tibet behaved as an Andean margin bounded by thrust faults, underlain by thick granitic intrusions, and presumably with thick crust and high elevation, as in the present-day central Andes [e.g., *Burg and Chen*, 1984; *Burg et al.*, 1983; *Dürr*, 1996; *England and Searle*, 1986; *Kapp et al.*, 2003; *Murphy et al.*, 1997; *van der Beek et al.*, 2009; *Volkmer et al.*, 2007]. Lithosphere, weakened by thermal processes associated with subduction and still warm following accretion of fragments, then facilitated deformation of southern Eurasia as India and Arabia penetrated into this region [e.g., *Molnar and Tapponnier*, 1981] but with Tibet apparently starting with a high, if narrow, plateau and most of Iran lying near sea level.

[12] Arabia and India are not only two of the largest fragments to be added to Asia since early Paleozoic time but also the most recent such fragments. For both the Zagros and the Himalaya, it appears that continents with rather typical rifted margins followed oceanic lithosphere into subduction zones and collided with southern Eurasia.

[13] Southwest of the Main Zagros Thrust, a wide zone of folded sedimentary rock crops out along the entire length of the Zagros. Sediment accumulated on the northeastern margin of the Arabian platform from Precambrian through Cretaceous time initially on a stable platform and later on a continental margin after fragments of crust rifted from the Arabian platform in Permian and Jurassic time [e.g., *Alavi*, 2004; *Colman-Sadd*, 1978; *Falcon*, 1974; *Koop and Stoneley*, 1982; *Stöcklin*, 1968, 1977; *Trowell*, 1995]. From analyses of sequences of sediment in many tens of drill cores, *Trowell* [1995] calculated that the region had undergone a total of ~10%–25% extension largely in two phases, in Permian and Jurassic times. Studies of crustal structure support the rifting interpretation: numerous estimates of crustal thickness based on seismic refraction [e.g., *Mooney et al.*, 1985], surface wave dispersion [*Al-Amri*, 1999; *Mokhtar et al.*, 2001; *Rodgers et al.*, 1999], receiver functions [*Al-Damegh et al.*, 2005; *Al-Lazki et al.*, 2002; *Pasyanos et al.*, 2007], or both of the latter methods combined [*Gök et al.*, 2008] suggest that crustal thicknesses in central Arabia exceed 40 km and may reach 45 km in places, but the thickness of the crystalline crust beneath the sediment just southwest of the Zagros (in Iraq and Kuwait) is only ~35 km, like that beneath the Zagros [*Hatzfeld et al.*, 2003; *Paul et al.*, 2006, 2010] (as discussed in section 3.3). Only in Cenozoic time has sediment derived from the Iranian side contributed to the sedimentary sequence in the Zagros.

[14] Comparable detail does not seem to exist for much of the ancient margin of India, but few doubt that a rifted margin existed before collision [e.g., *Corfield et al.*, 2005; *Garzanti*, 1999; *Liu and Einsele*, 1994; *Myrow et al.*, 2003; *Wiesmayr and Grasemann*, 2002]. *Gaetani and Garzanti* [1991] reported a history for the northwestern Himalaya resembling that of Arabia: shallow water deposition until Permian time, when rifting seems to have created a continental margin. Virtually all descriptions of the sedimentary sequence farther southeast include a similar history of sedimentation first on a platform and then on a rifted margin (also in Late Permian or early Triassic and again in Jurassic time, as for the Arabian margin) [e.g., *Burg and Chen*, 1984; *Corfield et al.*, 2005; *Garzanti*, 1999; *Liu and Einsele*, 1994;

Ratschbacher et al., 1994; *Searle*, 1983; *Steck*, 2003; *Steck et al.*, 1993a, 1993b; *Wiesmayr and Grasemann*, 2002]. Subsequent deformation, however, has left little evidence that constrains the details of the precollisional structure of the margin [e.g., *Garzanti et al.*, 2005], but thicknesses of sedimentary rock reach ~10 km and call for several kilometers of crust thinning during rifting [e.g., *Corfield et al.*, 2005; *Liu and Einsele*, 1994].

[15] As noted in section 1, the collision between India and Eurasia preceded that between Arabia and Eurasia. For both, however, controversies continue to highlight inconsistencies in the data used to infer the date of collision. For the Himalaya, most agree that collision occurred some time between 45 and 55 Ma, the age of the last marine sediment deposited on the northern margin of India [e.g., *Garzanti and Van Haver*, 1988; *Green et al.*, 2008; *Khan et al.*, 2009; *Najman*, 2006; *Rowley*, 1996, 1998; *Zhu et al.*, 2005], but recently, *Aitchison et al.* [2000, 2007, 2008] have challenged this view. They suggested that what is commonly treated as the collision between India and Eurasia actually was a collision between India and an island arc that lay south of Eurasia. They argued that the last seafloor vanished as recently as ~34 Ma. Indeed, the northwestern part of India seems to have collided with an arc before final closure of an ocean but much earlier than 34 Ma [e.g., *Tahirkheli et al.*, 1979; *Tapponnier et al.*, 1981]. *Coward et al.* [1987], for example, inferred an arc-continent collision in this region at 90–100 Ma. More recently, *Khan et al.* [2009] argued that the Kohistan arc in the Pakistan Himalaya was accreted to the Indian subcontinent near 60 Ma and both collided with Eurasia near 50 Ma. Similarly, *Ding et al.* [2005] inferred the emplacement of ophiolites onto India at ~65 Ma, but they too deduce that collision with southern Eurasia had occurred by 45 Ma. Following *Garzanti* [2008], we accept the evidence for collision closer to 50 Ma than 34 Ma as sound, but readers should note that this issue is not closed.

[16] The Zagros share aspects of Himalayan structure, but opinions about their history also differ. Some have argued for a Cretaceous collision [*Alavi*, 1994; *Berberian and King*, 1981; *Mohajjel and Fergusson*, 2000], some have suggested an early Cenozoic age [e.g., *Ghasemi and Talbot*, 2006; *Homke et al.*, 2009; *Mazhari et al.*, 2009], and some inferred from other data that collision manifested itself only as recently as ~5 Ma [e.g., *Allen et al.*, 2004; *Blanc et al.*, 2003; *Falcon*, 1974; *Haynes and MacQuillan*, 1974; *Kashfi*, 1976; *Ricou et al.*, 1977], but opinions are now converging on collision during the period between late Eocene and early Miocene, as suggested by *Koop and Stoneley* [1982]. Potential suture zones follow the Main Zagros Thrust [e.g., *Agard et al.*, 2005; *Haynes and MacQuillan*, 1974; *Ricou et al.*, 1977] and along the Sanandaj-Sirjan Zone to its northeast [e.g., *Alavi*, 1994; *Pamić et al.*, 1979]. *Agard et al.* [2005, 2006] inferred from dates of blueschist, intrusions, and ophiolitic material that subduction had occurred until ~35 Ma and that collision occurred before 23–25 Ma. Subsequent thrust slip on the Main Zagros Thrust (also known as the "Zagros Crush Zone" for the rock within it has been highly deformed) seems to have carried intact Arabian crust beneath the ancient southern margin of Eurasia. *Agard et al.* [2005] inferred 50–70 km of such convergence from repetitions of overthrust slices of sedimentary rock and other units, but we presume that their estimate is a lower bound on such convergence. Moreover, revised dating of the sedimentary sequence in the Zagros shows that beginning in early Miocene, and perhaps late Oligocene time (20–25 Ma), material eroded from terrain northeast of the present-day Zagros, and hence from the Eurasian side, was deposited on the ancient Arabian margin [*Ballato et al.*, 2010; *Fakhari et al.*, 2008]. Using zircons derived from the adjacent magmatic belt and deposited north of the Zagros, *Horton et al.* [2008] also argued for this date of collision. Finally, *Hessami et al.* [2001b] showed that folding in the Zagros had begun before late Miocene time, despite widespread belief that it began since that time.

[17] Part of the confusion over timing of collision in the Zagros has arisen because suturing may have involved two events, collision of an arc in Mesozoic time and with the Arabian platform later [e.g., *Ghasemi and Talbot*, 2006]. Controversy surrounds the number of ophiolite suites and ocean basins that closed in this region and keeps open the possibility of a date of collision earlier than late Eocene [e.g., *Mazhari et al.*, 2009; *Mohajjel et al.*, 2003]. Moreover, different phases of deformation within the Sanandaj-Sirjan Zone, which include not only thrust slip and convergence but also an important component of right-lateral shear [e.g., *Mohajjel and Fergusson*, 2000; *Sarkarinejad and Azizi*, 2008; *Sarkarinejad et al.*, 2008], make it difficult to distinguish the termination of subduction from other styles of deformation.

3. KINEMATICS

3.1. Kinematics of Plate Motions and the Early History of Deformation

[18] Plate reconstructions for both Arabia [*McQuarrie et al.*, 2003] and India [e.g., *Copley et al.*, 2010; *Dewey et al.*, 1989; *Molnar and Stock*, 2009; *Molnar and Tapponnier*, 1975; *Molnar et al.*, 1993; *Patriat and Achache*, 1984] with respect to Eurasia show sustained convergence for tens of millions of years before either collision occurred (Figure 3). In both regions decreases in convergence rates occurred near the times of collision. For the Himalaya, the convergence rate dropped 30%–40% near 45 Ma, from ~120 to ~80 mm/yr in the northeastern corner of India and from ~110 to ~60 mm/yr in the northwestern corner (Figure 4) [*Molnar and Stock*, 2009]. Subsequently, between 20 and 10 Ma, the convergence rates seem to have slowed another ~45% (~83 to ~44 mm/yr and ~59 to ~34 mm/yr in the two corners of India). Similarly, Arabia's convergence rate slowed by ~35% since ~20 Ma, following collision with Eurasia (from ~31 mm/yr to ~20 mm/yr for a point in the northwestern part of the Zagros) (Figure 4) [*McQuarrie et al.*, 2003]. The simple interpretation is that when collision occurred, the buoyancy of continental crust inhibited continued subduction, and with continued convergence, the

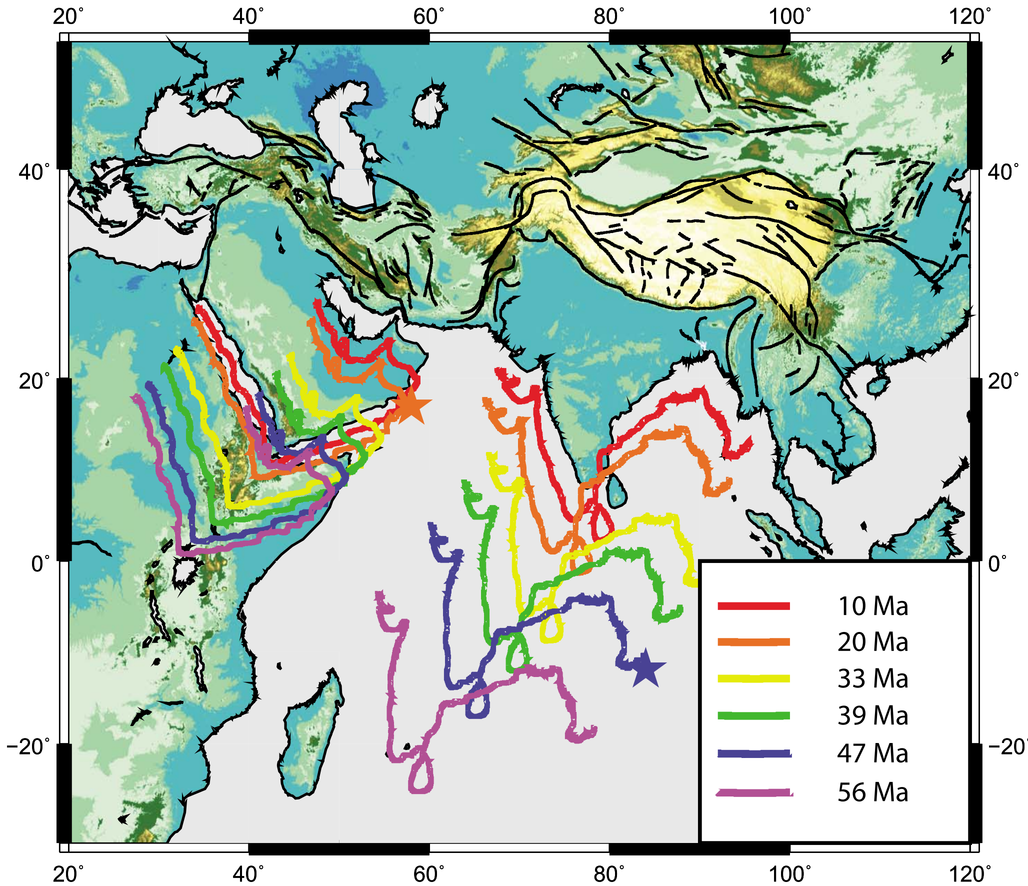


Figure 3. Positions of coastlines of Arabia and India with respect to Eurasia reconstructed to their positions at different times in the past, using parameters given by *McQuarrie et al.* [2003] and *Molnar and Stock* [2009]. Stars mark positions of coastlines of the two continents at the approximate times of their collisions with Eurasia.

force per unit length along the subduction zone provided by the weight of downgoing slabs decreased [e.g., *Copley et al.*, 2010; *Molnar and Tapponnier*, 1975].

[19] (Note that *Copley et al.* [2010], who used different reconstructions in the various oceans than did *Molnar and Stock* [2009], reported no major change in rate between 20 and 10 Ma but rather a monotonic reduction in convergence rate since ~50 Ma. The set of reconstructions used by *Copley et al.* [2010] would also call for a history of convergence between Arabia and Eurasia different from that of *McQuarrie et al.* [2003]. We proceed with the assumption that the reconstructed histories of plate motion lie within the uncertainties given by *McQuarrie et al.* [2003] and by *Molnar and Stock* [2009] and as shown in Figures 3 and 4.)

[20] Despite these similarities in the histories of convergence, two differences between the Zagros and the Himalaya stand out in Figure 4. First, India's average rate of penetration into Eurasia since collision has been roughly twice that of Arabia's rate (~50–65 mm/yr versus ~25 mm/yr), which makes all processes due to collision and penetration occur faster in the Himalaya and surrounding regions than in Iran. Second, the earlier collision between India and Eurasia requires not only a longer period over which the region responded to collision but also, because of the higher rate, a greater penetration of India than Arabia into Eurasia. Among consequences of these differences, the most obvious is the difference in mean elevations [e.g., *Ben Avraham and Nur*, 1976].

[21] It appears that deformation occurred across much of the Tibetan Plateau shortly after the India-Eurasia collision and somewhat later in the Iranian Plateau after the Arabia-Eurasia collision. For Tibet, numerous observations, summarized by *Dayem et al.* [2009], show deformation in northern Tibet shortly after collision. For the Zagros and the Iranian Plateau, observations and constraints on timing seem to be fewer than for Tibet, but some suggest that deformation has occurred far from the Zagros since late Miocene time, if not earlier. *Morley et al.* [2009] describe some 38 km of shortening between the Zagros and Alborz since late Miocene time, in the region just west of central Iran. Deformation in the Alborz Mountains seems to have begun later, near 20–17.5 Ma according to *Ballato et al.* [2008, 2010] or perhaps more recently at 12 Ma [*Guest et al.*, 2007], when exhumation there accelerated [*Guest et al.*, 2006]. From estimates of amounts of slip and present-day rates of slip for faults in the Kopet Dagh, *Hollingsworth et al.* [2006, 2008] suggested that the current phase of tectonic activity dates from ~10 Ma, though *Shabanian et al.* [2009a]

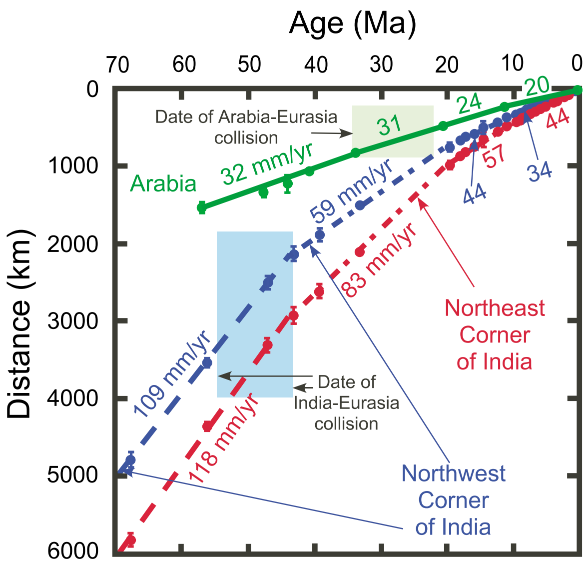


Figure 4. History of convergence between Arabia and Eurasia and between India and Eurasia. Two points on India (27°N, 95°E, and 30°N, 72°E) and one on Arabia (32.70°N, 50.38°E) were reconstructed to their positions at different times in the past (by *Molnar and Stock* [2009] and by *McQuarrie et al.* [2003], respectively), and here the distances that the points subsequently moved are plotted versus those times. Note the earlier collision, the faster convergence, and the greater amount of movement since collision between India and Eurasia than between Arabia and Eurasia.

argued for a more recent, more important change in tectonic style in the Kopet Dagh.

3.2. Geological Structure and Evolution of the Zagros and Himalaya

[22] In Figure 5, we contrast simplified cross sections across the Zagros and Himalaya. These are drawn to emphasize differences in the deeper structure, at the expense of details of folding in the uppermost layers, particularly in the Zagros.

[23] Although the large-scale structure of the Himalaya recognized by *Heim and Gansser* [1939] and *Gansser* [1964] has undergone modification, their basic image remains valid to first order (Figure 5). The suture between India and the southern margin of Eurasia now lies within the Tibetan Plateau. To its north, adjacent to virtually the entire suture zone, a belt of granitic rock records tens of millions of years of arc volcanism and subduction of oceanic lithosphere. To its south, between the high peaks of the Himalaya and the suture, one finds a sequence of sedimentary rock that was deposited on the northern edge of India from Paleozoic until early Cenozoic time. This sedimentary rock was deformed in a typical fold-and-thrust belt [e.g., *Burg and Chen*, 1984; *Corfield et al.*, 2005; *Ratschbacher et al.*, 1994; *Wiesmayr and Grasemann*, 2002]. Along much of the Himalaya, a system of gently northward dipping (essentially inactive) normal faults separate that sequence from the basement on which it was deposited [e.g., *Burchfiel et al.*, 1992; *Burg and Chen*, 1984; *Burg et al.*, 1984; *Dézes et al.*, 1999; *Herren*, 1987; *Valdiya and Goel*, 1983]. The crystalline rock of the Greater Himalaya has been thrust southward a minimum of ~100 km on the Main Central Thrust [*Brunel*, 1975; *Brunel and Andrieux*, 1977; *Fuchs and Sinha*, 1978; *Gansser*, 1966; *Pearson and DeCelles*, 2005; *Srivastava and Mitra*, 1994; *Steck*, 2003; *Steck et al.*, 1993a, 1993b; *Stöcklin*, 1980] and apparently much more [e.g., *DeCelles et al.*, 2002; *Robinson et al.*, 2006] onto the Lesser Himalaya, though some contend that placing tight constraints on values more than a few hundred kilometers is not possible [e.g., *Avouac*, 2003; *Hodges*, 2000; *Steck*, 2003]. We ignore here the fact that the Main Central Thrust consists of splays of faults, which now can be mapped as separate structures with substantial slip on each [e.g., *Burg and Chen* 1984; *Célérier et al.*, 2009a, 2009b; *Pearson and DeCelles*, 2005; *Steck et al.*, 1993a, 1993b; *Valdiya*, 1980a, 1980b]. Rock cropping out in the Lesser Himalaya, below klippen of crystalline rock from the Greater Himalaya, consists largely of sedimentary rock that has been weakly metamorphosed, if metamorphosed at all. The age of this sedimentary rock is poorly constrained in most regions, but where dated, it ranges from late Precambrian to as young as Miocene [e.g., *DeCelles et al.*, 1998; *Myrow et al.*, 2003; *Najman et al.*, 2002; *Pearson and DeCelles*, 2005; *Sakai*, 1983]. At the southern front of the range, the entire package of rock comprising the Lesser and Greater Himalaya has been thrust onto the Indian shield and onto sedimentary rock deposited on it. A wide sedimentary basin, the Ganga Basin overlain by the Indo-Gangetic Plain, lies south and southwest of the Himalaya and contains as much as 4–5 km of sediment derived largely by erosion of the Himalaya [*Karukaranan and Ranga Rao*, 1979; *Mathur and Evans*, 1964; *Raiverman et al.*, 1983; *Sahni and Mathur*, 1964; *Sastri et al.*, 1971].

[24] This structure, which characterizes most of the ~2500 km length of the Himalaya, results from a sequence of events that includes suturing at 55–45 Ma, with ophiolitic mélange thrust onto the Indian margin [e.g., *Gansser*, 1964; *Heim and Gansser*, 1939; *Le Fort*, 1975; *Mattauer*, 1975] and with folding of the sedimentary sequence beginning at that time [*Wiesmayr and Grasemann*, 2002]. Moreover, melting and intrusion of granite at 44 Ma into the sequence of sedimentary rock deposited on the northern margin of the Indian subcontinent attests to underthrusting of that rock to depth shortly after the collision [*Aikman et al.*, 2008]. Some time later, but not yet unambiguously dated, slip on the Main Central Thrust carried rock comprising the northern edge of the Indian shield southward onto the intact part of the shield. Sediment deposited both in the Bengal fan [*Galy et al.*, 1996] and in the Bengal Basin of Bangladesh [*Najman et al.*, 2008] contain metamorphic minerals that cooled from hotter than 350°C before ~30 Ma and whose isotopic content is similar to that of rock now exposed in the Greater Himalaya. Thus, a relatively deeply eroded, and hence possibly high, Himalayan range seems to have existed by ~30 Ma. The Main Central Thrust was active at ~20 Ma, for an apparently poorly constrained duration before then, and apparently until ~12 Ma [e.g., *Hodges et al.*, 1996;

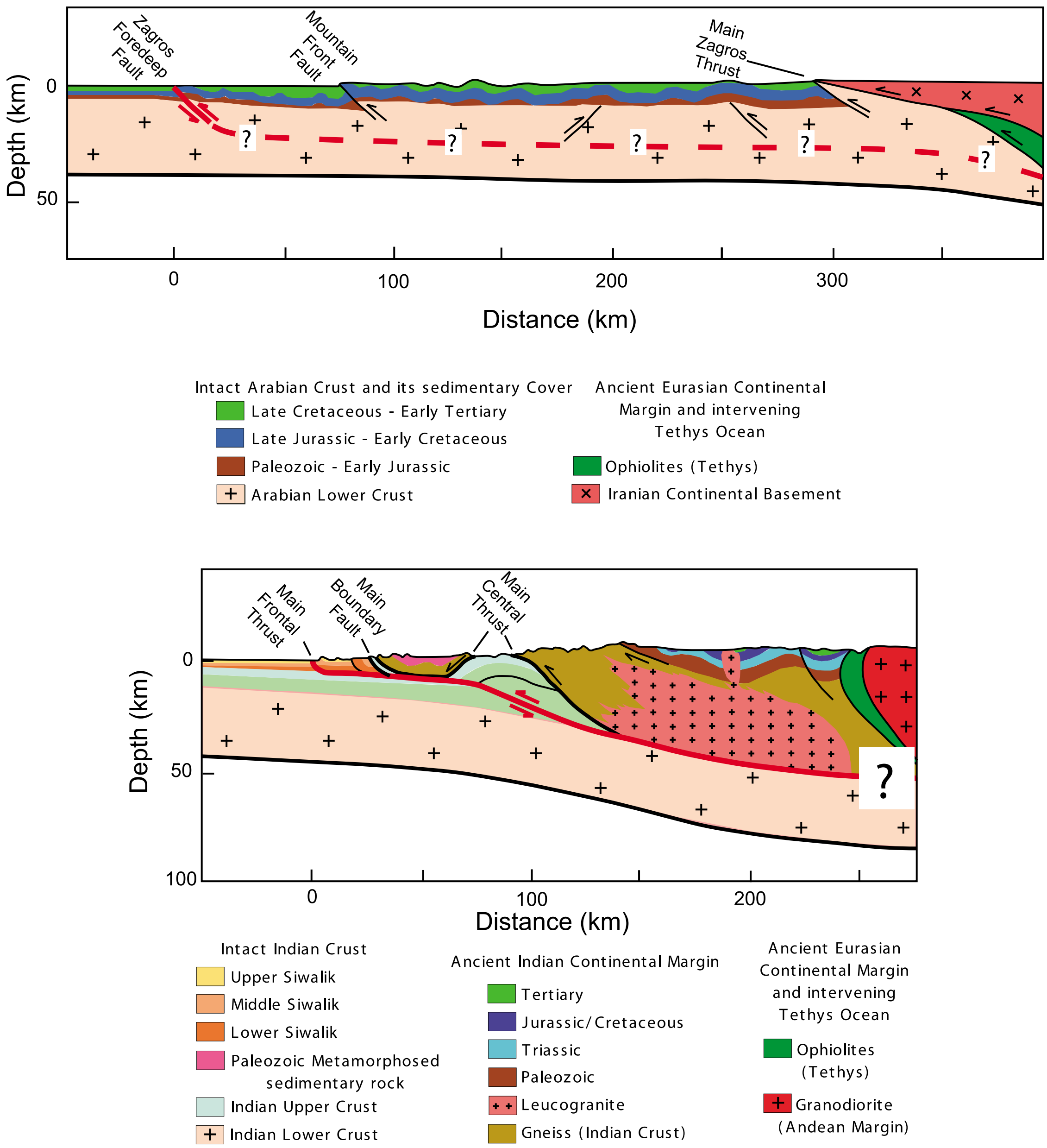


Figure 5. Simplified geologic cross sections across the (top) Zagros and (bottom) Himalaya, drawn to include the entire crust. The cross section for the Himalaya is modified from one given by *Avouac* [2003], and for the Zagros we used one from *Mouthereau et al.* [2007] as a guide. Note that the Zagros are built almost entirely of folded sedimentary rock deposited on the Arabian platform, but in the Himalaya major thrust faults cut the entire crust. The portion of the Himalaya equivalent to the Zagros currently crops out between the crystalline rock of the Greater Himalaya and Indus-Tsangpo suture. Reverse faults in the basement of the Zagros may be coalescing into a shear zone of finite width (shown by a red dashed line with question marks) within the middle to lower crust of Arabian platform beneath the sedimentary rock.

Hubbard and Harrison, 1989; *Kohn et al.*, 2001, 2004; *Pearson and DeCelles*, 2005], but little evidence suggests that it is active now, and it is certainly not active as a major fault. Although most seismicity, including both moderate earthquakes [e.g., *Baranowski et al.*, 1984; *Ni and Barazangi*, 1984] and microearthquakes [*Pandey et al.*, 1995] and discussed further in section 3.3, occurs beneath the surface trace of the Main Central Thrust, most of these earthquakes take place at depths near ~15 km, and most of those with reliable fault plane solutions show slip on planes dipping gently northward or northeastward. Thus, rapid active slip must occur on a fault below the Main Central Thrust.

[25] As noted in section 1, both the Zagros and Himalaya have been built adjacent to suture zones and to magmatic belts associated with subduction of oceanic lithosphere before collision occurred. Beyond these similarities, the surface

geology of the Zagros bears little resemblance to that of the Himalaya. Virtually all rock exposed in the Zagros is sedimentary in origin, deposited on the northeastern margin of Arabia. The Zagros have been built by northeast-southwest shortening, which has occurred largely by folding, and hence by thickening, of that shallow sedimentary rock approximately in isostatic equilibrium (Figure 5). Layers have been detached along weak horizons of evaporites, and for many geologists, the Zagros exemplify thin-skinned deformation [e.g., *Colman-Sadd*, 1978; *Hessami et al.*, 2001b; *McQuarrie*, 2004; *Molinaro et al.*, 2005; *Oveisi et al.*, 2007, 2009]. The idea that the sedimentary rock has been detached from the basement seems inescapable, given the open folding of it and the widespread presence of weak horizons at different depths, including abundant salt. In addition, unlike the Himalaya, whose width varies little along the ~2500 km long chain, the width of the Zagros in the southeast is virtually double that to its northwest, northwest of the Kazerun line. (See Figure 7a for the location of the Kazerun line, which, in present-day tectonics discussed in section 3.3, defines a roughly north-south zone of right-lateral shear.) Accordingly, the regional average slope in the southeast is approximately half that in the segment northwest of that line (Figure 8). Widespread salt domes emanating from the Cambrian Hormuz salt layer attest to the presence of a very weak layer at the base of the sedimentary sequence southeast of the Kazerun line. Their sparse presence northwest of it has led many to deduce that the greater width of the southeastern Zagros results from the presence of the Hormuz salt, which allows only low shear stress on the base of the overlying wedge of stronger sedimentary rock [e.g., *Jahani et al.*, 2009; *Mouthereau et al.*, 2006; *Sepehr and Cosgrove*, 2004, 2005; *Sepehr et al.*, 2006; *Talbot and Alavi*, 1996]. Many balanced cross sections across the Zagros, in turn, begin with the assumption that all of the deformation of sedimentary rock occurs in strata detached from the basement.

[26] Recently, a new view has emerged in which some thrust or reverse faults in the basement penetrate the sedimentary cover. Although weak horizons within the sedimentary rock redistribute strain, both recent efforts to balance cross sections across the Zagros that take into account the elevation of the basement [e.g., *Berberian*, 1995; *Molinaro et al.*, 2005; *Mouthereau et al.*, 2006, 2007] or that exploit geophysical imaging of the top of the basement [e.g., *Bosold et al.*, 2005; *Sherkati et al.*, 2005] suggest that the basement now deforms across much of the Zagros and no longer merely serves as an effectively rigid substratum. *Mouthereau et al.* [2006] buttress these arguments with an analysis of the mechanics involved in deformation of both sedimentary layers overlying salt and a straining basement. In section 3.4, we exploit this new view further.

[27] The locus of deformation in the Zagros seems to have propagated from the suture, at the Main Zagros Reverse Fault, beginning near the end of Eocene time southwestward toward the Persian Gulf shore by the Mid-Pleistocene time [e.g., *Hessami et al.*, 2001b; *Shearman*, 1976]. If we assume that the folding occurs concurrently with modest deformation of the basement, as suggested by the similar distribution of surface deformation evidenced by GPS measurements, active seismicity, and late Cenozoic folding, we might infer that faulting in the basement has also propagated in time from the inner to the outer part of the Zagros Mountains [*Hatzfeld et al.*, 2010].

[28] For eastern Asia, assuming a date of collision of 45–55 Ma, northwestern and northeastern India have moved ~2500–3500 km and 3000–4000 km (Figure 4), respectively, toward stable Eurasia since collision. Estimates of the amount of underthrusting of India beneath southern Tibet, all of which may be lower bounds, range from 300 to 700 km [e.g., *Coward and Butler*, 1985; *Coward et al.*, 1987, 1988a; *DeCelles et al.*, 2002; *Hauck et al.*, 1998; *Johnson*, 2002; *Ratschbacher et al.*, 1994; *Schelling*, 1992; *Schelling and Arita*, 1991; *Searle*, 1986; *Searle et al.*, 1997; *Srivastava and Mitra*, 1994], and *Robinson et al.* [2006] allowed for as much as 900 km. This latter value is consistent with steady underthrusting of India at its current rate of ~20 mm/yr since 45 Ma.

[29] Estimates of both crustal shortening in the Zagros and total plate convergence between Arabia and Eurasia are much smaller. Northwest of the Kazerun line balanced cross sections yield amounts of shortening from as little as 25 km [*Sherkati and Letouzey*, 2004], to ~45 km [*Bosold et al.*, 2005] and 49 km [*Blanc et al.*, 2003], and to as much as 65 km [*Sherkati et al.*, 2006] or 57 and 67 km [*McQuarrie*, 2004]. Southeast of this line they range from 45 km [*Molinaro et al.*, 2005], to 50 km [*Sherkati et al.*, 2006], and to 65–78 km [*Mouthereau et al.*, 2007] and even 85 km [*McQuarrie*, 2004]. These values approximately match the 50–70 km that *Agard et al.* [2005] inferred for thrust slip at the Main Zagros Thrust north of the Kazerun line or that *Paul et al.* [2006, 2010] inferred from the thickness of the crust. Depending upon how one interprets the relationship of that thrust faulting and the folding in the Zagros, we might treat the folding and the overthrusting as independent estimates of the same amount of convergence but manifested differently in different sections of a cross section; alternatively, we might add them, treating them as having formed during separate phases of deformation after collision. In the latter view, surfaces of décollement, where sedimentary units separated from the basement, must pass into the crust southwest of the Main Zagros Thrust on thrust faults that have not yet been defined clearly. In the former view convergence of Arabia with central Iran would occur by folding of the sedimentary rock in the Zagros and by thrust slip farther north where the surface of décollement is rooted in the deeper crust. The widespread belief, however, that folding in the Zagros is largely young, since 10 Ma, and perhaps since 5 Ma, compared with earlier thrust slip at the Main Zagros Thrust, favors treating these estimates of shortening as separate.

[30] Even if one adds the amounts for the two regions (folding of sedimentary rock and slip on the Main Zagros Thrust), the sum of ~100–150 km is notably smaller than the calculated amount of convergence of ~500–800 km between Arabia and Eurasia since 35–23 Ma (Figure 4)

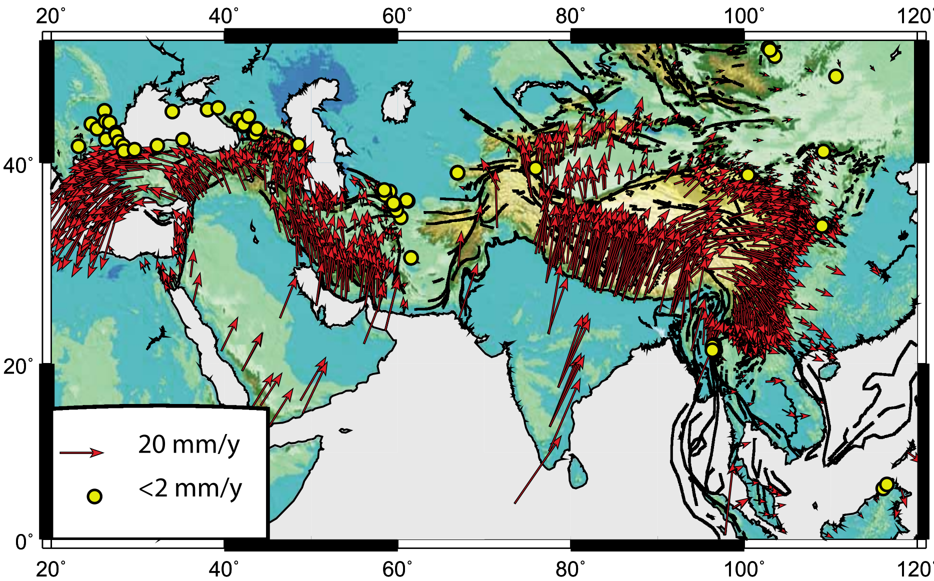


Figure 6. Map of Asia showing topography and GPS velocities in a Eurasian reference frame. Yellow dots show GPS control points with speeds smaller than 2 mm/yr; arrows show velocities of points moving more rapidly than 2 mm/yr with respect to Eurasia. Note the more rapid convergence of India than Arabia with respect to Eurasia, and in both regions much of the convergence is absorbed across a wide region north of the Zagros and Himalaya. Note also the rapid eastward and southward lateral transport of eastern Tibet around northeastern India and rapid westward transport of the region north of northwestern Arabia. The region between the Indian and Arabian plates and east of the Iranian Plateau seems to be part of Eurasia, for speeds of control points there are less than 2 mm/yr. Velocities are taken from *Bettinelli et al.* [2006], *Gan et al.* [2007], *Jade et al.* [2004], *Mohadjer et al.* [2010], *Reilinger et al.* [2006], *Socquet et al.* [2006], *Tavakoli et al.* [2008], *Walpersdorf et al.* [2006], and A. Walpersdorf (personal communication, 2009).

[*McQuarrie et al.*, 2003]. If we take into account the northeast-southwest orientation of the crustal shortening but the nearly north-south orientation of relative plate motion, we might assume that the amount of convergence absorbed by deformation across the Zagros has been yet larger, by roughly 1.4 (= 1/cos 45°) times, but such an assumption requires that a large strike-slip component be absorbed within the Zagros. Some paleomagnetic declination anomalies as large as 20° [*Aubourg et al.*, 2008] do allow for this possibility, but other such measurements show no resolvable declination anomalies [*Homke et al.*, 2004].

[31] In summary, it seems unlikely that deformation in the Zagros can account for as much as half of the convergence between Arabia and Eurasia, and quite possibly, it accounts for only ~20% of that convergence. Both of these amounts, shortening across the Zagros and total convergence between Arabia and Eurasia since collision, are much smaller than the corresponding amounts of the Himalaya and for India with respect to Eurasia. Moreover, estimated amounts of shortening across the Himalaya cannot be constrained well, and balancing the budget of crustal material in both eastern and mideastern Asia continues to pose a challenge.

3.3. Present-Day Kinematics Based on GPS Measurements

[32] The penetrations of Arabia and of India into Eurasia manifest themselves in active deformation over broad regions north and east of the margins of the Arabian and Indian plates (Figure 6). Much of the convergence of Arabia with Eurasia occurs northeast of the Zagros, across the Iranian Plateau and surroundings, and that of India with Eurasia occurs north and northeast of the Himalaya across Tibet and farther north. As discussed below, some of the remaining convergence results in crustal thickening and mountain building, but some is absorbed by lateral transfer of material out of the paths of Arabia and of India with respect to Eurasia (Figure 6). The oblique convergence of Arabia with Eurasia at the northwest trending Zagros induces right-lateral shear across the northwestern Zagros, which in turn transforms into westward translation of crust in northwestern Iran and Turkey [e.g., *McClusky et al.*, 2000; *Reilinger et al.*, 2006]. Similarly, part of the north-northeastward penetration of India into Tibet is absorbed by eastward and southward transfer of material around the eastern end of the Himalaya [e.g., *Gan et al.*, 2007; *Holt et al.*, 1991, 2000; *King et al.*, 1997; *Wang et al.*, 2001; *Zhang et al.*, 2004].

[33] Approximately half of the present-day plate convergence rate between Arabia or India and Eurasia is absorbed at the two mountain belts. Both GPS measurements [e.g., *Bettinelli et al.*, 2006; *Bilham et al.*, 1997; *Feldl and Bilham*, 2006; *Jade et al.*, 2004; *Jouanne et al.*, 2004; *Larson et al.*, 1999] and detailed studies of Quaternary faulting [*Lavé and Avouac*, 2000; *Powers et al.*, 1998] show convergence across the Himalaya of ~18–20 mm/yr, compared to India's

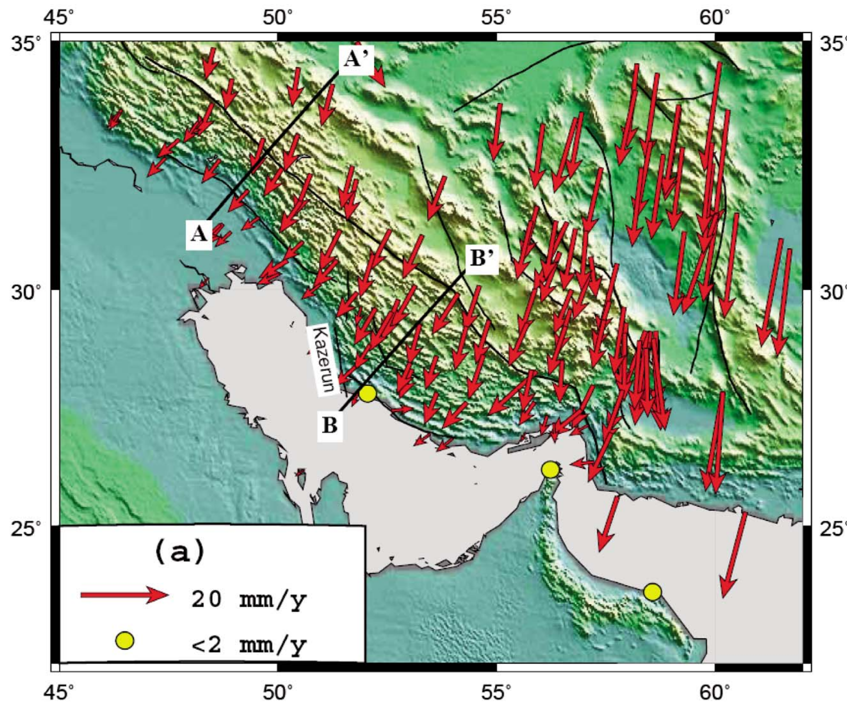
present-day convergence rate with Asia of ~35 mm/yr in the west and 45 mm/yr in the east. Because of the obliquity of convergence in the western Himalaya, the northeast-southwest shortening accounts for a smaller rate of plate convergence than farther east, and much of that convergence is absorbed farther north, with nearly 20 mm/yr across the western Tien Shan alone [*Abdrakhmatov et al.*, 1996; *Reigber et al.*, 2001]. The shortening rate across the Zagros, which decreases from 8 to 10 mm/yr in the segment southeast of the Kazerun line to 4–6 mm/yr northwest of it (Figures 7 and 8) [*Hatzfeld et al.*, 2010; *Hessami et al.*, 2006; *Nilforoushan et al.*, 2003; *Tatar et al.*, 2002; *Vernant et al.*, 2004; *Walpersdorf et al.*, 2006], is less than half of the convergence rates between Arabia and Eurasia of ~18 mm/yr in the northwest and ~25 mm/yr at the Hormuz strait [e.g., *Reilinger et al.*, 2006; *Sella et al.*, 2002; *Vernant et al.*, 2004; *Vigny et al.*, 2006]. Even with allowance for a 45° obliquity of the range to the direction that Arabia moves with respect to Eurasia, oblique convergence across the Zagros cannot account for more than half of the Arabia-Eurasia relative plate motion. To appreciate this, suppose that the fraction of plate convergence accommodated by shortening across the Zagros equals the shortening rate divided by the cosine of the obliquity (cos 45°) and hence is 1.4 times greater than the shortening rate. Then oblique convergence across the Zagros would account for 6–9 mm/yr of the 18 mm/yr of relative plate motion in the northwest and 11–14 mm/yr of 25 mm/yr in the southeast.

[34] In both regions, most of the convergence detected at the Earth's surface occurs at the fronts of the belts, on the southwestern edge of the Zagros and on the southern and southwestern margin of the Himalaya. For the Himalaya, thrust faulting and folding in the sedimentary rock at the foot of the range absorb most of the present-day convergence [e.g., *Baker et al.*, 1988; *Lavé and Avouac*, 2000; *Powers et al.*, 1998], a result consistent with the discovery of surface traces of ruptures associated with huge earthquakes there [*Kumar et al.*, 2001, 2006; *Lavé et al.*, 2005]. For the Zagros a detailed study of Quaternary folding in one region also demonstrates that most of the present-day convergence is absorbed across three or four frontal anticlines at the foot of the Zagros [*Oveisi et al.*, 2007, 2009], though this region, near the southeastern end of the Kazerun line (discussed below), may not be representative of the tectonic style along the Zagros. GPS velocities for Iran also suggest concentrated shortening at the foot of the range (Figures 7 and 8) [e.g., *Hatzfeld et al.*, 2010; *Walpersdorf et al.*, 2006]. GPS data from the Himalaya imply that the main thrust fault is presently locked at shallow depths to at least 15 km and therefore beneath the southernmost 80–100 km of the belt (Figures 7 and 8) [e.g., *Bilham et al.*, 1997; *Feldl and Bilham*, 2006; *Larson et al.*, 1999]. A small amount of permanent strain is absorbed within the Himalaya by slip at ~1–2 mm/yr [*Wobus et al.*, 2005], where topographic profiles steepen abruptly and perhaps above where the underlying thrust fault also steepens [e.g., *Lyon-Caen and Molnar*, 1983; *Molnar*, 1987; *Pandey et al.*, 1995].

[35] The relatively narrow widths, approximately 50–100 km, where strain at the surface is concentrated at the edges of the Zagros and Himalaya, imply that faults that slip at depth are locked at relatively shallow depths of 10–15 km. Thus, most of the convergence at upper crustal depths seems to occur either by slip on a gently dipping thrust fault in the Himalaya or by slip on a nearly flat shear zone (décollement) at the base of (or within) the sedimentary rock of the Zagros. We presume that shear within the middle to lower crust of the Zagros is still sufficiently slow that detecting the resulting northeast-southwest contraction of the upper crust above it with GPS is not yet possible. Convergence of 2–4 mm/yr of the southwestern edge of the Zagros with respect to the Arabian platform, however, suggests that some deformation may occur beneath the Persian Gulf, either aseismically and permanently or as elastic strain accumulation (Figure 7).

[36] The orientations of convergence across the two mountain ranges and the respective styles of deformation differ. For the Himalaya, both slip vectors of earthquakes (discussed in section 3.4) [e.g., *Baranowski et al.*, 1984; *Molnar and Chen*, 1982; *Molnar and Lyon-Caen*, 1989] and GPS velocities of points in the high Himalaya and southern Tibet (Figure 7) [*Jade et al.*, 2004] show radially outward displacement of southern Tibet and Himalaya onto the Indian plate [*Bendick and Bilham*, 2001]. This radially outward convergence with the effectively rigid Indian lithosphere implies divergence of material in southern Tibet, which manifests itself clearly both with grabens oriented orthogonally to the Himalaya [e.g., *Armijo et al.*, 1986] and in fault plane solutions of earthquakes within Tibet (Figure 9, discussed in section 3.4) [e.g., *Molnar and Lyon-Caen*, 1989].

[37] Relative to Arabia, GPS velocities of control points lying in the higher part of the Zagros and the adjacent Iranian Plateau are not perpendicular to the belt as those in the lower part are. Thus, this velocity field includes a strike-slip component along the Zagros (Figures 7 and 8) [e.g., *Hatzfeld et al.*, 2010; *Walpersdorf et al.*, 2006]. In the northwestern Zagros, such shear occurs both as slip on the Main Recent Fault (Figure 8) and by distributed shear across the Zagros [*Authemayou et al.*, 2006; *Sarkarinejad and Azizi*, 2008]. In the southeastern Zagros, the strike-slip shear is smaller but again seems to be distributed across the range. Moreover, paleomagnetic declinations show evidence of rotation consistent with a broad zone of simple shear [*Aubourg et al.*, 2008]. The two segments of the Zagros are separated by a zone of strike-slip shear, the Kazerun line, where not only GPS measurements [e.g., *Tavakoli et al.*, 2008; *Walpersdorf et al.*, 2006] but also disruptions to the overall northwest-southeast trend of folds [e.g., *Hessami et al.*, 2001a; *Lacombe et al.*, 2006; *Mobasher and Babaie*, 2008], fault plane solutions of some earthquakes [*Baker et al.*, 1993; *Talebian and Jackson*, 2004], and offsets of mapped surface traces [*Authemayou et al.*,



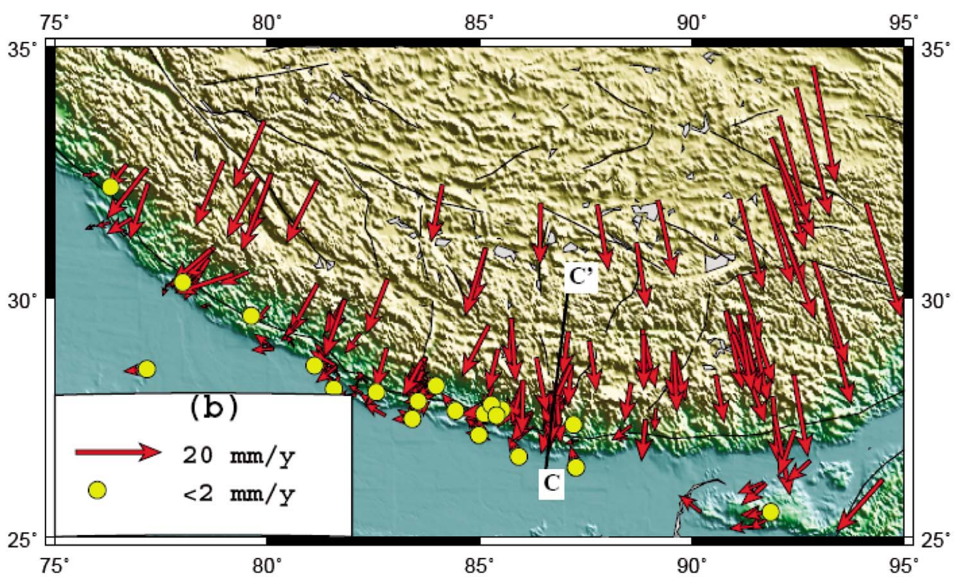
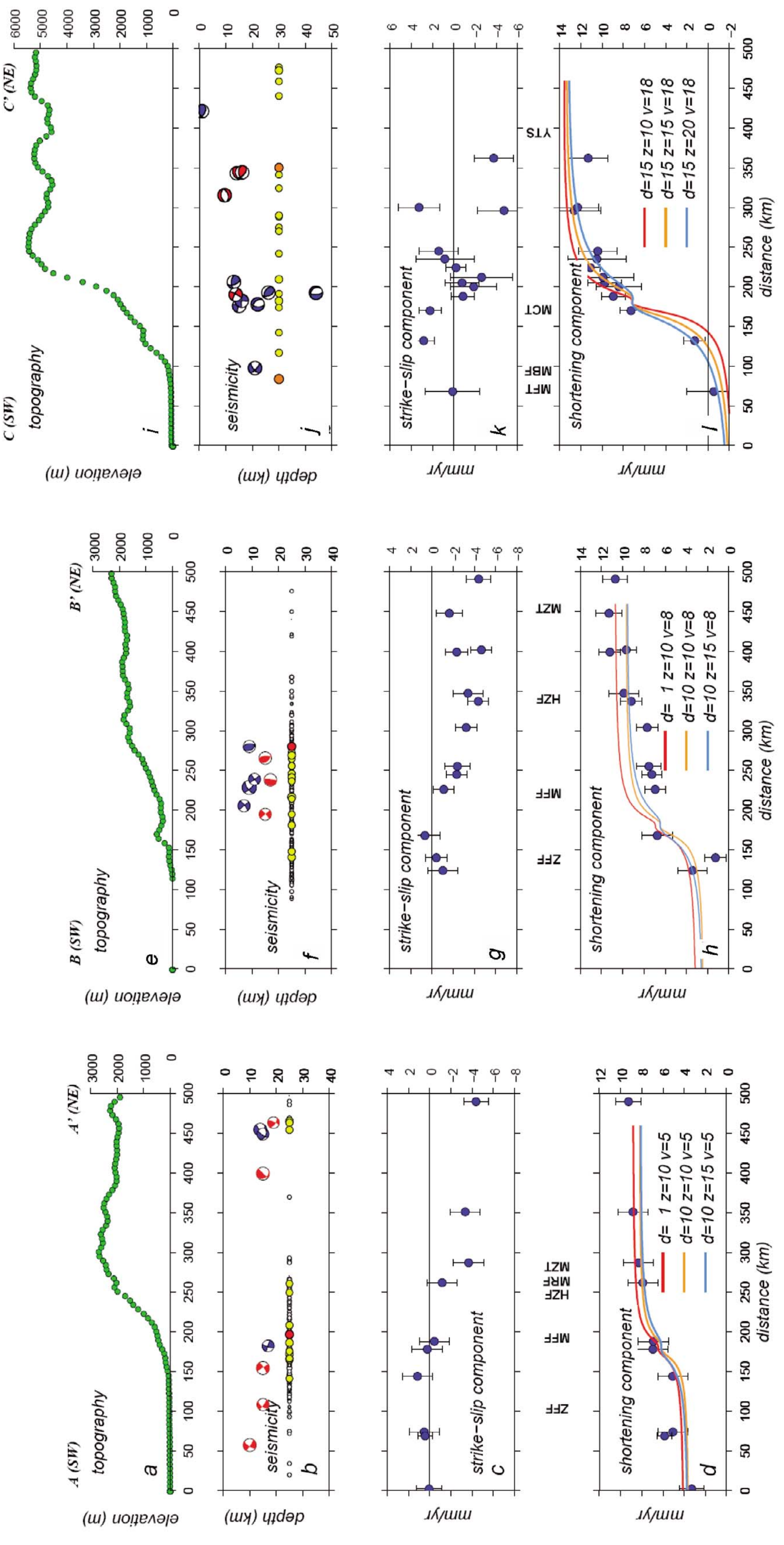


Figure 7. GPS velocities for (a) the Zagros and adjacent areas in a reference frame fixed to Arabia (A. Walpersdorf, personal communication, 2009) and (b) the Himalaya and southern Tibet in a reference frame fixed to India, with velocities taken from *Bettinelli et al.* [2006], *Gan et al.* [2007], and *Jade et al.* [2004]. For Figure 7a, angular velocities of *Vigny et al.* [2006] were used to convert velocities relative to Eurasia or to the ITRF2000 reference frame given in published studies, and for Figure 7b, those of *Bettinelli et al.* [2006] were used. Notice that for control points within the Zagros the component of velocity parallel to the trend of the Zagros is larger in the northwest than in the southeast, but in general, speeds increase from northwest to southeast. By contrast, velocities point radially outward from Tibet, perpendicular to the local trend of the Himalayan arc, and no obvious change in speed occurs from west to east. Black lines labeled A–A′ and B–B′ in Figure 7a and C–C′ in Figure 7b show locations of profiles of GPS velocities and other data in Figure 8.



A (SW)
A' (NE)
B (SW)
B' (NE)
C (SW)
C' (NE)
topography
seismicity
strike-slip component
shortening component
elevation (m)
depth (km)
mm/yr
distance (km)
ZFF
MFF
HZF
MRF
MZT
MFT
MBF
MCT
YTS
d= 1 z=10 v=5
d=10 z=10 v=5
d=10 z=15 v=5
d= 1 z=10 v=8
d=10 z=10 v=8
d=10 z=15 v=8
d=15 z=10 v=18
d=15 z=15 v=18
d=15 z=20 v=18

Figure 8

2009] show strike-slip faulting on several roughly parallel faults.

[38] The Kazerun line has no obvious equivalent in the Himalaya. As noted in section 3.2, the width of the Zagros and its northeast-southwest regional slope northwest of the line differ from those to the southeast. Moreover, although exceptions exist [e.g., *Sherkati et al.*, 2006], most imagine a greater amount of shortening across the Zagros southeast than northwest of the Kazerun line, consistent with the difference in present-day rates. These southeastward increases in rates and in amounts of shortening result in part from the axis of rotation describing Arabia's motion with respect to Eurasia lying closer to the northwestern than southeastern Zagros. As important, however, is the fact that the northerly movement of Arabia relative to central Iran includes a large strike-slip component in the Zagros, which is partly accommodated by slip on the Main Recent Fault north of the Kazerun line but not south of it [e.g., *Authemayou et al.*, 2009; *Tavakoli et al.*, 2008]. Right-lateral slip along Kazerun line then transfers slip on the Main Recent Fault into a convergent component in the southeastern Zagros. The present-day rate of strike-slip shear across the Kazerun line at its northwestern end equals, approximately, the difference in convergence rates across the Zagros northwest and southeast of it.

[39] In detail, the variation in the oblique component of convergence across the Zagros reflects complexity absent in the Himalaya. First, the orientation of Arabia's plate motion with respect to Eurasia, shown by velocities of GPS control points on Arabia (Figure 6), varies from north-northeast at the southeastern end of the Zagros where rates are highest (~25 mm/yr) to due north in the central Zagros to slightly north-northwest in the northwestern Zagros, where rates are lowest (~18 mm/yr). Because the trend of the Zagros does not change along strike, shortening across the northwestern part contributes less to that overall convergence than does convergence across the southeastern part. We might expect the same to apply to the western Himalaya, but in that region little evidence suggests that convergence is slower than farther east [e.g., *Jade et al.*, 2004], despite more rapid India-Eurasia relative plate motion at the longitude of the eastern Himalaya than at the longitude of the western Himalaya. Both the amount and the distribution of shear parallel to the Zagros also vary along the belt (Figures 7 and 8). In the northwest, shear is partitioned into pure right-lateral strike slip along the Main Recent Fault, and largely, but not completely, convergent movement across that segment of the Zagros. The Main Recent Fault follows closely the suture zone between Arabia and Eurasia, but right-lateral slip on it ends where it connects to the Kazerun line [e.g., *Authemayou et al.*, 2009]. Right-lateral slip along fault segments in the Kazerun line then transforms the slip on the Main Recent Fault to oblique convergence across the Zagros to the southeast. Slip on these segments in the Kazerun line decreases toward the south. The region of localized crustal shortening studied by *Oveisi et al.* [2007, 2009] lies just east of where the Kazerun line projects to the southern margin of the Zagros fold belt.

[40] In summary, present-day GPS velocities across the Zagros, with their large oblique convergence and varying rates along the belt, differ markedly from those across the Himalaya. Convergence across the Himalaya, between

Figure 8. Cross sections across the Zagros and Himalaya summarizing topography, seismicity and fault plane solutions, and both horizontal components of GPS velocities relative to (a–h) Arabia and (i–l) India. Figures 8a–8h show profiles A–A′ and B–B′ across the Zagros in Figure 7a, and Figures 8i–8l show profile C–C′ across the Himalaya in Figure 7b. In Figures 8a, 8e, and 8i showing profiles of topography, note the much steeper topographic front for the northwestern profile A–A′ than for the southeastern profile B–B′ across the Zagros. In Figures 8b, 8f, and 8j, we show background seismicity and back hemisphere stereographic projections of focal spheres for earthquakes with reliable fault plane solutions and focal depths. Background seismicity is plotted at a constant depth because errors in focal depths could give misleading impressions of the distribution of activity; red circles show events with $M \geq 6$, yellow circles are for $M \geq 5$, and white circles are for $M < 5$. Seismicity beneath the Zagros (A–A′ and B–B′) is concentrated in the crust beneath the lower, southwestern half of the belt. In the Himalaya (C–C′), it lies beneath the segment where the range steepens. Fault plane solutions of moderate earthquakes show steep dips (30°–60°) of nodal planes beneath the Zagros (see also Figure 9a). Blue beach balls are from a compilation by J. A. Jackson (personal communication, 2009), and red ones are from the Harvard Centroid Moment Tensor (CMT) catalog (http://www.seismology.harvard.edu/CMTsearch.html). For the Himalaya, for moderate earthquakes (blue beachballs, CMT and studies cited in section 3.3) at depths of ~15 km beneath the Himalaya, fault planes dip gently (~15°) northward beneath the Himalaya. At greater depth and beneath Tibet, solutions of both moderate earthquakes and microearthquakes (red beach balls [*de la Torre et al.*, 2007]) show reverse and normal faulting (see Figures 9b and 9c). In Figures 8c, 8g, and 8k, the components of velocity parallel to the ranges (strike-slip components) are larger on profile A–A′ across the Zagros than on the southwestern profile B–B′ and profile C–C across the Himalaya. Positive values correspond to right-lateral shear. In Figures 8d, 8h, and 8l, convergent components are greater on profile B–B′ than on A–A′ across the Zagros and greatest across the Himalaya (profile C–C′). In all cases, present-day strain accumulation is concentrated near the foot of each belt. In Figures 8d, 8h, and 8l, where velocities perpendicular to the belts are plotted, we also show velocities calculated assuming slip on gently dipping thrust faults beneath the belts but locked at depths shallower than 10 or 15 km; d gives the dip in degrees of the fault, z is the locking depth in km, and v is the slip rate in mm/yr on the portion free to slip. In A–A′ and B–B′, ZFF marks the trace of the Zagros Foredeep Fault, MFF marks the Mountain Front Fault, HZF marks the High Zagros Fault, MRF marks the Main Recent Fault, and MZT marks the Main Zagros Thrust (Figure 5) [*Berberian*, 1995]. In C–C′, MFT marks the Main Frontal Thrust, MBF marks the Main Boundary Fault, MCT marks the Main Central Thrust, and YTS marks the Yarlung Tsangpo Suture (Figure 5).

southern Tibet and the Indian shield, occurs at a nearly identical rate along the arc and is oriented normal to it.

3.4. Present-Day Kinematics of Deformation From Fault Plane Solutions of Earthquakes

[41] Fault plane solutions of earthquakes also reveal different patterns in the Himalaya and Zagros (Figure 9) that may reflect different stages of development of the two belts. Alternatively, they might result from differences in material properties.

[42] Fault plane solutions of earthquakes in the Himalaya resemble those at subduction zones. Solutions for occasional events beneath the Ganga Basin show normal faulting due to the stretching of the upper part of the lithosphere as it bends down to form the basin, analogous to those for earthquakes beneath deep-sea trenches (Figures 7 and 9) [*Isacks et al.*, 1968; *Stauder*, 1968]. Most solutions (Figures 8 and 9) show thrust slip, perpendicular to the belt, on gently dipping planes (dips <25° and commonly <15°). These earthquakes reflect the underthrusting of intact Indian lith-

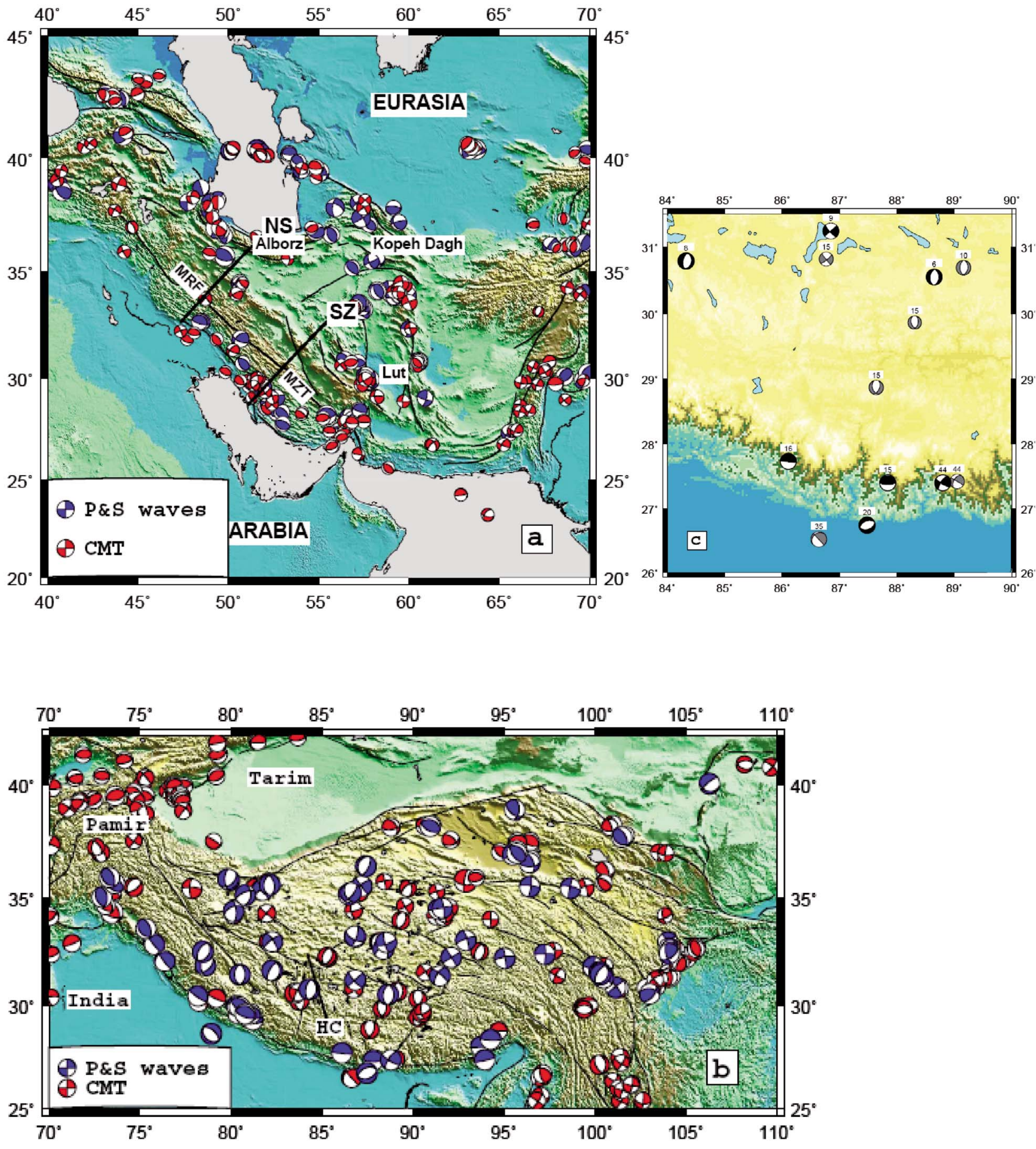


Figure 9

osphere beneath the Lesser Himalaya, if not the entire belt [e.g., *Baranowski et al.*, 1984; *Molnar and Chen*, 1982; *Molnar and Lyon-Caen*, 1989; *Ni and Barazangi*, 1984]. It appears that the underthrusting of India beneath the Himalaya occurs mostly by slip in very large earthquakes, not just those with magnitudes of ~8, like those that have occurred in the twentieth century [e.g., *Bilham et al.*, 2001; *Seeber and Armbruster*, 1981], but more importantly in much larger events with recurrence intervals of several hundred to a thousand years [e.g., *Kumar et al.*, 2001, 2006; *Lavé et al.*, 2005]. In addition, a few earthquakes deep in the crust or upper mantle show reverse faulting at depth within the Indian plate [*de la Torre et al.*, 2007; *Ni and Barazangi*, 1984]; they may reflect compression of the bottom of the plate due to its flexure, but because other events that occur within the Indian plate also show reverse faulting [e.g., *Gahalaut et al.*, 2004; *Saikia*, 2006], those beneath the Himalaya may simply result from large horizontal compressive stresses perpendicular to the range.

[43] We are not aware of important seismicity in Arabia southwest of the folds in the Zagros or fault plane solutions of earthquakes near that region that show normal faulting and stretching of the top surface of a flexed Arabian plate (Figure 9). Thus, either flexure is modest, or the horizontal compressive stress within the Arabian lithosphere suffices to nullify the extensional deviatoric stress that develops on the top of a flexed plate. Nearly all earthquakes within the Zagros show reverse faulting, with planes dipping at 30°–60° and with *P* axes oriented roughly perpendicular to the belt (Figures 8 and 9) [e.g., *Jackson and McKenzie*, 1984; *Maggi et al.*, 2000; *McKenzie*, 1972a; *Talebian and Jackson*, 2004]. A few solutions also show strike-slip faulting associated with segments of the Kazerun fault system. Of particular importance is the fact that most of these earthquakes occurred in the basement beneath the thick (~10 km) sedimentary cover of the Zagros [*Hatzfeld et al.*, 2010; *Jackson and Fitch*, 1981; *Maggi et al.*, 2000; *Ni and Barazangi*, 1986; *Talebian and Jackson*, 2004; *Tatar et al.*, 2004]. Although controversy persists, most imagine that the faulting in the basement occurs independently of the folding of the sedimentary cover because the cover is detached from the basement along weak layers of evaporites [e.g., *Falcon*, 1974; *Jahani et al.*, 2009; *Talbot and Alavi*, 1996]. *Walker et al.* [2005b] showed one example of surface faulting associated with an earthquake in the Zagros, but they emphasized that this was a rare occurrence. Thus, most moderate earthquakes reflect straining of the subducting Arabian plate, not the underthrusting of that plate beneath the mountain belt. Yet that seismic strain rate is low; summation of seismic moment tensors of these earthquakes cannot account for the rate of shortening measured with GPS [e.g., *Jackson and McKenzie*, 1988; *Masson et al.*, 2005; *North*, 1974].

[44] Fault plane solutions of earthquakes within Tibet show a mixture of normal and strike-slip faulting with both requiring east-west extension [e.g., *Langin et al.*, 2003; *Molnar and Chen*, 1983; *Molnar and Lyon-Caen*, 1989; *Molnar and Tapponnier*, 1978; *Ni and York*, 1978; J. R. Elliott et al., Extension on the Tibetan Plateau: Recent normal faulting measured by InSAR and body-wave seismology, submitted to *Geophysical Journal International*, 2010]. A summation of moment tensors of earthquakes within Tibet suggests that east-west extension occurs at roughly twice the rate of north-south compression [e.g., *Molnar and Chen*, 1983; *Molnar and Lyon-Caen*, 1989]. Normal faulting leads to vertical compression (crustal thinning), and the conjugate strike-slip faulting, with left-lateral slip on northeast striking planes and right-lateral on northwest striking planes, contributes north-south shortening. We are aware of no reliable solution that shows thrust or reverse faulting within the plateau where it is higher than ~4000 m [*Molnar et al.*, 1993; Elliott et al., submitted manuscript, 2010]. Thrust and reverse faulting, however, do occur north and northeast of Tibet: in the Tien Shan [e.g., *Maggi et al.*, 2000; *Nelson et al.*, 1987; *Tapponnier and Molnar*, 1979], on the northeast margin of Tibet in the Qilian Shan [e.g., *Molnar and Lyon-Caen*, 1989], and farther northeast in the Altay and Gobi-Altay [e.g., *Bayasgalan et al.*, 2005].

[45] By contrast, no normal faulting occurs within the Iranian Plateau (Figure 9). Instead, fault plane solutions show largely strike-slip and some reverse faulting [e.g.,

Figure 9. Plots of lower hemisphere diagrams of fault plane solutions of shallow focus earthquakes in (a) the Zagros and Iranian Plateau, (b) the Himalaya and Tibet, and (c) a small portion of the Himalaya and Tibet that illustrates well the contrast in fault plane solutions of earthquakes in these regions. Beach balls show lower hemisphere stereographic projections of the focal sphere, with dark quadrants showing regions of extensional strain, white quadrants showing regions of compressional strain, and boundaries between light and dark quadrants giving the orientations of the two nodal planes. One nodal plane ruptured during the earthquake, and the normal to the other defines the orientation of slip on the fault plane. For the Zagros (Figure 9a), most solutions show reverse faulting, horizontal compressive strain, and nodal planes that dip 30°–60°. Across the Iranian Plateau, most solutions show strike-slip faulting, but some show reverse faulting. For the Himalaya (Figures 9b and 9c), a couple of solutions show normal faulting beneath the Ganga Basin, south of the range, but most earthquakes within the Himalaya show one nodal plane dipping gently north and thrust slip on it. For Tibet, however, a mixture of normal and strike-slip faulting occurs. Thick black lines labeled NS and SZ in Figure 9a and HC in Figure 9b show locations of profiles in Figure 11. CMT solutions are from http://www.seismology.harvard.edu/CMTsearch.html. Solutions based on *P* and *S* waves are from *Molnar and Lyon-Caen* [1989] for Tibet and the Himalaya and from a compilation by J. A. Jackson (personal communication, 2009) for Iran. In Figure 9c, black beach balls shows solutions from the CMT catalog or from *Molnar and Lyon-Caen* [1989], and those in gray are from *de la Torre et al.* [2007]. Numbers next to each give focal depths in kilometers.

Berberian and Yeats, 1999; *Hatzfeld et al.*, 2010; *Jackson and McKenzie*, 1984, 1988; *Jackson et al.*, 1992; *Priestley et al.*, 1994; *Talebian and Jackson*, 2004]. Active faulting and surface ruptures of most major earthquakes in Iran northeast of the Zagros show strike-slip faulting [e.g., *Ambraseys*, 1963; *Ambraseys and Jackson*, 1998; *Ambraseys and Melville*, 1982; *Berberian et al.*, 1984, 1992, 1999, 2001; *Binet and Bollinger*, 2005; *Fattahi et al.*, 2007; *Funning et al.*, 2005; *Hessami et al.*, 2003; *Hollingsworth et al.*, 2006, 2007; *Jackson et al.*, 2006; *Landgraf et al.*, 2009; *Le Dortz et al.*, 2009; *Meyer and Le Dortz*, 2007; *Meyer et al.*, 2006; *Nazari et al.*, 2009a, 2009b; *Peyret et al.*, 2008; *Ritz et al.*, 2006; *Shabanian et al.*, 2009a, 2009b; *Talebian and Jackson*, 2002; *Tirrul et al.*, 1983; *Walker and Jackson*, 2002, 2004; *Walker et al.*, 2005a, 2009], but some require reverse faulting [e.g., *Ambraseys and Jackson*, 1998; *Berberian et al.*, 2000; *Fattahi and Walker*, 2007; *Fattahi et al.*, 2006; *Fielding et al.*, 2004; *Parsons et al.*, 2006; *Talebian et al.*, 2006; *Tatar et al.*, 2007; *Walker et al.*, 2003, 2004]. Similarly, microearthquake studies demonstrate reverse slip in several regions [e.g., *Tatar and Hatzfeld*, 2009; *Yamini-Fard et al.*, 2006, 2007]. Thus, deformation of the Iranian Plateau differs markedly from that of Tibet.

[46] As noted in section 3.3, in both the Himalaya and the Zagros, current surface deformation seems to be localized at the fronts of the belts, where gently dipping thrust faults or surfaces of deécollement approach or reach the surface. Two major differences between the Himalaya and Zagros, however, stand out.

[47] First, as discussed in section 3.3, shortening across the Himalaya is oriented normal to the belt, and the rate is nearly constant along it. In the Zagros, a large component of oblique convergence occurs, and both the rate and orientation across the belt vary along it.

[48] Second, whereas the construction of the present-day Himalaya owes its existence to large magnitudes of thrust slip on major faults that formed within Indian lithosphere, no obvious equivalent seems to have developed within the Arabian lithosphere. *Agard et al.* [2005] suggested that slip continues on segments of the Main Zagros Thrust, albeit at only 3–4 mm/yr since 15–20 Ma. Yet GPS measurements give no indication of present-day convergence, and we suspect that slip on this thrust fault has ceased. From the tens of kilometers of shortening associated with folding of sedimentary rock detached from the underlying strata and basement, we, like most workers, presume that much of the convergence between Arabia and the southwest margin of the Iranian Plateau has accumulated by slip on a flat fault, or surface of décollement, beneath the thick, folded sedimentary cover of the Zagros and above the basement. Following a more recent view [e.g., *Berberian*, 1995; *Molinaro et al.*, 2005; *Mouthereau et al.*, 2006, 2007; *Sherkati et al.*, 2005], however, we also imagine that that style of deformation is giving way to an increasingly important role for deformation within the underlying basement.

[49] In some sense the Himalayan equivalent to the wide belt of folded sedimentary rock that dominates the Zagros landscape lies north and northeast of the belt of high Himalayan peaks in what is often termed the Tethyan Himalaya, but this folding is no longer active [e.g., *Corfield et al.*, 2005; *Ratschbacher et al.*, 1994; *Wiesmayr and Grasemann*, 2002]. The zone of folded sedimentary rock deposited on the Indian shelf is now less than 100 km wide, much narrower than the Zagros, but it may have been comparably wide before erosion removed much of it, leaving only metamorphic rock to crop out today in the Greater Himalaya and farther south. As we discuss further below, the equivalents of the Main Central Thrust and Main Boundary Fault (which separates the largely Mesozoic and older rock of the Himalaya from the late Cenozoic sedimentary rock of the Ganga Basin) (Figure 5) do not seem to have formed yet in the Zagros.

[50] Since *Jackson* [1980] suggested that reverse faulting may occur by reactivation of normal faults, such as those forming at continental margins, many now imagine that the deformation in the basement beneath the Zagros reflects such reactivation [e.g., *Mouthereau et al.*, 2007; *Ni and Barazangi*, 1986]. Presumably, the same process occurred in the Himalaya before the Main Central Thrust became the locus of convergence between India and southern Tibet. The rupturing of a major fault like the Main Central Thrust, however, might not occur instantaneously in geologic time. Instead, reverse faulting within the northern Indian continent might have been distributed over a wide area until one such fault could coalesce with a presumably ductile shear zone in the lower crust, which then became the Main Central Thrust. Finding evidence for such reverse faulting in the highly deformed rock of the Greater Himalaya must be hard, but *Steck* [2003] and his colleagues [*Robyr et al.*, 2002; *Steck et al.*, 1993a, 1993b, 1999; *Vannay and Steck*, 1995; *Wyss et al.*, 1999] showed that early in the history of deformation of the high Himalayan crystalline rock, thrust or reverse faulting took place on faults that dip south. *Schlup et al.* [2003] showed that such deformation must have occurred early in the collision history, before 40 Ma. *Brown and Nazarchuk* [1993] inferred similar structures 1000 km east [see also *Godin et al.*, 1999a, 1999b], and *Godin et al.* [2001] showed this deformation to have occurred before 35 Ma. It is easy to imagine that most evidence of such reverse faulting in the brittle upper crust has been destroyed, particularly that of faults that dipped northward, if they existed. Moreover, the faults recognized by Steck, Brown, Godin, and colleagues might have formed initially as normal faults, though we are not aware of evidence suggesting this. In any case, it follows that we may be living in a time of transition during which the upper 10–15 km of the Zagros crust shortens by distributed strain, before a major thrust fault can slice through the entire crust, as occurred when the Main Central Thrust formed in the Himalaya.

4. DEEP STRUCTURE

4.1. Crustal and Upper Mantle Structure of the Zagros and Himalaya

[51] The two belts share the similarity that lithospheric plates have been flexed down and have underthrust the

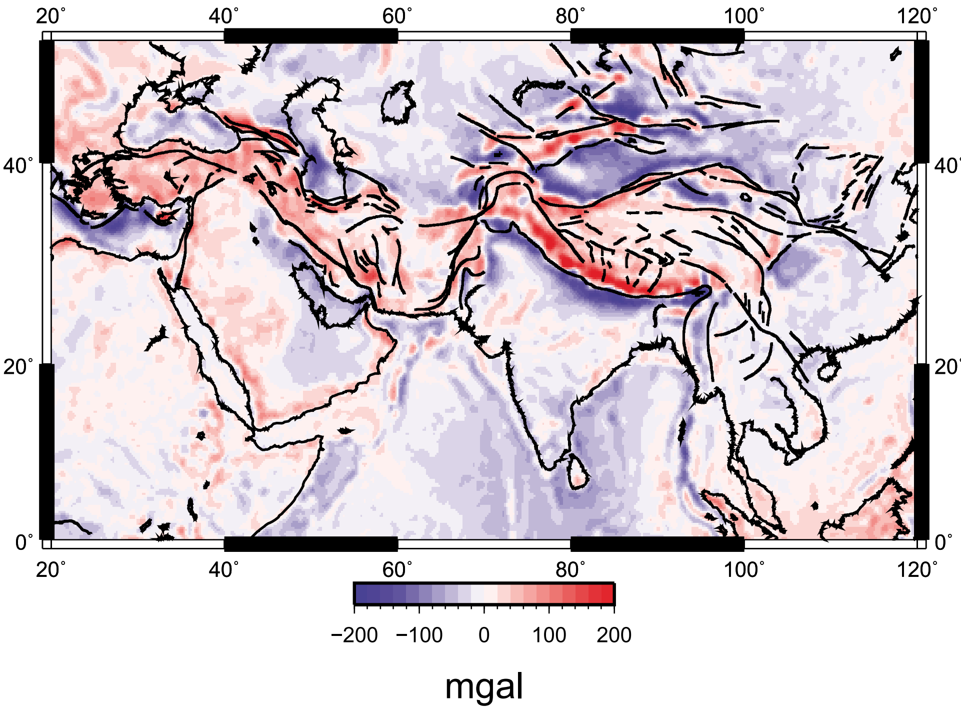


Figure 10. Map of Asia showing free-air gravity anomalies of the Gravity Recovery and Climate Experiment (GRACE) gravity model GGM02C [*Tapley et al.*, 2005]. Scale in mgal at the bottom. Note the small values over the stable regions; the belts of negative anomalies adjacent to the Zagros, the Himalaya, and other mountain ranges; the belts of positive anomalies over the Zagros, the Himalaya, and other ranges; and the small values over the Tibetan and Iranian plateaus.

southern margin of Eurasia, but the deeper structures of the belts differ in important ways. It appears that these differences arise largely because of the greater amounts of convergence in the Himalaya than in the Zagros or from an alternative point of view, the earlier stage in the development of the Zagros than that of the Himalaya [e.g., *Barazangi*, 1989; *Ni and Barazangi*, 1986]. Studies of the deep structure beneath the Zagros and Himalaya contribute to and confirm the geological cross sections (Figure 5) discussed in section 3.2.

[52] Gravity anomalies across the Zagros and Himalaya demonstrate large deviations from isostasy in some areas, which require lithospheric strength for their support (Figure 10). For both regions, Bouguer anomalies become progressively negative as mean elevations increase, as they must in a state approximately in isostatic equilibrium. Profiles of Bouguer anomalies across both belts, and especially across the Himalaya, have proven to be useful for constraining deeper structure, but such constraints are demonstrated more clearly by isostatic anomalies and by free-air anomalies. For both regions, free-air anomalies, based on satellite measurements and terrain-corrected terrestrial measurements (Figure 10), define belts of negative anomalies over the basins adjacent to the mountain ranges, positive anomalies over the ranges themselves, and negligibly small anomalies over the adjacent plateau. The paired belts of negative and positive free-air anomalies result in part from the "edge effect" associated with the juxtaposition of crust and mantle with different crustal thicknesses in isostatic equilibrium, but these anomalies also indicate mass deficits and excesses. The near-zero free-air anomalies over the two plateaus attest to isostatic equilibrium, but they provide no constraint on the degree to which that equilibrium derives from thick crust or low-density mantle.

[53] The belts of negative free-air anomalies beneath both the sedimentary basin southwest of the Zagros, which includes the Persian Gulf, and the Ganga Basin south and southeast of the Himalaya result in part from mass deficits. As calculations that match observed anomalies show, they attest to a gentle dip of the Moho beneath these areas, where lithosphere has been flexed down by the weight of the Zagros thrust atop the Arabian plate [*Snyder and Barazangi*, 1986] and of the Himalaya atop the Indian plate [e.g., *Cattin et al.*, 2001; *Duroy et al.*, 1989; *Jin et al.*, 1996; *Karner and Watts*, 1983; *Lyon-Caen and Molnar*, 1983, 1985; *McKenzie and Fairhead*, 1997; *Tiwari et al.*, 2006; *Warsi and Molnar*, 1977]. The much more negative anomaly beneath the Ganga Basin than beneath the Persian Gulf accords with greater lithospheric flexure.

[54] The positive free-air anomalies over the Zagros and the Himalaya require an excess of mass, which in both cases is supported by the strong lithosphere beneath the belts. *Snyder and Barazangi* [1986] had used the gradient in Bouguer anomalies across the Zagros to infer marked crustal thickening from southwest to northeast, but receiver functions (Figure 11) indicate a deepening of the Moho of, at most, a few kilometers [*Paul et al.*, 2006, 2010]. *Paul et al.* [2006] showed that the negative Bouguer and relatively

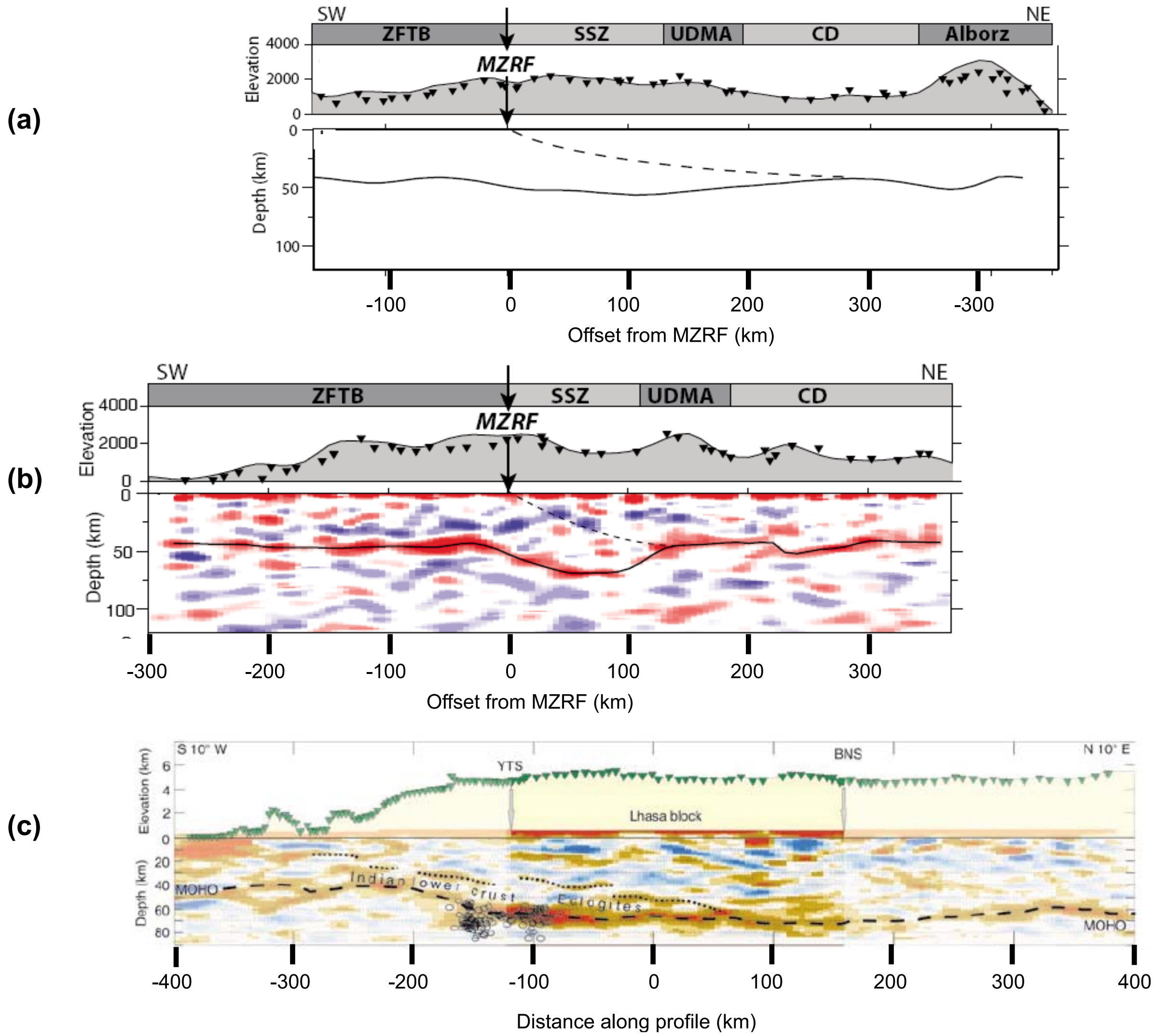


Figure 11. Cross sections (based on receiver functions) for (a and b) the Zagros and (c) the Himalaya aligned at the suture zones (MZRF in Figures 11a and 11b and YTS in Figure 11c). In Figure 11a, we show an interpretation of a migrated receiver function presented by *Paul et al.* [2010], Figure 11b is from *Paul et al.* [2006], and Figure 11c is from *Wittlinger et al.* [2009]. For locations of profiles, see Figure 9a for Figures 11a and 11b and Figure 9b for Figure 11c. In Figures 11a and 11b, distance is measured with respect to the Main Recent Fault. The average elevation profile is shown on the top of each, with the locations of seismograph stations (black triangles) projected onto the profile. In Figure 11b, we show a common conversion point, depth-migrated cross section, and in Figure 11c, we show common conversion point, depth-migrated cross sections for *S* waves converted to *P* waves. In Figures 11b and 11c, red and brown show depths of interfaces at which the impedance above the interface is smaller than that below and hence where, in general, low-speed material overlies higher-speed material. Blue shows the opposite, interfaces where, in general, high-speed material overlies lower-speed material. Solid black lines in Figures 11a and 11b and the dashed black line in Figure 11c show the Moho. The dashed lines in Figures 11a and 11b show the crustal-scale thrust fault inferred by *Paul et al.* [2006, 2010], based also on gravity data for Figure 11b. For Figures 11a–11c, time has been converted to depth using wave speeds inferred from the migrated seismograms. Open circles at horizontal distances between −100 and −150 km and depths between 60 and 80 km show the earthquake clusters located by *Monsalve et al.* [2006] farther east and projected onto the profile.

small positive free-air anomaly seem to be due to a thickening toward the northeast of sedimentary rock and low-density upper crust.

[55] The greater positive free-air anomalies over the Himalaya than the Zagros (Figure 10) require a substantially greater excess mass. Moreover, although not easily seen from the map in Figure 10, calculations of gravity anomalies for profiles across the Himalaya indicate that insofar as the principal density contrasts are those at the Earth's surface and at the Moho, the Moho must dip more steeply, by 10°–15°, beneath the Greater Himalaya than farther south [e.g., *Warsi and Molnar*, 1977]. Accordingly,

the flexural rigidity of the lithosphere beneath the Greater Himalaya must be smaller than it is farther south [e.g., *Cattin et al.*, 2001; *Duroy et al.*, 1989; *Jin et al.*, 1996; *Karner and Watts*, 1983; *Lyon-Caen and Molnar*, 1983, 1985; *Tiwari et al.*, 2006]. Images of crustal thickness, using both seismic reflections from controlled sources [e.g., *Hauck et al.*, 1998; *Hirn et al.*, 1984a; *Zhao et al.*, 1993] and receiver functions [*Mitra et al.*, 2005; *Nábělek et al.*, 2009; *Schulte-Pelkum et al.*, 2005; *Wittlinger et al.*, 2009], confirm a steepening of the Moho beneath the Himalaya (Figure 11). As calculations of gravity anomalies show, support for the large free-air anomalies over the Ganga Basin and Himalaya apparently derives from the strength of the Indian plate, which has been flexed down to form the basin but is strong enough to support part of the weight of the Himalaya, if with a markedly smaller flexural rigidity beneath the Greater than Lesser Himalaya [*Burov and Watts*, 2006; *Cattin et al.*, 2001; *Duroy et al.*, 1989; *Hetényi et al.*, 2006; *Jin et al.*, 1996; *Karner and Watts*, 1983; *Lyon-Caen and Molnar*, 1983, 1985; *Tiwari et al.*, 2006].

[56] Two profiles of receiver functions across the Zagros differ somewhat from one another but show some thickening of crust northeast of the Main Zagros Thrust and indications of a northeastward continuation of that fault to depth (Figure 11) [*Paul et al.*, 2006, 2010]. The northwest profile, in particular, displays an obvious dipping zone where *P* waves are converted to *S* waves and where the underlying material has the lower *S* wave speed [*Paul et al.*, 2010]. When projected to the surface, this interface approaches the surface trace of the Main Zagros Thrust, suggesting that it marks the continuation of that fault at depth. It dips gently, at ~15°, and can be traced ~250 km farther northeast to the Moho beneath central Iran. The approximate thickness of the low-speed material of 5–8 km is consistent with its being sedimentary rock that has been underthrust beneath crystalline basement of central Iran. A similar, but weaker and less continuous, interface can be seen on the southeast profile (Figure 11), and it too might mark the continuation of the Main Zagros Thrust.

[57] Northeast of the surface trace of that fault, crust thickens from 43 ± 2 km beneath the Zagros fold-and-thrust belt to 69 ± 2 km along the southeast profile but to only 56 ± 2 km beneath the northwest profile (Figure 11) [*Paul et al.*, 2006, 2010]. The thickened crust for the southeastern profile accounts for the negative Bouguer gravity anomalies along this segment of the gravity profile [*Paul et al.*, 2006]. Thus, the extent of low-density crust and the projection of the Main Zagros Thrust to the Moho yield estimates of the extent to which the Arabian platform has underthrust the southwestern edge of the Iranian Plateau. *Paul et al.* [2010] estimated 59 km of such shortening for the southern profile and 49 km for the northern profile, values that accord with the geologic estimates of 50–70 km by *Agard et al.* [2005] and *Mouthereau et al.* [2007].

[58] The dipping surface that *Paul et al.* [2006, 2010] imaged using receiver functions (Figure 11) resembles a surface beneath the Himalaya that has been imaged both with reflections from controlled seismic sources [e.g., *Hauck et al.*, 1998; *Zhao et al.*, 1993] and with receiver functions, *S* (or *P*) waves converted from *P* (or *S*) waves at it [e.g., *Mitra et al.*, 2005; *Nábělek et al.*, 2009; *Schulte-Pelkum et al.*, 2005; *Wittlinger et al.*, 2009]. They differ, however, in that the surface beneath the Zagros seems to mark the suture zone between Arabia and Eurasia, which we infer no longer to be active. By contrast, the surface beneath the Himalaya marks the active thrust fault along which apparently intact Indian crust slides beneath slivers of Indian crust detached from that below it and beneath southernmost Tibet.

4.2. Crustal Thickness and Upper Mantle Structure of the Iranian and Tibetan Plateaus

[59] The two plateaus differ in crustal thickness. The Iranian Plateau, which stands roughly 1000–1500 m high, is underlain by crust whose thickness seems to differ little from that of adjacent regions (Figure 12). Beneath the plateau crustal thicknesses are reported as ~40–50 km, and beneath surrounding lowlands they seem to be closer to 35–40 km. Beneath Tibet, which stands 4500–5500 m high, crustal thicknesses are everywhere greater than 60 km, typically 70 km, and in places possibly as large as 80 km (Figure 12). Thus, they exceed crustal thicknesses of surroundings (35–40 km) from possibly as little as 20 km to as much as 45 km.

[60] Studies using receiver functions for Iran show variability in crustal thickness, but except near the Main Zagros Thrust [e.g., *Paul et al.*, 2006, 2010; *Yamini-Fard et al.*, 2006], most estimates are between 40 and 50 km, compared with only ~35 km for the Arabian platform immediately adjacent to the Zagros [e.g., *Gök et al.*, 2008; *Pasyanos et al.*, 2007] and for much of the Zagros, where crustal thicknesses also include ~35 km of basement with ~10 km of sedimentary rock above [e.g., *Hatzfeld et al.*, 2003; *Paul et al.*, 2006, 2010]. Northeast of the thickest crust and, hence, beneath the southwest edge of the Iranian Plateau, *Paul et al.* [2006] measured a crustal thickness of ~42 km using receiver functions. Elsewhere, again using receiver functions, *Doloei and Roberts* [2003] obtained 46 ± 2 km for the region near Tehran, just south of the Alborz, if beneath the Alborz values reach ~51–54 km [*Sodoudi et al.*, 2009]. Farther east, *Nowrouzi et al.* [2007] inferred crustal thickness of 44–50 km beneath the Iranian portion of the Kopet Dagh. Thus, although crustal thickening has occurred in the mountain belts that surround the Iranian Plateau, there is little evidence to suggest that the crust beneath the interior of the plateau has thickened much, no more than 10 km and perhaps none at all in some areas.

[61] For Tibet, data of many types call for crustal thicknesses in excess of 60 km: seismic reflection profiling [e.g., *Galvé et al.*, 2002a, 2006; *Hirn et al.*, 1984a, 1984b; *Ross et al.*, 2004; *Zhao et al.*, 2001], wide-angle reflections of converted phases [*Tseng et al.*, 2009], receiver functions [e.g., *Chen et al.*, 2005; *Hetényi et al.*, 2006; *Kind et al.*, 2002; *Owens and Zandt*, 1997; *Schulte-Pelkum et al.*, 2005; *Shi et al.*, 2004; *Wittlinger et al.*, 2004, 2009; *Xu et al.*, 2007; *Yuan et al.*, 1997], and surface wave dispersion [e.g., *Chen and Molnar*, 1981; *Cotte et al.*, 1999; *Holt*

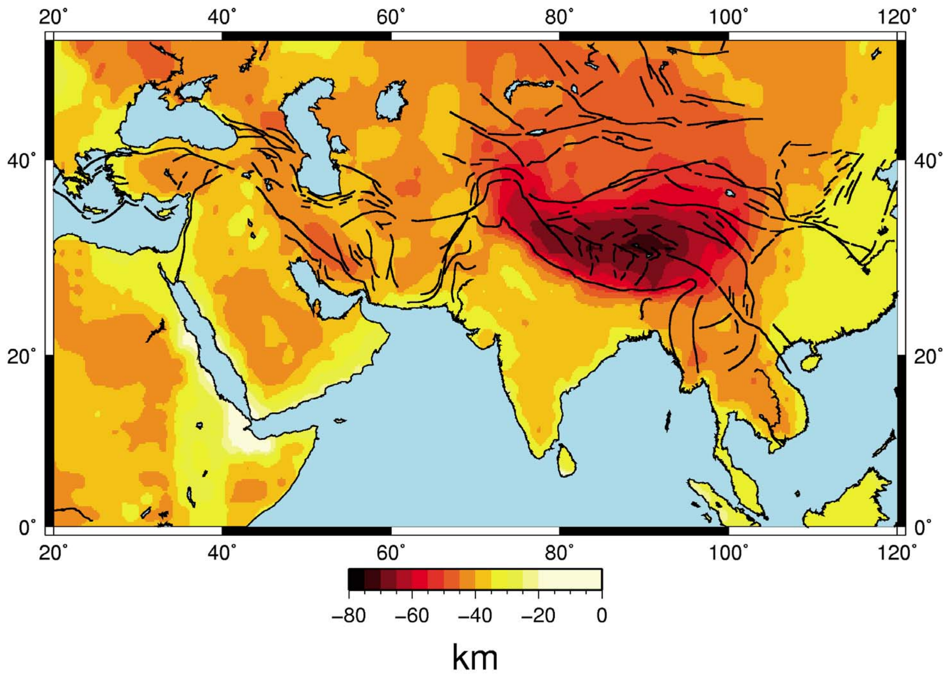


Figure 12. Map of Asia showing crustal thickness smoothed from the 2° × 2° compilation, CRUST 2.0, of G. Laske, G. Masters, and C. Reif [*Bassin et al.*, 2000; http://mahi.ucsd.edu/gabi/rem.html]. Note the much thicker crust beneath Tibet than beneath Iran. Depths are only approximate; the gradient across Tibet shown well by *Tseng et al.* [2009], for instance, is not resolved here.

and Wallace, 1990; *Priestley et al.*, 2008; *Rodgers and Schwartz*, 1997, 1998] or combinations of such data [*Galvé et al.*, 2002b; *Mejia*, 2001; *Rai et al.*, 2006]. Collectively, these data also suggest that the crust is thickest in the south and southwest, decreasing from 70–80 to 60–70 km in northern and northeastern Tibet. This northward decrease is demonstrated most clearly by *Tseng et al.* [2009], who used *S* waves from nearby earthquakes, sufficiently close that the *P* wave that reflects off the Earth's surface from the incident *S* wave is totally reflected at the Moho; thus, this *SsPwp* phase is recorded as a strong signal following the direct *S* wave. The time interval between the direct *S* wave and *SsPwp* is directly proportional to crustal thickness. *Tseng et al.* [2009] showed a monotonic decrease in crustal thickness from ~75 km beneath southern Tibet to less than 65 km beneath the southern part of northern Tibet. Although some of this thick crust may have existed before the collision with India, most of it, particularly beneath northern Tibet, presumably results from India's penetration into Eurasia.

[62] Insofar as mean elevations, E, of both areas are isostatically compensated, even only approximately, by a Moho deepened by an amount H, we expect E and H to be related by $H = E\rho_c/\Delta\rho$, where $\Delta\rho = \rho_m - \rho_c$, with ρ_m and ρ_c being the densities of mantle and crust. The excess crustal thickness, which is what seismologists commonly estimate, is given by $H + E = E\rho_m/\Delta\rho$. Without allowance for different densities of upper and lower crust, we might assume that $\rho_c = 2.8 \times 10^3$ kg/m^3, $\rho_m = 3.3 \times 10^3$ kg/m^3, and $\Delta\rho = 0.5 \times 10^3$ kg/m^3 and obtain $H = 5.6E$. For different average densities of upper and lower crust, we might allow for $\rho_c = 2.7 \times 10^3$ kg/m^3 but with $\Delta\rho = 0.35 \times 10^3$ kg/m^3, for which $H = 7.7E$. Thus, for perfect Airy isostatic equilibrium, we should expect crustal thicknesses to be 7–8 times those of the mean elevations. For Iran, the ~1500 m mean elevation would call for excess crustal thicknesses of ~10–12 km; assuming both Airy isostasy and that initial thicknesses were 35–40 km, comparable to that of surrounding regions, crustal thicknesses beneath Iran would be 45–50 km. For Tibet, we expect excess thicknesses of 30–35 km from the ~5000 m mean elevation and hence crustal thicknesses of ~70 km. To a first, if only crude, approximation, the two plateaus are in Airy isostatic equilibrium.

[63] Differences in upper mantle structure beneath the plateaus and their surroundings almost surely require that part of the isostatic compensation of both high plateaus be due to density differences in the mantle, if the contribution from low-density mantle is surely smaller than that due to thick crust. In Figure 13a we present a map of anomalies in *S* wave speeds at 125 km depth computed by E. Debayle (personal communication 2009) on a global scale with a horizontal resolution of ~400 km from fundamental and higher-mode surface wave dispersion. This map shows a low-speed anomaly beneath Tibet and a smaller one beneath Afghanistan, but that beneath Iran is not obvious, in part because the few regional stations make resolution poorer than ~400 km. In any case, surface wave dispersion [e.g., *Asudeh*, 1982a; *Kaviani et al.*, 2007; *Kustowski et al.*, 2008; *Maggi and Priestley*, 2005; *Villaseñor et al.*, 2001], *Pn* speeds [*Al-Lazki et al.*, 2003, 2004; *Asudeh*, 1982b; *Chen et al.*, 1980; *Hearn and Ni*, 1994], and *P* and *S* wave tomogra-

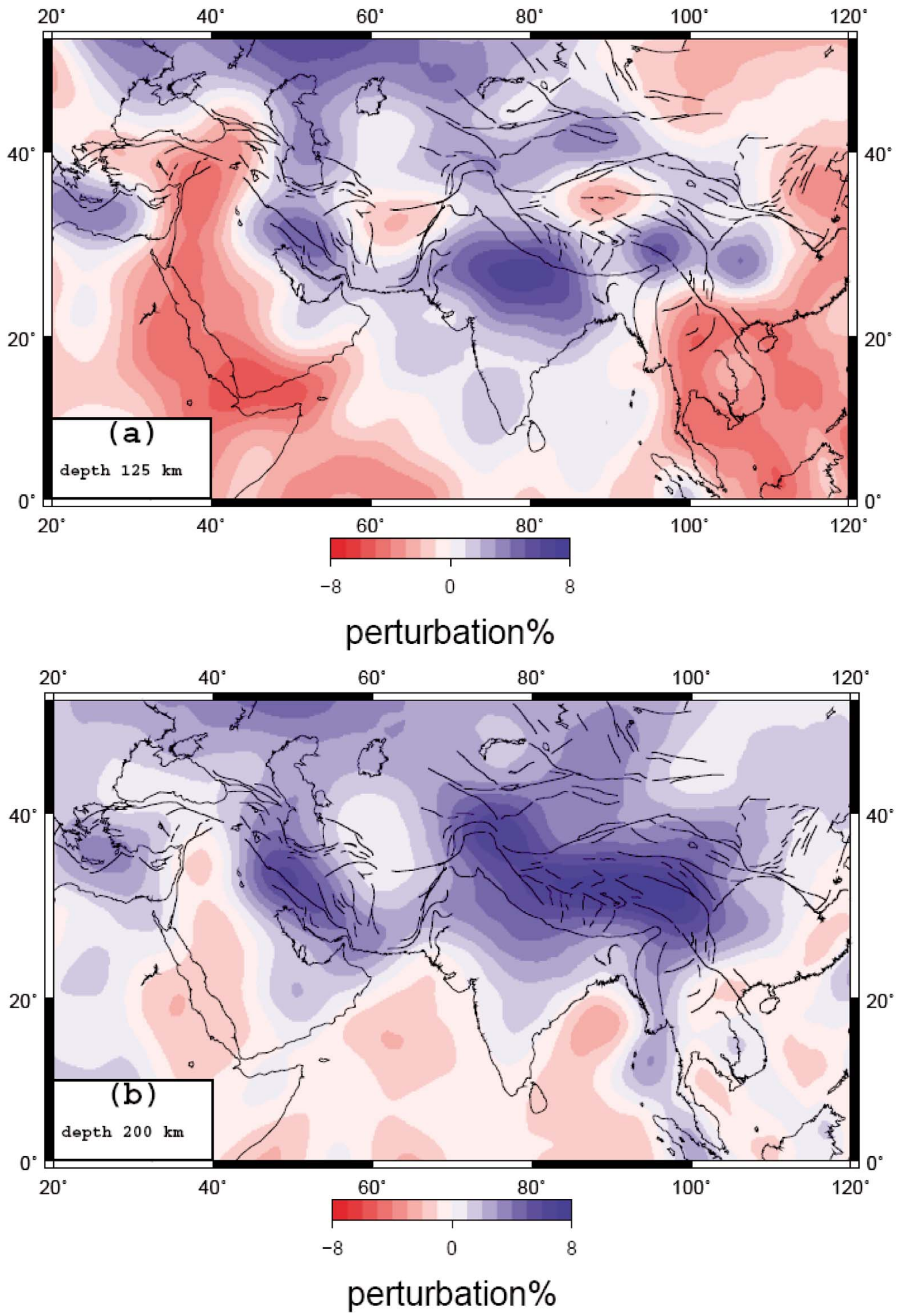


Figure 13. Maps of Asia showing *S* wave speeds at (a) 125 km and (b) 200 km based on surface wave tomography and provided by E. Debayle (personal communication, 2009). They are based on both fundamental and higher-mode Rayleigh and Love waves [*Debayle and Sambridge*, 2004; *Debayle et al.*, 2001; *Maggi and Priestley*, 2005; *Priestley et al.*, 2006]. Note that at 125 km, much of the northern part of the Tibetan Plateau is underlain by relatively low speed material, but low speeds are not apparent beneath Iran. A similar plot (not shown) for 100 km, however, reveals relatively low speeds beneath the Iranian Plateau. At 200 km, however, both are underlain by high-speed material, particularly in their southern parts. This difference between low speed near 100 km and high speed near 200 km is common to studies of the upper mantle of this region [e.g., *Debayle et al.*, 2001; *Kustowski et al.*, 2008; *Priestley and McKenzie*, 2006; *Priestley et al.*, 2006; *Shapiro and Ritzwoller*, 2002; *Villaseñor et al.*, 2001].

phy conducted on a regional scale [*Alinaghi et al.*, 2007; *Hafkenscheid et al.*, 2006; *Kaviani et al.*, 2007; *Paul et al.*, 2010] all call for lower *P* and/or *S* wave speeds in the uppermost mantle beneath the Iranian Plateau, and Turkey, than beneath the Zagros and adjacent Arabian platform. Correspondingly, attenuation of *Sn*, the shear wave that propagates in the uppermost mantle, is high beneath the Iranian and Turkish plateaus but not beneath the Arabian platform or Eurasia [e.g., *Al-Damegh et al.*, 2004; *Kadinsky-Cade et al.*, 1981; *Molnar and Oliver*, 1969; *Sandvol et al.*, 2001]. Low speeds and high attenuation are usually interpreted as implying relatively high temperatures, higher than those beneath shields and stable platforms. Moreover, the contrast between these regions is commonly revealed to be sharp, with high wave speeds beneath the Zagros and lower speeds beneath the region to the northeast [e.g., *Al-Lazki et al.*, 2003, 2004; *Hafkenscheid et al.*, 2006; *Kaviani et al.*, 2007; *Paul et al.*, 2010]. Surface wave tomography consistently shows these lower speeds and higher attenuation to characterize the upper mantle to depths of ~100 km [e.g., *Debayle et al.*, 2001; *Kaviani et al.*, 2007; *Kustowski et al.*, 2008; *Priestley and McKenzie*, 2006; *Villaseñor et al.*, 2001] (Figure 13b). Readers may notice the apparent contradiction between abundant evidence cited above for relatively low *P* and *S* wave speeds and high attenuation beneath the Iranian Plateau and normal if slightly high speeds in Figure 11a; this is due to the choice of 125 km. A plot for shallower depths, however, would introduce the risk of error beneath Tibet because of contamination by the thick crust there. At depths greater than ~100 km and beneath the Zagros, higher speeds, at least of *S* waves, are found in all surface wave studies (Figure 11).

[64] For southern Tibet, surface wave dispersion shows that at depths near 100 km *S* wave speeds are quite high, like those beneath a shield, but they are much lower under northern Tibet [e.g., *Brandon and Romanowicz*, 1986; *Holt and Wallace*, 1990; *Kustowski et al.*, 2008; *Priestley et al.*, 2006; *Rodgers and Schwartz*, 1997, 1998; *Shapiro and Ritzwoller*, 2002; *Villaseñor et al.*, 2001]. The relatively low speeds in the upper mantle beneath northern Tibet are corroborated by a variety of other observations: low *Pn* speeds [*Beghoul et al.*, 1993; *Hearn et al.*, 2004; *Liang and Song*, 2006; *McNamara et al.*, 1997; *Meissner et al.*, 2004; *Zhao and Xie*, 1993], delays of *S* waves from earthquakes there [*Molnar*, 1990; *Molnar and Chen*, 1984], delays in *SS* phases reflected beneath this area [*Dricker and Roecker*, 2002; *Lyon-Caen*, 1986; *Woodward and Molnar*, 1995], and high attenuation of *Sn* [*Barazangi and Ni*, 1982; *Molnar and Oliver*, 1969; *Ni and Barazangi*, 1983]. Thus, a wealth of evidence calls for a low-speed, highly attenuating uppermost mantle beneath northern Tibet and hence the suggestion that today this area is relatively hot, as would be the case if mantle lithosphere were thin.

[65] At greater depth, centered near ~200 km, *S* wave speeds are relatively high beneath most of Tibet (Figure 13b) [e.g., *Debayle et al.*, 2001; *Kustowski et al.*, 2008; *Priestley and McKenzie*, 2006; *Priestley et al.*, 2006; *Rapine et al.*, 2003; *Shapiro and Ritzwoller*, 2002; *Villaseñor et al.*, 2001; *Zhou and Murphy*, 2005]. Thus, the structure beneath Tibet shares similarities with Precambrian shields in that high speeds exist at depths of 200–250 km [*Priestley and McKenzie*, 2006; *McKenzie and Priestley*, 2008]. Two very different interpretations have been assigned to this widespread high-speed zone, and reasons to doubt both are not hard to see.

[66] Shortly after *Jordan* [1975] proposed that the lithosphere beneath Precambrian shields was much thicker than beneath old oceanic lithosphere, he suggested that the thick lithosphere (which he dubbed "tectosphere") might form where horizontal shortening over a wide region not only thickened crust, as has occurred beneath Tibet, but also the mantle lithosphere [*Jordan*, 1978, 1988]. Recently, *McKenzie and Priestley* [2008] revived this view for the high *S* wave speed regions beneath both Iran and Tibet and proposed a very thick (~250 km) lithosphere beneath both Tibet and the Zagros. Obviously, the low speeds in the uppermost mantle, down to ~100 km or more beneath northern Tibet and beneath the Iranian Plateau, which seem inescapable from all inferences of seismic wave speeds, appear to be inconsistent with thick lithosphere. Both *Chen and Molnar* [1981] and *McKenzie and Priestley* [2008] showed, however, that for sufficient radiogenic heating in the thickened crust, albeit with different initial conditions and different bottom boundary conditions to the lithosphere, the adjacent mantle just below the Moho could warm hundreds of degrees Celsius tens of millions of years after thickening occurred. The conditions necessary for sufficient warming are, at least in our opinion, near the extreme for plausibly accounting for the low speeds beneath southern Tibet. *McKenzie and Priestley* [2008] illustrated the case of crust initially 25 km thick and shortened and thickened instantaneously to 75 km. They assumed a mantle lithosphere sufficiently thick that initially, the surface might have lain below sea level, a state not appropriate for northern Tibet. The concentration of radioactivity in the upper 20 km of 2 μW/m^3 is relatively large. Even after 40 Ma, their calculated warming of the uppermost mantle is modest, with depths initially at ~500°C reaching only ~650°C and with the 900°C isotherm remaining below 160 km for 150 Myr. Thus, if applied to northern Tibet, where the present-day crustal thickness of <65 km [*Tseng et al.*, 2009] was probably not reached until 10–20 Ma, radiogenic heating from the crust seems only remotely plausibly capable of accounting for the low-speed uppermost mantle. Moreover, with less crustal thickening beneath the Zagros and Iranian Plateau than Tibet and a more recent collision, even if *McKenzie and Priestley*'s [2008] calculation did account for the low seismic wave speeds beneath northern Tibet, radiogenic heating in thickened crust seems unlikely to account for the low speeds in the uppermost mantle beneath the Iranian Plateau.

[67] Alternatively, the high-speed material may mark the presence of cold lithospheric mantle that has been convectively removed from beneath Tibet and either has ponded at these depths or is passing through this portion of the upper mantle [e.g., *Houseman and Molnar*, 2001]. We

adopt this view, but readers should recognize that it is neither required nor devoid of difficulties. Accordingly, the inference that the high-speed zone beneath Tibet centered near 200 km reflects such ponding of convectively removed mantle lithosphere requires adopting additional assumptions. First, essentially all studies showing this high-speed zone show that speeds are higher beneath southern and southwestern Tibet than beneath northern or northeastern Tibet. In this respect, images of *P* wave and *S* wave speeds in the upper mantle beneath Tibet between 100 and 200 km from body and surface wave tomography, respectively, resemble one another [see *Li et al.*, 2008, Figure 10]. These images differ beneath northern Tibet, however, where *P* wave speeds are shown to be relatively low but *S* wave speeds are relatively high. For southern Tibet, many imagine that the high-speed zone between ~100 and ~300 km results simply from Indian lithosphere plunging at a gentle angle beneath southern Tibet [e.g., *Tilmann et al.*, 2003; *Zhou and Murphy*, 2005], but *Grand* [2002] inferred that high *S* wave speeds extend deeper than 300 km.

[68] Perhaps more problematic for an interpretation in terms of ponding of mantle lithosphere beneath northern Tibet is the apparent termination of high *S* wave speeds at a depth between 200 and 250 km; if lithosphere were convectively removed, why would it collect at this depth and not deeper? In fact, *Chen and Tseng* [2007] found relatively high *P* wave speeds at depths near 650 km beneath Tibet, and they inferred that this material represents foundered mantle lithosphere from beneath Tibet. Later, *Tseng and Chen* [2008] reported normal *S* wave speeds in the same region and postulated that hydrated mantle lithosphere, which would have accumulated beneath Tibet during subduction, had later become detached and then had sunk to these depths. Anyhow, regarding ponding of convectively removed lithosphere at ~250 km, we remind readers that in the depth range of ~100 to ~250 km, temperature profiles approach the solidus for typical mantle material; if the presence of partial melt reduced viscosity substantially, then the base of that partially molten zone would also mark an increase in viscosity. We are not aware of direct evidence, however, of a marked increase in viscosity at this depth.

[69] The high *S* wave speeds centered near 200 km beneath both plateaus may be due, at least in part, to a greater sensitivity of wave speeds to temperature near this depth than at greater depths. The temperature dependences of *P* and *S* waves ($\partial V/\partial T$) are relatively constant at low temperatures, but as temperatures approach the solidus, they become much larger [e.g., *Goetze*, 1977], so that wave speeds are more sensitive to temperature near 200 km than deeper [e.g., *Cammarano et al.*, 2003]. Thus, the presence of a high-speed zone near 200 km might give us an exaggerated image of lateral heterogeneity. In this respect, perhaps it is noteworthy that the most sharply resolved example of an apparently sinking blob of mantle lithosphere, the "Isabella anomaly" beneath the southwest edge of the Sierra Nevada and southeast portion of the Great Valley of California, is clear to a depth of ~220 km but not deeper [*Reeg et al.*, 2007].

[70] Surface waves, whose lateral resolution is ~400 km [e.g., *Debayle et al.*, 2001; *McKenzie and Priestley*, 2008; *Priestley and McKenzie*, 2006], cannot distinguish lateral heterogeneity on the horizontal length scale likely for convective processes in the uppermost mantle. If mantle lithosphere beneath Tibet were being removed today, however, its upper mantle should be laterally heterogeneous, and several studies suggest that it is. In an attempt to match body wave seismograms for paths passing under Tibet, *Zhao et al.* [1991] reported that no single layered model worked and that lateral heterogeneity was required. Constraints on the structure deeper than ~200 km beneath Tibet derive largely from *P* and *S* wave tomography, and although differences in wave speeds are not great, essentially all studies call for lateral heterogeneity. *Zhou et al.* [1996] showed that differences in *P* wave arrival times from earthquakes in central and southwestern Tibet require marked differences in upper mantle structure, and assuming that the structure beneath southern Tibet accounted for most of the residuals, they inferred that a high-speed zone striking parallel to the Himalaya must plunge nearly vertically to a depth of ~400 km beneath southwestern Tibet and the adjacent Himalaya. *Pandey et al.* [1991], relying on recordings in Nepal, also showed high speeds reaching to 300 km beneath the Karakorum of southwesternmost Tibet. Seismograph stations in most of the Himalaya and Tibet, however, are sparse, and body wave tomography does not sample its upper mantle well. Nevertheless, *Hafkenscheid et al.* [2006] and *Li et al.* [2008] also show high *P* wave speeds reaching to such depths on some profiles across the Himalaya and the adjacent parts of southwestern and southern Tibet. Using a north-south line of seismographs deep into the interior of Tibet, *Tilmann et al.* [2003] reported a nearly vertical, high-speed zone at depths between ~100 and ~400 km. Using these and other data, *Li et al.* [2008] show a high-speed zone in the same area but shifted deeper to ~300–600 km. Although interpretations of the origin of this high-speed material differ, its presence argues for lateral heterogeneity beneath the depth interval (150–250 km) where high *S* wave speeds are common. Finally, in eastern Tibet, *Ren and Shen* [2008] show regions of high and low *P* wave speeds plunging into the mantle to depths of nearly 400 km and with lateral dimensions of ~100 km. These zones also require marked lateral heterogeneity. Only a small fraction of Tibet is instrumented well enough to allow definition of lateral heterogeneity with lateral dimensions of ~100 km, but where sufficiently dense networks have been deployed, such heterogeneity seems to be present in the upper mantle beneath the plateau.

[71] Body wave tomography of the upper mantle beneath the Iranian Plateau also allows for lateral heterogeneity but with less resolution than for Tibet. *Hafkenscheid et al.* [2006] show a high-speed zone continuing beneath the Zagros and perhaps marking subducted ocean lithosphere that led the Arabian shield into the subduction zone. By contrast, in a focused attempt to detect such a slab of lithosphere, *Kaviani et al.* [2007] found that delays within the crust masked an equivocal travel time advance due to such

a slab, and they could neither demonstrate nor deny its existence. *Kadinsky-Cade and Barazangi* [1982] interpreted a single earthquake at a depth of 107 km and with a T axis plunging northeast as a hint of remnant lithosphere beneath the southeast end of the Zagros. *Maggi et al.* [2000], however, associated this event with subduction of oceanic lithosphere beneath the Makran coast, although the epicenter lies slightly west of the west end of that subduction zone. Unequivocal evidence of either oceanic lithosphere attached to Arabia or heterogeneity in the upper mantle that might reflect ongoing convective removal of lithosphere beneath the Iranian Plateau remains to be found.

[72] In summary, the crust and upper mantle structures of Iranian and Tibetan plateaus differ in several respects. Crust is much thicker beneath Tibet. The upper mantle beneath both includes high-speed material not only beneath the Zagros and Himalaya but also beneath the adjacent portions of the Iranian and Tibetan plateaus, consistent with the underthrusting of platform- or shield-like lithosphere beneath the belts. Beneath the Iranian Plateau and the northern part of Tibet, however, seismic wave speeds from the Moho to depths of ~100 km are lower than those typical of stable shields and platforms, and attenuation is high. Thus, we presume that the upper mantle is relatively warm and lithosphere is thin there. At greater depths (>150 km) beneath both plateaus (Figure 13b), S wave speeds again are quite high, suggesting the presence of relatively cold material whose origin remains a subject of discussion and which might indicate thick lithosphere, whose upper part is warm because of radiogenic heating in the overlying crust [*McKenzie and Priestley*, 2008]. At greater depths, several studies show lateral variations in P and S wave speeds beneath Tibet, consistent with active mantle dynamics beneath the plateau, but such evidence from Iran remains inconclusive. Finally, as with all differences between Iran and Tibet, the contrasts are greater beneath Tibet, with lower wave speeds just below the Moho and with a greater extent of high speeds at depths of ~200 km.

5. DISCUSSION: IMPLICATIONS FOR MOUNTAIN BUILDING AND THE GROWTH OF HIGH PLATEAUS

[73] We have focused mostly on how the Zagros and Himalaya and their adjacent plateaus differ from one another, but perhaps it is worth recounting first how they are similar. Both the Zagros and the Himalaya formed after stable continental crust (a platform and a shield) and sedimentary rock deposited on their continental shelves plunged into subduction zones along the southern margin of Eurasia. Virtually all rock of both mountain ranges was part of Arabian or Indian, not Eurasian, crust. In both cases, when collision occurred, the rates of convergence between the two plates, Arabia and Eurasia in one case and India and Eurasia in the other, slowed by ~35% and ~30%–45%, respectively, presumably because the thick continental crust made the subducting plates more buoyant than when only oceanic lithosphere was consumed. As these continents penetrated into Eurasia, sedimentary rock deposited on their northeastern and northern margins shortened horizontally by folding and thrust faulting as the loci of deformation progressively moved onto the Arabian and Indian continents. Later the underlying basement deformed by reverse and thrust faulting, with the locus of deformation stepping southward into the impinging continents. Convergence continues today, as both GPS and seismicity attest. Following both collisions, wide plateaus grew within the adjacent southern portions of Eurasia, as Arabia and India penetrated into them. Shortening and isostatically compensated thickening of crust account for most of the present-day elevations of the plateaus, but portions of both are also underlain by uppermost mantle with lower seismic wave speeds than beneath the adjacent stable regions. Hence, it appears that mantle lithosphere beneath at least parts of both plateaus is thin or absent and presumably has been removed since the collisions (but as noted in section 4.2, not all agree either that lithosphere is thin or that it has been removed). Crustal shortening continues in the mountain belts that surround the plateaus, not just in the Zagros and Himalaya, suggesting that both plateaus are growing outward.

[74] Turning to the differences between the Zagros and Himalaya and between the Iranian and Tibetan plateaus, we discuss those that seem to offer insights into how collision zones evolve. These include the development of major thrust fault systems, the growth of high plateaus including the roles of crustal thickening and underlying mantle dynamics, and the significance of movements of apparently rigid crustal blocks in that growth.

5.1. Initiation of "Main" Throughgoing, Intracrustal Thrust Faults

[75] The Himalaya seems to be at a stage that is far advanced beyond that of the Zagros [e.g., *Barazangi*, 1989; *Ni and Barazangi*, 1986]. Convergence has occurred by slip on major, termed "Main," thrust faults that dip gently beneath the range (Figures 5 and 14). Following collision, the first such fault, the Main Central Thrust, broke through the Indian crust, and today it appears as a wide shear zone with several splays. Subsequently, the locus of slip at the surface moved southward to the Main Boundary Fault, which separates pre-Mesozoic rock from Cenozoic deposits. Currently, slip occurs on the Main Frontal Thrust, a listric splay that cuts the late Cenozoic sedimentary cover and manifests itself as a piedmont fold belt within the Ganga Basin south of where it joins the gently dipping fault that presumably was active when slip occurred on the Main Boundary Fault.

[76] The Zagros also have been built by the horizontal shortening, initially by folding of thick sedimentary rock at the surface and, apparently more recently, also by reverse faulting in the basement that once underlay Arabia's continental margin (Figures 5 and 14). Although Arabian crust appears to have been underthrust beneath the margin of the Iranian Plateau [e.g., *Agard et al.*, 2005, 2006; *Paul et al.*, 2006, 2010], crustal thickening within much of the Zagros has primarily involved the sedimentary strata that formed on

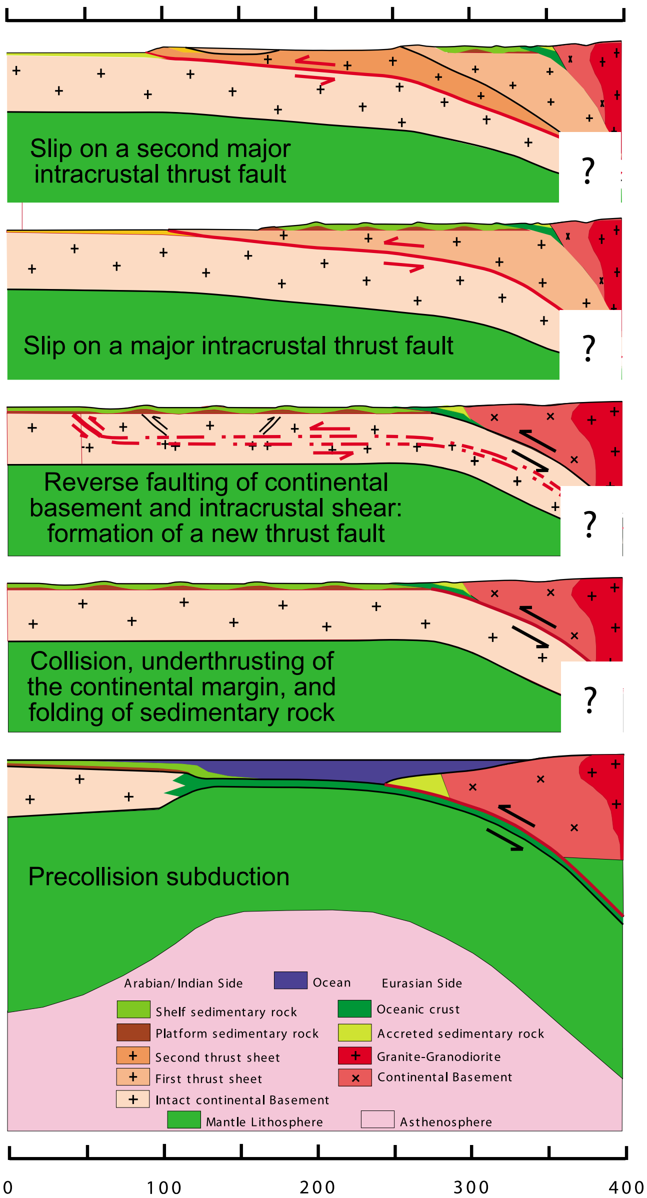


Figure 14. Cartoons illustrating a hypothesized development of major thrust faults that cut through the entire crust, using the Zagros and Himalaya to illustrate stages in that development. The fifth panel shows underthusting of oceanic lithosphere beneath a continent, with Arabia or India following that lithosphere. The fourth panel shows the state after the margin of that continent has been thrust to depth beneath the overriding continent and folding of the sedimentary cover has begun. In the third panel, reverse faulting occurs over a broad part of the upper crust, while aseismic shear occurs within the lower crust; we imagine that this represents the current state in the Zagros. In the second panel, displacement of tens or hundreds of kilometers has occurred, and the shear zone has contracted to become a fault as the underthrusting continent slides beneath colder upper crust. In addition, the lithosphere to its left has been flexed down to form a foreland basin. In the first panel, this process has been repeated, a second major thrust fault has developed, and slip of 150 km or more on it has accrued; we imagine this to mimic crudely the upper lithospheric structure of the present-day Himalaya.

the ancient margin of Arabia. Apparently, only little thickening of the basement itself has occurred. The same folding of sedimentary rock occurred in the Himalaya, but all of this record now lies north or northeast of the high peaks in what is sometimes also called the Tethyan Himalaya and geographically is now part of southern Tibet. The active tectonic processes that continue to build the Himalaya reveal themselves south of and beneath the Greater Himalaya. They have involved the shearing of deeper basement of India's northern continental margin from the rest of India and the thrusting of these thin sheets of basement rock atop the rest of India, as the intact portion of the Indian subcontinent has plunged beneath southern Tibet (Figure 5). When (and if) the analogous process in the Zagros happens, it should involve the creation of a gently dipping thrust fault through the basement beneath the present-day Zagros and the thrusting of that sheet of crystalline basement and the overlying sedimentary rock onto the Arabian shield. The surface outcrop of such a fault would lie within and northwest of the present-day Persian Gulf.

[77] The present-day seismicity and active tectonics of the Zagros suggest that this initiation of a new thrust fault may be occurring today and will be established soon [e.g., *Mouthereau et al.*, 2006, 2007]. Fault plane solutions of earthquakes in the Zagros do not show a gently dipping nodal plane like those for earthquakes in the Himalaya, and thus, no major thrust fault seems to have formed yet. At the same time, most modern analyses of the deformation of the Zagros suggest that the basement of the Zagros is involved in the current deformation [*Berberian*, 1995; *Bosold et al.*, 2005; *Molinaro et al.*, 2005; *Mouthereau et al.*, 2006, 2007; *Sherkati et al.*, 2005]. Although much of the horizontal shortening of the sedimentary layers in the Zagros has occurred by folding of sedimentary rock detached from the basement along weak layers of salt or other material, deformation in the Zagros is no longer strictly, nor is it entirely, thin skinned. The seismicity shows that active faulting beneath the Zagros occurs largely by reverse faulting [e.g., *Jackson and McKenzie*, 1984; *Maggi et al.*, 2000; *McKenzie*, 1972a; *Talebian and Jackson*, 2004] but that activity is more intense beneath the lower, southwestern half of the Zagros than beneath the higher part (Figures 8 and 9). Moreover, GPS measurements show a concentration of strain accumulation in this southwestern part of the Zagros. Where accurately located, small earthquakes occur in clusters with a different spacing from the folds at the surface, suggesting that those folds do not result from slip on adjacent, active faults in the basement [*Tatar et al.*, 2004]. Fault plane solutions of these accurately located earthquakes suggest that most, but not all, clusters mark faults dipping northeast in the upper crust. We imagine that these seismic faults merge into a ductile shear zone in the middle to lower crust (Figure 14).

[78] In most images of the strength of the crust, a brittle layer overlies a ductile layer, which in turn overlies a stronger uppermost mantle [e.g., *Brace and Kohlstedt*, 1980]. Many associate the boundary between brittle and ductile deformation with an isotherm near 350°C, where

quartz flows at relatively low stress [e.g., *Chen and Molnar*, 1983; *Meissner and Strehlau*, 1982; *Sibson*, 1982]. Because quartz deforms more readily than feldspars or pyroxenes at crustal temperatures [e.g., *Bürgmann and Dresen*, 2008; *Mackwell et al.*, 1998; *Rybacki and Dresen*, 2004; *Rybacki et al.*, 2008], many allow for a stronger lowermost crust, rich in feldspar and pyroxene, than middle crust. (We note also that some consider the strength of continental lithosphere to lie largely in the crust and not in the mantle [*Jackson et al.*, 2008; *McKenzie and Priestley*, 2008; *Priestley et al.*, 2008].) Thus, it makes sense for a major, gently dipping thrust fault to form within the middle to lower crust (and not at the Moho). Yet such a fault must also rupture the upper crust, which almost surely requires slip initially on a reverse fault (dip ~ 45°). Reverse faulting currently occurs within the Zagros and specifically in the portion of the Zagros where elevations are relatively low and closer to the intact Arabian shield than the high, northeastern part of the Zagros. Thus, we imagine that the Arabian shield has plunged beneath the Iranian Plateau, as images based on receiver functions suggest [*Paul et al.*, 2006, 2010], but that convergence along this zone is resisted and slip on it no longer occurs. Today, the underthrusting of the Arabian shield induces horizontal compressive strain, which manifests itself in both seismicity and reverse faulting beneath the folded and thrust-faulted sedimentary cover. Loading of that basement may include some ductile shear at middle to lower crustal depths beneath the higher parts of the Zagros.

[79] With some reasonable assumptions, we may estimate the total strain within the basement. The increase in elevation from the Persian Gulf to central Iran includes steps in the basement due to reverse faulting. *Berberian* [1995] inferred 6 km of total vertical difference, from offsets in sedimentary layers at different faults. *Mouthereau et al.* [2006] showed a similar amount. If we assumed that the reverse faults dip at 30°–60°, as fault plane solutions of earthquakes imply, this 6 km would translate into 3.5–10 km of horizontal shortening. Recognizing that reverse slip on faults dipping at 60° is not likely [e.g., *Sibson*, 1990], however, and that a range of dips of 30°–45° seems more sensible, likely amounts of shortening of the basement are 6–10 km. At GPS-measured convergence rates of 5–9 km/Myr across the Zagros, these amounts suggest that this basement involvement began between ~0.7 and 2 Ma.

[80] Constraints on the structure, seismicity, and GPS geodesy imply that beneath the southwestern part of the Zagros, the upper Arabian crust not only has been thrust beneath the Zagros but also currently undergoes northeast-southwest shortening. The absence of seismicity beneath the northeast portion of the Zagros and its approximately constant rate of convergence with the stable Arabian shield suggest that this section of upper crust moves southwest with respect to Arabia with little active deformation and hence effectively rigidly (Figure 14). The convergent strain across the low-elevation, southwestern parts of the Zagros, which is seen at the surface with GPS (Figure 8) and with active folding [e.g., *Oveisi et al.*, 2007, 2009] and in the upper crust as seismicity, must be absorbed at depth. We presume that aseismic shear within the middle and lower crust accommodates the convergence between the intact portion of the Arabian plate and the upper crust of the high-elevation, northeastern parts of the Zagros. Thus, we imagine that shear occurs within a nearly horizontal zone of finite thickness in the middle and lower crust beneath the higher Zagros (Figure 14). Shear on that zone would load the brittle upper crust farther southwest, where earthquakes release accumulated strain. The third panel of Figure 14 presents this view, and the first and second panels show subsequent evolution that would lead to a structure similar to that in the present-day Himalaya.

[81] This inference of straining of the basement beneath the Zagros and the eventual initiation of a new thrust fault in the lower crust is not without inconsistencies. The summation of seismic moments of earthquakes [e.g., *Jackson and McKenzie*, 1988; *Masson et al.*, 2005; *North*, 1974] yields a rate of shortening of the basement that is markedly lower than the rate of shortening measured with GPS. Accordingly, much of Arabia's convergence with Eurasia must occur by a process that earthquakes do not record.

5.2. Growth of High Plateaus and a Role for Mantle Dynamics

[82] Again, the different histories of deformation of Tibet and Iran point to different stages of development. Tibet has undergone large (if poorly quantified) north-south shortening [e.g., *Chang et al.*, 1986; *Chen et al.*, 1993; *Coward et al.*, 1988b; *Dewey et al.*, 1988], but today it extends east-west. The Iranian Plateau, by contrast, has undergone less shortening, rose more recently, and does not yet extend by normal faulting.

[83] A thin viscous sheet in a gravity field and penetrated by a rigid object provides a simple model that accounts for many aspects of large-scale deformation of continental lithosphere [e.g., *Bird and Piper*, 1980; *England and Houseman*, 1986; *England and McKenzie*, 1982, 1983; *Flesch et al.*, 2001; *Houseman and England*, 1986; *Jiménez-Munt and Platt*, 2006; *Vilotte et al.*, 1986]. With only two free parameters, the thin viscous sheet yields surface elevations that match the hypsometry and the pattern of the regional rotations in eastern Asia [*England and Houseman*, 1986], as well as the simple relationship between gradients in gravitational potential energy and gradients in strain rates [*England and Molnar*, 1997]. A basic assumption used to exploit the thin viscous sheet is that deformation varies little with depth through the thickness of the lithosphere, and two sets of observations support this assumption, at least as applied to eastern Asia. First, fault plane solutions of shallow focus earthquakes in southern Tibet and those in the underlying uppermost mantle show the same pattern of east-west extension by normal faulting on north-south planes and conjugate strike-slip faulting [e.g., *Chen et al.*, 1981; *Chen and Yang*, 2004; *de la Torre et al.*, 2007; *Molnar and Chen*, 1983; *Zhu and Helmberger*, 1996]. Second, orientations of strain or strain rate in Tibet measured from geological or geodetic observations at the Earth's surface are aligned parallel to orientations of the

faster and slower quasi–*S* wave deduced in studies of shear wave splitting [e.g., *Davis et al.*, 1997; *Flesch et al.*, 2005; *Holt*, 2000; *Lev et al.*, 2006; *Sol et al.*, 2007; *C.-Y. Wang et al.*, 2008]. Of course, in its simplest form with no shear stresses on horizontal planes, the thin viscous sheet is inapplicable to the margins of Tibet, like the Himalaya.

[84] Insofar as the thin viscous sheet is applicable to a region in front of strong indenting continental lithosphere, like Arabia or India, the lateral extent of the deforming region should scale with the width of the indenter, so that an extensive region should undergo strain concurrently [*England et al.*, 1985]. In addition, the locus of the most rapid shortening and thickening should migrate into the sheet away from the edge of indenter as the latter penetrates into the sheet [e.g., *England and Houseman*, 1986; *England and McKenzie*, 1982, 1983; *Houseman and England*, 1986; *Vilotte et al.*, 1986]. As noted in section 3.1, geologic evidence for Tibet suggests not only an initially wide zone of deformation, facilitated by the presence of strong regions on Tibet's northern margin [*Dayem et al.*, 2009; *England and Houseman*, 1985; *Flesch et al.*, 2001; *Vilotte et al.*, 1984], but also the suggestion of a subsequent northward growth of the locus of the most intense deformation [e.g., *Métivier et al.*, 1998; *Tapponnier et al.*, 2001]. Although the temporal pattern of deformation in Iran is less well constrained than in Tibet, the apparent rise of the Alborz near ~20 Ma [*Ballato et al.*, 2008, 2010] or perhaps more sharply at only 12 Ma [*Guest et al.*, 2007] and the inference that the Kopet Dagh dates from ~10 Ma [*Hollingsworth et al.*, 2006, 2008] suggest that deformation on the surroundings of the Iranian Plateau began after Arabia collided with Eurasia but not necessarily long after that time.

[85] At present, deformation in the two regions differs in two obvious ways. First, in Tibet, approximately east-west crustal extension occurs by a combination of normal faulting, which implies crustal thinning, and conjugate strike-slip faulting (Figure 9). In Iran, normal faulting is essentially absent. Second, the Himalaya is overthrust onto India in a direction radially outward, perpendicular to the local trend of the belt, not parallel to India's convergence with respect to Eurasia (Figure 7). In the Zagros, however, the obliquity of Arabia's convergence with Eurasia pervades deformation across much of the Zagros and the Iranian Plateau, and convergence across the belt varies in both rate and orientation along it (Figure 7). These two differences can be understood as reflecting a difference in the development of the two regions.

[86] In the Zagros and the Iranian Plateau, the penetration of Arabia into Eurasia manifests itself in part by lateral transport of material out of Arabia's path, as northwestern Iran and eastern Turkey move westward past the relatively strong lithosphere beneath the Black and Caspian seas and toward the Aegean Sea (Figure 6). Reverse faulting and crustal thickening occur not just beneath the Zagros but also beneath the Alborz and Kopet Dagh to the north and beneath short ranges that link strike-slip faults in the region surrounding the Lut Block. Throughout most of Iran, however, crust beneath the plateau has not thickened much, if at all (see Figure 12 for a general illustration of this). Yet this region stands high, and its surface uplift must have occurred since the collision for Oligocene marine sediment covers parts of central Iran [e.g., *Berberian and King*, 1981; *Davoudzadeh et al.*, 1997; *Morley et al.*, 2009; *Reuter et al.*, 2009; *Schuster and Wielandt*, 1999; *Stöcklin*, 1968, 1974, 1977]. We suppose, but are by no means convinced, that mantle lithosphere has been removed from beneath central Iran, presumably as a result of convective instability, and the added buoyancy of hotter material beneath this region requires that its surface stand high. The forces that conspire to sustain Arabia's penetration into Eurasia seem, however, to have done so with little diminution in strength for the speed with which Arabia moves seems to have changed little since 20 Ma (Figure 4).

[87] By contrast, the widespread normal faulting and crustal thinning in Tibet call attention to a change in the style of deformation and in the balance of forces acting on the plateau. First, high plateaus cannot be built by normal faulting. Almost surely, the crust beneath most of Tibet has thickened since India collided at 45–55 Ma [e.g., *Chang et al.*, 1986; *Chen et al.*, 1993; *Coward et al.*, 1988b; *Dewey et al.*, 1988], but normal faulting implies crustal thinning, not thickening. When crustal thickening of Tibet ended and crustal thinning began are not constrained well. We do not concern ourselves here with normal faulting that occurred along the southern edge of Tibet near 20 Ma [e.g., *Burchfiel et al.*, 1992; *Burg et al.*, 1984; *Herren*, 1987; *Valdiya and Goel*, 1983]. The oldest precisely dated northerly trending normal fault within Tibet seems to have been active at ~13.5 Ma [*Blisniuk et al.*, 2001], and other normal faults date from ~8 (±3) Ma [e.g., *Harrison et al.*, 1995; *Pan and Kidd*, 1992]. One implication of both this normal faulting and the radially outward overthrusting of southern Tibet onto India is that some change occurred in the balance of forces on the margin of Tibet and stress within it. *England and Houseman* [1989] discussed possible changes and argued that the only sensible possibility is that mantle lithosphere was removed from beneath Tibet and the plateau rose 1000 m or more. The increased elevation of the plateau would then increase the vertical compressive stress, transforming it from the minimum principal compressive to the maximum compressive stress. Concurrently, the force per unit length that the plateau and surrounding areas apply to one another would increase and resist further horizontal shortening within Tibet, so that deformation of lower surrounding territory would become favored.

[88] *England and Houseman* [1989] discussed the possibility that the plateau became weaker and that this weakening facilitated normal faulting, and they showed this possibility to be illogical. In regions of crustal shortening and thickening, the forces (per unit length) that build the thick crust must do work against dissipative processes, such as friction on faults or viscous flow in continuous medium, and against gravity to create potential energy. Accordingly, if the region became weaker but the horizontal driving force (per unit length) remained the same, the resistance to gravity must necessarily increase; if that

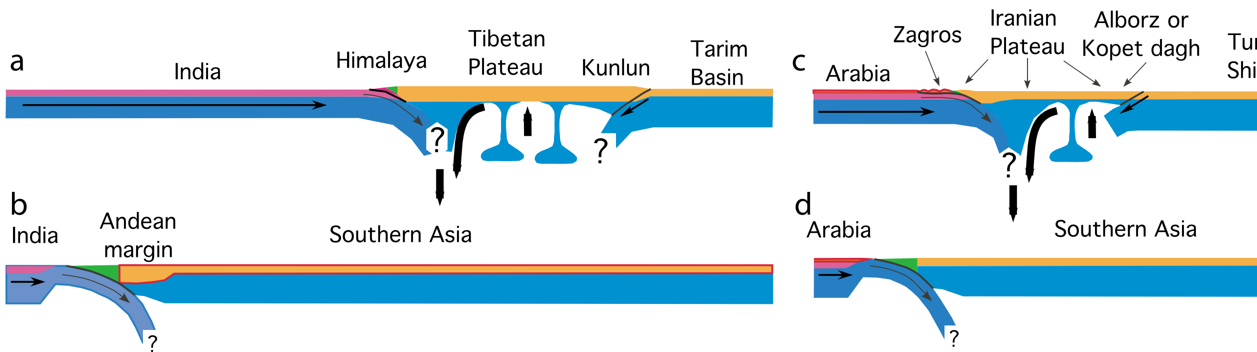


Figure 15. Cartoon cross sections contrasting hypothesized evolutions of deep structure beneath the (a and b) Tibetan and (c and d) Iranian plateaus. Figures 15b and 15d show precollision states with India and Arabia following oceanic lithosphere into the asthenosphere beneath southern Eurasia. For Tibet (Figures 15a and 15b), we imagine continued convergence of 2000–3000 km (but not with amounts scaled accurately); thickening of Eurasian crust to build the wide, high Tibetan Plateau; removal of mantle lithosphere possibly in several downwelling plumes or sheets; possible ponding of that material at depths where viscosity increases sufficiently to retard further sinking; and the absence of intact mantle lithosphere beneath much of northern Tibet. Evidence for such plumes or sheets is sparse, but some tomograms do show lateral heterogeneity on this scale [e.g., *Li et al.*, 2008; *Ren and Shen*, 2008; *Tilmann et al.*, 2003]. For Iran (Figures 15c and 15d), there has been notably less convergence between Arabia and Eurasia than between India and Eurasia. Nevertheless, with lithospheric thickening beneath parts of Iran, we suppose (tentatively) that mantle lithosphere became unstable and sank beneath the margins of the Iranian Plateau, leaving thinned mantle lithosphere beneath the plateau. As beneath Tibet, that detached mantle lithosphere is shown to have ponded where viscosity may increase, but this suggestion is yet more tentative than for Tibet. For Iran obliquity of various mountain ranges (Zagros, Alborz, Kopet Dagh, and others) makes a 2-D cross section more misleading than for Tibet, and assigning specific localities to regions of downwelling is risky.

happened, the crust would thicken, which would require reverse or thrust, not normal, faulting. Weakening of a high terrain built by horizontal compression and crustal thickening cannot, at least in the absence of other changes or additional assumptions, initiate crustal thinning.

[89] We presume that this major difference between the Tibetan and Iranian plateaus results from their being at different stages. The lower seismic wave speeds beneath the plateaus than beneath their surroundings suggest warmer upper mantle, consistent with both having lost at least some of their mantle lithosphere (though again, we remind readers that this inference is controversial).

[90] For Tibet, numerous observations suggest that the region surrounding Tibet underwent accelerated deformation since ~15 Ma, possibly in response to removal of mantle lithosphere. Such removal of cold dense material and its replacement with warmer material would have led to a rise of its surface of ~1000 m [e.g., *Molnar and Stock*, 2009]. Moreover, as discussed in section 3.1, India's rate of convergence with Tibet decreased by 45% near 15 Ma, as might be expected if the force per unit length that Tibet applies to India increased (though as noted in section 3.1, *Copley et al.* [2010] question this decrease in rate). Thus, insofar as the deformation surrounding Tibet and the change in the convergence rate of India with Eurasia offer tests of the idea that mantle lithosphere was removed from beneath Tibet, such removal passes these tests. Also, as noted in section 4.2, virtually all studies of upper mantle structure of Tibet call attention to low speeds in the uppermost mantle beneath northern Tibet but high speeds at depths of ~200 km (and possibly greater). Speeds in this depth range resemble those of Precambrian shields, and some have suggested that horizontal compression of Tibetan lithosphere if manifested by thickening of it would create such a structure [*Jordan*, 1978, 1988; *McKenzie and Priestley*, 2008]. We suspect instead that this high-speed material marks the active removal, and perhaps ponding, of cold material that formerly was part of Tibet's mantle lithosphere that became convectively unstable during horizontal shortening and that is sinking or has sunk through the underlying asthenosphere (Figure 15) [e.g., *Houseman and Molnar*, 2001]. We can also imagine that mantle lithosphere may have been removed more than once; following removal, subsequent thickening, again due to horizontal shortening, could have made mantle lithosphere unstable again. The volcanic history of Tibet suggests that melting of mantle lithosphere occurred not just in the past 15 Ma but much earlier in some areas [*Arnaud et al.*, 1992; *Chung et al.*, 1998, 2003, 2005; *Ding et al.* 2003; *Guo et al.*, 2006; *Hou et al.*, 2004; *Liu et al.*, 2008; *Miller et al.*, 1999; *Roger et al.*, 2000; *Wang et al.*, 2005; *Q. Wang et al.*, 2008; *Yin and Harrison*, 2000], consistent with earlier periods of removal.

[91] The absence of significant crustal thickening within central Iran, but with crustal thickening on the margins, suggests (at least to us) that thin lithosphere beneath central Iran may have formed because lithosphere beneath the surrounding ranges became unstable as it thickened; then

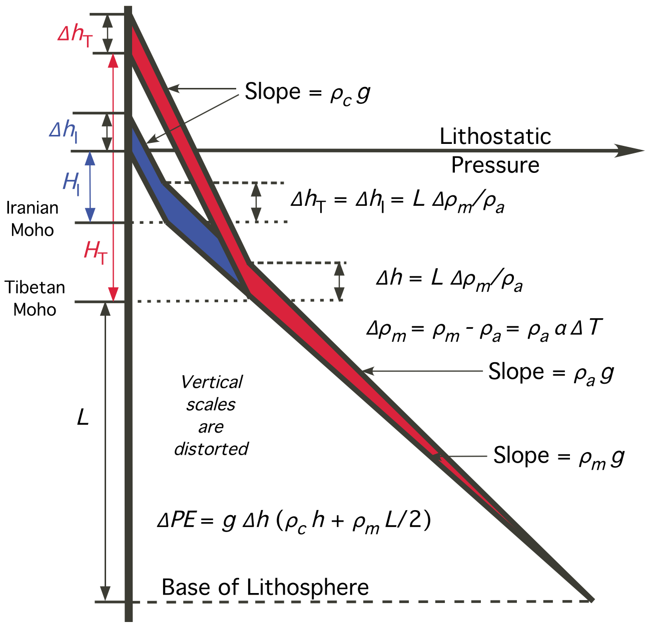


Figure 16. Profiles of lithostatic pressure versus depth for regions with different initial elevations and different crustal thicknesses, before and after mantle lithosphere is removed. (Scales are distorted so that differences between profiles are apparent; thus, differences in surface elevations of order 1–5 km ought not be compared directly with crustal thicknesses of 35–70 km.) Pressure increases with depth with a gradient proportional to the product of density and gravity. For Airy isostatic equilibrium, the excess crustal thickness at depth is proportional to the elevation of the surface. If mantle lithosphere of mean density ρ_m is removed and replaced by asthenospheric material of mean density ρ_a, the crustal thickness remains unchanged, but the surface will rise because of the addition of buoyant material, and the profile of lithostatic pressure versus depth will shift. The pressure at any depth below sea level beneath a high surface is greater than that beneath a lower surface. The integrated difference in profiles of lithostatic pressure, from the surface to the depth of compensation, here taken to be the base of the lithosphere, gives the potential energy per unit area and equivalently the force per unit length that the two columns of mass apply to one another [e.g., *Molnar and Lyon-Caen*, 1988]. Thus, thickening of crust or removal of mantle lithosphere, both of which lead to a rise in surface elevation, creates potential energy per unit area and leads to an increase in the force per unit length that the two columns apply to one another. Moreover, the increase in potential energy per unit area is greater if mantle lithosphere is removed from beneath thicker than thinner crust.

sinking of that thickened mantle lithosphere drew at least the deeper part of the mantle lithosphere beneath central Iran into downwelling sheets or plumes (Figure 15). We do not know when (or if) mantle lithosphere was removed from central Iran, but a guess might be ~10–12 Ma, when deformation in the Alborz and Kopet Dagh seems to have accelerated. Compared with Tibet, however, such an event (if it occurred) manifested itself much more subtly than it seems to have beneath Tibet. For instance, no marked decrease in the convergence rate between Arabia and Eurasia has occurred since the slowing of convergence at the time of collision (Figure 4).

[92] One might ask why removal of mantle lithosphere beneath both plateaus has not manifested itself similarly. First, clearly, the Iranian Plateau has not grown to the lateral dimensions or to the height of Tibet. More interesting and important, however, is the fact that crust within the Iranian Plateau has not yet thickened much, if at all. Removal of mantle lithosphere and replacement with hotter asthenosphere requires that the remaining crust rise in order to maintain isostatic equilibrium. In principle, removal of the same amount of mantle lithosphere from beneath two regions of different crustal thickness will lead to identical amounts of surface uplift. Yet when such mantle lithosphere is removed, the change in potential energy per unit area is greater beneath the region with the thicker crust (Figure 16) [e.g., *Molnar and Stock*, 2009]. For regions in isostatic equilibrium the change in potential energy per unit area turns out to be equal to the change in horizontal force per unit length that the elevated region applies to its surroundings [e.g., *Molnar and Lyon-Caen*, 1988]. Thus, removal of mantle lithosphere from beneath central Iran should have increased the force per unit length that Iran applies to Arabia or to Eurasia, but that increase should have been smaller than the change in the force per unit length that Tibet applies to India or Eurasia if the same amount of mantle lithosphere were removed from beneath it. We suspect that removal of mantle lithosphere from beneath Tibet altered the balance of forces (per unit length) beneath Tibet sufficiently to cause deformation in Tibet to change from dominantly north-south crustal shortening to the present-day east-west extension. The increased pressure difference would cause Tibet to flow radially outward onto the Indian plate, whether that radial translation is attributed to change in force per init length applied to a brittle substance, like a Coulomb solid [e.g., *Dahlen and Suppe*, 1988], or to a viscous substance acting like a gravity current [e.g., *Copley and McKenzie*, 2007]. The removal of the same amount of mantle lithosphere from beneath Iran, however, might merely lead to a redistribution of crustal shortening and thickening beneath the Iranian Plateau and, in particular, beneath its margins.

[93] The change in potential energy per unit area, or force per unit length, due to removal of mantle lithosphere is given by [e.g., *Molnar and Stock*, 2009]

$$\Delta PE = \Delta E g(\rho_c h + \rho_m L/2), \tag{1}$$

where ΔE is the amount of surface uplift, g is gravity, h is the crustal thickness, and L is the initial thickness of mantle lithosphere. With $g = 9.8$ m/s^2, $\rho_c = 2.8 \times 10^3$ kg m^{-3}, and $\rho_m = 3.3 \times 10^3$ kg m^{-3}, if $\Delta h = 1.5$ km and $L = 100$ km, then for $h = 35$ or 70 km, $\Delta PE = 3.9 \times 10^{12}$ or 5.3×10^{12} N/m, respectively. The former is comparable with typical values assigned to "ridge push," the force per unit length that thin lithosphere at mid-ocean ridges applies to old, thick oceanic lithosphere [e.g., *Chapple and Tullis*, 1977; *Forsyth and*

Uyeda, 1975; *Frank*, 1972; *McKenzie*, 1972b], and the latter is somewhat greater. Obviously, these estimates of ΔPE or force per unit length are uncertain because of uncertainties in the various parameters, but changes in potential energy and the force balance associated with removal of mantle lithosphere will be greater for thick than for thin crust by an amount that varies linearly with its thickness. Suppose these arguments are correct: that first, mantle lithosphere was removed from both plateaus and second, thicker crust beneath Tibet sufficed to cause a much different response to the Tibetan Plateau than to the Iranian Plateau. Then, the arguably small difference in the changes in force per unit length, $\sim 1 \times 10^{12}$ N/m, which is only a fraction of "ridge push," suggests that deformation in continental regions is in a delicate state of balance, where such changes can have markedly different responses in different settings.

[94] If, in accounting for isostatic compensation of the elevation of the Iranian Plateau, one allows for modest crustal thickening beneath the Iranian Plateau, of, say, 6 km, which is permissible given our ignorance of both the present-day crustal thickness and its initial value, the change in potential energy per unit area for that region would be yet smaller. For this case, ~1000 m of the present-day mean elevation of ~1500 m would be due to crustal thickening, and only 500 m would be due to removal of mantle lithosphere. From (1), the change in potential energy per unit area would be proportionally smaller, or only 1.3×10^{12} N/m, and the difference between that change and the one for Tibet would be $\sim 4 \times 10^{12}$ N/m, more than the effect of "ridge push." Given our ignorance of all of the requisite quantities, we cannot determine the changes in potential energy per unit length for Tibet and Iran well enough to quantify their differences, but surely, that for Tibet is larger and quite likely much larger.

5.3. Rigid Blocks in a Deforming Continuum

[95] The word "block" arises often in discussions of large-scale tectonics, but usages differ with context and often with some confusion. On the one hand, essentially rigid blocks lie within deforming Asia, with the Lut Block in Iran and the Tarim Basin in China standing out as prominent examples. Both seem to have separated from Gondwana as parts of essentially rigid plates, when accreted to Eurasia they remained intact, and they undergo little deformation today. Much of the accretion of material that now comprises both Iran and its surroundings and eastern Asia is also often described in terms of blocks, though unlike the Lut Block or Tarim, most such regions have undergone deformation. In this context, the word "block" implies an entity with the same stratigraphy without regard to its strength. In recent terminology, however, the word is commonly used to define regions undergoing strain that is too small to be measured.

[96] The rapid accumulation of GPS data in and around Tibet has led to a number of attempts to subdivide that region into such blocks, regions undergoing negligible strain, except on their margins where elastic strain accumulates between earthquakes. Whereas *Calais et al.* [2006] and *Shen et al.* [2005] found little evidence for rigid or elastic blocks within Tibet, *Meade* [2007] and *Thatcher* [2007] argued that a division of Tibet into elastic blocks can describe well the velocity field of Tibet, regardless of the facts that their boundaries of blocks differ from one another [*Flesch and Bendick*, 2008] and that the slip rates that they calculated for some major faults do not match those measured [e.g., *Zhang et al.*, 2007]. Thus, the issue of whether blocks describe well the wide deforming Tibet remains controversial.

[97] Attempts to subdivide Iran into blocks have not progressed to the level of disagreement that prevails for Tibet, but GPS data do seem to be sufficient to define one region of central Iran where deformation is slow and where seismicity is low [*Vernant et al.*, 2004]. Given the absence of unusually thick crust and the presence of apparently hot uppermost mantle beneath the Iranian Plateau, the GPS measurements present a peculiarity that might seem puzzling. Much of central Iran seems to behave as a rigid block, for distances between GPS control points in this area show no resolvable changes, and hence, the region undergoes no measurable strain; yet the presence of a rigid block might seem unlikely for a region apparently underlain by a weak uppermost mantle. The rate of deformation of a region, however, scales with the ratio of the deviatoric stress acting on it to some measure of its average viscosity. Although a weak region, one with low viscosity, will obviously deform more rapidly than a stronger one with the same imposed stress, if deviatoric stresses (or forces per unit length) are small, rates of deformation can also be small. We infer this to be the case in central Iran. As such, it is a reminder that even if parts of Tibet do prove to be amenable to treatments in terms of blocks that are not deforming, such blocks need not be strong, let alone rigid.

5.4. Crustal Budgets, Transformation to Eclogite, and Channel Flow

[98] Finally, we note, without discussing either in detail, two controversies surrounding the structure of the Himalaya. They might also apply to the Zagros.

[99] First, estimates of budgets of crust in eastern Asia leave open the possibility that crust has not been conserved during India's collision and penetration into Eurasia. *England and Houseman* [1986] showed that if all present-day elevations were compensated isostatically by a crustal root that formed since India collided with Eurasia, crustal thickening could account for present-day elevations if the collision occurred near 45 Ma and if no material were displaced eastward out of India's northward path into Eurasia. We now know that some isostatic compensation occurs because of hot upper mantle, we suspect that some regions were high before collision, and both very long baseline interferometry and GPS data show that southern China moves eastward at approximately 10 mm/yr [e.g., *Heki et al.*, 1995; *Molnar and Gipson*, 1996; *Simons et al.*, 2007]. Thus, some additional process must account for part of the crustal budget involved in India's penetration into Eurasia. Two possibilities exist: lateral transport of material [e.g., *Le*

Pichon et al., 1992; *Molnar and Tapponnier*, 1975; *Peltzer and Tapponnier*, 1988; *Tapponnier et al.*, 1982, 1986] (called "extrusion" or "tectonic escape" by some) and conversion of subducted crust to eclogite [e.g., *Le Pichon et al.*, 1992].

[100] Much recent attention has been given to the possibility that lower crust beneath the Himalaya has converted to eclogite facies [*Cattin et al.*, 2001; *Henry et al.*, 1997; *Hetényi et al.*, 2007; *Le Pichon et al.*, 1997; *Monsalve et al.*, 2008; *Nábělek et al.*, 2009; *Sapin and Hirn*, 1997; *Schulte-Pelkum et al.*, 2005; *Wittlinger et al.*, 2009]. If the lower crust in such regions has transformed to eclogite, it should have become dense, and seismic wave speeds should resemble those of mantle material. Thus, eclogitization provides a mechanism for hiding subducted crust. Similarly, plate reconstructions call for 500–800 km of convergence between Arabia and Eurasia since collision (Figure 4), but as discussed in section 3.1, crustal shortening within the Zagros seems capable of accounting for less than a third of this amount. Eclogitization, if it occurred beneath the Zagros or the Iranian Plateau, might hide crustal material there and lead to an underestimate of the extent to which the Arabian platform has underthrust the southwestern edge of the Iranian Plateau.

[101] Given the uncertainties in estimating the amount of continental crust underthrust in collision zones, especially for the Himalaya, it is perhaps risky to draw any conclusion from the difference between the budgets for the Himalaya and the Zagros, but let us note another possibility: that early in the collision process, much material was transported laterally out the path of the converging continents, presumably at a higher rate than southern China currently moves eastward relative to Eurasia. Some have argued that material that now comprises much of Indochina lay in southern Tibet, and shortly after India collided with Eurasia, it moved eastward out of India's path [e.g., *Akciz et al.*, 2008; *Peltzer and Tapponnier*, 1988; *Replumaz and Tapponnier*, 2003; *Royden et al.*, 2008; *Tapponnier et al.*, 1982, 1986, 1990, 2001]. Rapid lateral transport of material occurs in this region now, as shown best by GPS measurements (Figure 6) [e.g., *Gan et al.*, 2007; *King et al.*, 1997; *Zhang et al.*, 2004]. Rapid present-day westward transport of material also occurs across Turkey (Figure 6), but the total offset of only ~85 km for the North Anatolian fault [e.g., *Armijo et al.*, 1999; *Hubert-Ferrari et al.*, 2002; *Şengör*, 1979b; *Şengör et al.*, 2005], which accommodates virtually all of that movement relative to Eurasia, makes large magnitudes of lateral transport in this region appear unlikely. It may seem more likely that crustal material would flow around the southeast end of the Zagros, where a relatively free boundary along the Makran coast might allow southward flow of material analogous to that at the eastern end of the Himalaya, but at present, GPS velocities give no indication of such movement (Figure 6). To our knowledge, geologic evidence for large horizontal movement of material in this region has not been found.

[102] A second controversy surrounding the Himalaya is the possible role played by flow of lower crust, "channel flow," such that rock cropping out in the Greater Himalaya today has flowed southward many tens to perhaps hundreds of kilometers with respect to both India and the upper crust of southern Tibet [e.g., *Beaumont et al.*, 2001, 2004; *Grujic et al.*, 1996, 2002; *Hodges et al.*, 2001; *Hollister and Grujic*, 2006; *Jamieson et al.*, 2004; *Law et al.*, 2006]. This view is controversial, and many reject it [e.g., *Bollinger et al.*, 2006; *Kirstein et al.*, 2006; *Kohn*, 2008; *Murphy*, 2007; *Robinson et al.*, 2006; *Webb et al.*, 2007]. The profiles of receiver functions across the Zagros (Figure 11) do not lend themselves to an interpretation in terms of channel flow. The dipping interfaces and zones of low-speed material beneath them, especially that on the northwestern profile, pass from very shallow depths to the Moho without obvious distortion. If flow of lower crust from beneath central Iran to beneath the Zagros had occurred in a channel within the lower crust, these interfaces and low-speed zones should have been distorted by that flow and not show a linear cross section. The absence of evidence for such distortion, and hence for such a process, in the Zagros might weigh against the general concept of channel flow in such a setting, though obviously, these data do not bear on flow parallel to the Zagros and perpendicular to the cross sections. Alternatively, many who advocate such channel flow see the need for rapid erosion of the Greater Himalaya as a necessary participant in such flow [e.g., *Beaumont et al.*, 2001, 2004; *Hodges et al.*, 2001]. Thus, its absence in the Zagros might merely result from the absence of rapid erosion and might simply be the result of the Zagros being at a stage too early for such a process to develop or the absence of climatic conditions that lead to rapid erosion.

6. CONCLUSIONS

[103] The principal differences between the Zagros and Himalaya and between the Iranian and Tibetan plateaus seem to reflect different degrees of development of a similar set of processes, which result from a greater penetration of India than Arabia into Eurasia (Figures 3 and 4). These include both different amounts of underthrusting of India as compared to Arabia beneath Eurasia and different degrees of crustal thickening beneath Tibet as compared to Iran.

[104] The construction of the Zagros has occurred in large part by folding and thickening of the sedimentary cover on the edge of the Arabian platform as Arabia plunged beneath central Iran perhaps as much as 100 km but apparently less (Figures 5 and 11). The Himalaya, however, has been built largely by the stacking of thrust sheets consisting of slices of India's crystalline basement as deeper portions of India's lithosphere slid hundreds of kilometers beneath southern Eurasia (Figures 5, 8, and 11). The folding of sedimentary rock that built the Zagros also occurred within the Himalaya but ended before 20 Ma and perhaps much earlier. We see the current seismicity, fault plane solutions of earthquakes, and distribution of deformation within the Zagros measured with GPS (Figure 8) to be consistent with the possibility that a major intracontinental thrust fault, similar to the Main

Central Thrust of the Himalaya, is in the process of forming beneath the Zagros as a seismically active reverse fault in the upper crust but as a nearly horizontal ductile shear zone within deforming middle to lower crust (Figure 14).

[105] Crust has thickened beneath virtually all of Tibet and has started to thicken beneath the mountain belts flanking the Iranian Plateau (Figure 12). Both plateaus undergo deformation that includes strike slip, but whereas Tibet undergoes normal faulting and crustal thinning, the Iranian Plateau undergoes minor crustal shortening and thickening (Figure 11). As high plateaus cannot be built by normal faulting and crustal thinning, at least not without positing a conspiracy of processes that seem unlikely to occur together, some change in the balance of forces must have occurred beneath Tibet. Because both plateaus seem to be underlain, at least in part, by relatively warm upper mantle, as implied by low *P* and *S* wave speeds and high attenuation of seismic waves (Figure 13a), we presume that some mantle lithosphere has been removed from beneath each plateau (Figure 15), but readers should realize that this interpretation is speculative. In any case, the warmer material beneath the plateaus than beneath normal mantle lithosphere contributes to the isostatic balance of the plateaus. Because of Tibet's greater crustal thickness, removal of mantle lithosphere beneath it should lead to greater change in the balance of forces between the plateau and its surroundings (Figure 16). The greater change in convergence rate between India and Eurasia than between Arabia and Eurasia accords with a bigger change in the balance of forces for Tibet than Iran.

[106] Whereas Tibet undergoes widespread deformation that includes crustal thinning, the Iranian Plateau in central Iran seems not to be deforming by internal strain, despite its relatively thin mantle lithosphere. This difference in ongoing deformation suggests that Tibet stands higher than it must in order to balance the forces (per unit length) that it and its surroundings apply to one another, but the Iranian Plateau is more closely in balance with the forces it applies to its surroundings (Figure 16). The absence of ongoing deformation in central Iran, therefore, derives not from strength but from a state of low deviatoric stress. The same state may have characterized Tibet before ~15 Ma when a change occurred, which we ascribe to removal of mantle lithosphere (Figure 15), that led to widespread deformation on the surroundings of the Tibetan Plateau.

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