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Cenozoic tectonic evolution of Asia: A preliminary synthesis

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ABSTRACT

Asia has been a major testing ground for various competing models of continental deformation due to its relatively well-understood plate boundary conditions in the Cenozoic, exceptional exposure of active structures, and strain distribution, and widespread syn-collisional igneous activity as a proxy for the thermal state of the mantle and crust. Two Cenozoic orogens dominate the continent: the Himalayan–Tibetan orogen in the east induced by the India–Asia collision and the Turkish–Iranian–Caucasus orogen in the west induced by the Arabia–Asia collision. The development of the two orogens was accomplished by shortening in the early stage followed by strike-slip faulting and extension in the late stage. In the Himalayan–Tibetan orogen, shortening across two discrete thrust belts at 55–30 Ma in southern and northern Tibet created a large intracontinental basin (the Paleo-Qaidam basin) in between. Subsequent crustal thickening and a possible thermal event in the mantle (e.g., convective removal of central Tibetan mantle lithosphere) may have raised the elevation of this early intra-plateau basin up to ~ 2–3 km to its current height. Collision between India and Asia also caused lateral extrusion of southeast Asia between 32 Ma and 17 Ma. The latest stage of the India–Asia collision was expressed by north-trending rifting and the development of trench-facing V-shaped conjugate strike-slip faults in central Mongolia, central Tibet, eastern Afghanistan and southeast Asia. In the Turkish–Iranian–Caucasus orogen, early crustal thickening in the orogenic interior began at or prior to 30–20 Ma. This style of deformation was replaced by strike-slip faulting at ~15–5 Ma associated with further northward penetration of Arabia into Asia, westward extrusion of the Anatolia/Turkey block, and rapid extension across the Sea of Crete and Sea of Aegean. The late stage extension in both orogens was locally related to extensional core-complex development. The continental-margin extension of east Asia was developed in two stages: initially in a widely distributed zone that has an east-west width of 500–800 km during 65–35 Ma, which was followed by localized extension and opening of back-arc basins associated with the development of spreading centers at 32–17 Ma (e.g., Japan Sea or East Korea Sea, Bohai Bay, and South China Sea). Opening of the back-arc basins could be induced by (1) rapid eastward migration of the western Pacific trench system or (2) oblique subduction of Pacific plate beneath Asia that had produced a series of en echelon right-slip primary shear zones linking with back-arc spreading centers oriented obliquely to the strike of the nearby trench. Since ~15 Ma, the eastern margin of Asia became contractional in the east–west direction, as indicated by the collapse of back-arc basins in the western Pacific and the development of fold-thrust belts along the eastern continental margin. Coeval with the contraction is widespread east–west extension in Siberia, North China, and the Tibetan plateau. The above observations can be explained by a change in boundary condition along the eastern margin of Asia that allowed the thickened Asian continent to spread eastward, causing east-west extension in its trailing edge and east-west compression in its leading edge. In west Asia, continental-margin extension started at about 25–20 Ma in the Aegean and Cretan regions, which was associated with a rapid southward retreat of the Hellenic arc. The complex evolution of Cenozoic deformation in Asia may be explained by a combined effect of temporal changes in plate boundary conditions, thermal evolution of the upper mantle perturbed by collisional tectonics, and the built-up of gravitational energy through crustal thickening and thermal heating. Although the past research in Asia has treated the India–Asia and Arabia–Asia convergence as separate collisional processes, their interaction may have controlled the far-field Cenozoic deformation in Asia. The most pronounced result of this interaction is the creation of a northeast-trending 300–400-km wide and >1500-km long zone of northwest-striking right-slip faults, which extends from the Zagros thrust belt in the south to western Mongolia in the north and links with the active Tian Shan and Altai Shan intracontinental orogens. Cenozoic deformation and coeval igneous activity spatially overlap with one another in the Himalayan–Tibetan and Turkish–Iranian–Caucasus orogens. A large Cenozoic magmatic gap exists between Tibet in the south and Mongolia in the north where Cenozoic

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deformation has not been associated with any coeval igneous activity. Finally, Cenozoic igneous activity is always associated with Jurassic–Cretaceous magmatic arcs, suggesting a causal relationship between the early arc magmatism and later syn-collisional magmatism.

1. Introduction

No place on Earth displays more diverse and more complex patterns of active deformation in Asia (Fig. 1). Understanding the geometry and kinematics of this tectonic system has profound implications for deciphering the mechanisms of continental deformation, the feedback processes between lithospheric deformation and atmospheric circulation and the crustal evolution of other planets within the solar system (McKenzie, 1972, 1978; Molnar and Tapponnier, 1975; Tapponnier et al., 1982, 2001; England and Houseman, 1986; Davy and Cobbold, 1988; Cobbold and Davy, 1988; Harrison et al., 1992; Solomon et al., 1992; Molnar et al., 1993; Yin and Harrison, 2000; Yang and Liu, 2002; Liu and Yang, 2003; Yin, 2006; Basilevsky and Head, 2007). Although it is well known that the Cenozoic

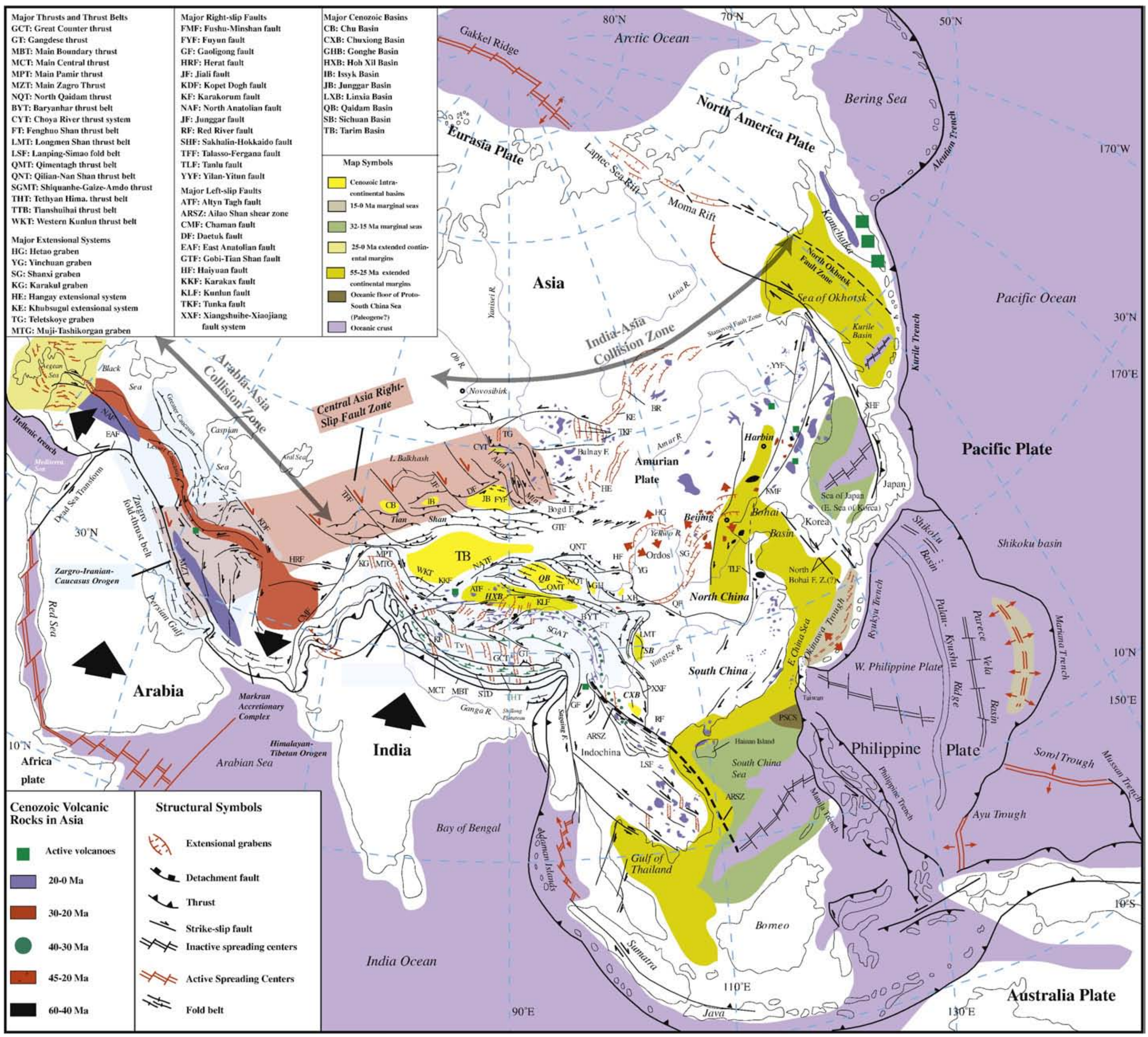

Fig. 1. Cenozoic structures and distribution of volcanic rocks. Ages of igneous rocks are from the following sources. Northern China, Wang (1982), Ye et al. (1987), Zhou et al. (1988), R.X. Liu et al. (1992), J.Q. Liu et al. (2001); Far East Russia, Okamura et al. (1998); Baikal rift system, Rasskazov (1994); Mongolia, Whitfotd-Stark (1987), Traynor and Sladen (1995), Barry and Kent (1998), Cunningham (2005); central and east Tibet, Pan et al. (1990), Yu (1991), Turner et al. (1993), Deng (1998), Chung et al. (1998, 2005), Wang et al. (2001); Pamirs and the Karakorum Mountains, Ratschbacher et al., 1994; Murphy et al., 2002; Robinson et al., 2004; Murphy and Copeland, 2005; Robinson et al., 2007; southeast China, R.X. Liu et al. (1992), Ho et al. (2000, 2003); southern Tibet, Yin et al. (1994, 1999a), Miller et al. (1999), Williams et al. (2001); Indochina, Lee et al. (1998); Iran, Berberian and Berberian (1981). The age of the Cenozoic volcanic rocks in eastern Asia is divided into the following periods: 60–40 Ma, 40–30 Ma, 30–20 Ma, 10–8 Ma, and 8 Ma to present. In Iran, Cenozoic volcanism is only divided into Eocene–Early Oligocene episode and Late Oligocene–Miocene episode.

deformation of Asia was a combined result of continental collision and oceanic subduction (McKenzie, 1972; Molnar and Tapponnier, 1975), how the two classes of plate boundary processes interact with one another in creating widespread intracontinental deformation remains debated [see discussions in Davy and Cobbold (1988) and Fournier et al. (2004)]. One thought is that the collisional tectonics drives the opening of the marginal seas, forcing the subduction zones to retreat (e.g., Tapponnier et al., 1982; Jolivet et al., 1990, 1992, 1994). In this view, the subduction plate boundaries are passive and may migrate due to large-scale deformation of Asian continent (Peltzer and Tapponnier, 1988). Alternatively, the subduction of the oceanic plates and related opening of marginal sea basins were independent of collisional tectonics and exerted a strong role in controlling the timing, location and style of intracontinental deformation in Asia (e.g., LePichon and Angelier, 1979; Northrup et al., 1995: Yin, 2000; Chough et al., 2000; Lewis et al., 2002; Hall, 2002; Royden et al., 2008). In addition, Cenozoic deformation of Asia may not only be a product of plate boundary processes as discussed above, but also a result of changes in the thermal state of the lithosphere induced by crustal and mantle processes (e.g., England and Houseman, 1989; England et al., 1992; Molnar et al., 1993; Harrison et al., 1998; Tapponnier et al., 2001) and erosion (e.g., Beaumont et al., 2001). A proxy for the thermal conditions in the mantle is the occurrence of widespread Cenozoic igneous activities across Asia (e.g., Deng, 1998; Flower et al., 1998; Wang et al., 2001; Liu et al., 2001; Ding et al., 2003; Ho et al., 2003; Chung et al., 2005; Pan et al., 2006; Mo et al., 2007). One way to determine the relationship between lithospheric deformation and thermal state of the mantle and crust is to examine the spatial and temporal correlation of structural development vs. igneous activity. Finally, it remains unknown how the India–Asia and Arabia–Asia collision has interacted with one another to shape the overall deformation history of Asia.

The main purpose of this paper is to provide a brief summary of the style, pattern, and timing of Cenozoic deformation across major tectonic domains in Asia. Although the primary emphasis is on the India–Asia collision in east Asia, I will also touch briefly the deformation history related to the Arabia–Asia collision in west Asia and its possible effect on the development of Cenozoic fault systems in central Asia. The synthesis below is based on a series of paleo-tectonic maps compiled from existing literature and my own interpretations of the existing data. They formed the basis for a sequential palinspastic reconstruction of Cenozoic deformation of whole Asia that in turn allows an informative discussion on the dynamic causes of continental deformation.

2. Cenozoic deformation

Asia constitutes two broad Cenozoic deformation zones: the India–Asia collision zone in the east and the Arabia–Asia collision zone in the west (Fig. 1). The boundary between the two is diffuse and lies generally along longitude 61–64°E through the middle part of Afghanistan. West of this boundary is a north- to northwest-trending right-slip fault system in eastern Iran and westernmost Afghanistan, accommodating northward penetration of Arabia into Asia. East of the boundary is the left-slip Chaman fault system in easternmost Afghanistan, accommodating northward penetration of India into Asia (Fig. 1). The India–Asia collision zone consists of the following major tectonic domains: (1) the Himalayan orogen, (2) the Tibetan Plateau, (3) the southeast Asia extrusion system, (4) the central Asia deformation domain stretching from the Tian Shan in the south to the Baikal rift zone in the north, (5) the North China deformation domain, and (6) the eastern Asia margin deformation domain extending from the eastern continental margin of Asia to the western Pacific trench system in the east-west direction and from the Sea of Okhotsk in the north to South China Sea in the south in the north-south direction. The Arabia–Asia collision zone consists of (1) the Turkish–Iranian–Caucasus orogen in front of the Arabia indenter, (2) the Turkey (also known as the Anatolian) extrusion system, and (3) the Aegean–Cretan extensional system. In addition to deformation related to collisional tectonics, the northeastern edge of Asia is also affected by the propagation of the Arctic mid-ocean ridge onto the continent, which is expressed by the development of the Laptev–Moma rift zone (Fig. 1).

2.1. Himalayan orogen

The Himalayan orogen is bounded by a frontal thrust zone along its southern range front in the south and the Indus–Tsangpo suture zone in the north (Fig. 1) (Yin, 2006). The suture zone is strongly modified by Cenozoic deformation and marked by a north-directed thrust (e.g., Ratschbacher et al., 1994; Yin et al., 1994, 1999a; Murphy et al., 2002; DiPietro et al., 2008). The oldest Cenozoic crustal shortening occurred in the Tethyan Himalaya in the Middle and Late Eocene (50–40 Ma) south of the Indus–Tsangpo suture (e.g., Ratschbacher et al., 1994; Wiesmayr and Grasemann, 2002; DeCelles et al., 2004; Yin, 2006). Cenozoic deformation in the Himalaya has been divided into an early phase between the Middle Eocene to Late Oligocene and a later phase between the Early Miocene and the present (Le Fort, 1996; Hodges, 2000). The evidence for the early event is best preserved in the sedimentary record of the Himalayan foreland basin (DeCelles et al., 1998a,b, 2004; Najman, 2006), the cooling history of the Tethyan Himalayan thrust belt (Ratschbacher et al., 1994; Wiesmayr and Grasemann, 2002), and 44-Ma plutons cutting highly folded strata of the Tethyan Himalayan Sequence (Aikman et al., 2008). The older event is also recorded in the high-grade core of the Himalayan orogen as indicated by Oligocene $^{40}Ar/^{39}Ar$ mica and hornblende cooling ages and Eocene–Oligocene monazite-inclusion ages in garnets (Vannay and Hodges, 1996; Argles et al., 1999; Godin et al., 2001; Catlos et al., 2001; Kohn et al., 2004; Martin et al., 2007). The younger phase of deformation is marked by the Early–Middle Miocene development of the south-directed Main Central Thrust, the north-dipping South Tibet Detachment fault, and the north-directed Greater Counter Thrust (Hubbard and Harrison, 1989; Burchfiel et al., 1992; Yin et al., 1999a; Johnson et al., 2001; Kohn et al., 2004; Webb et al., 2007; Kohn, 2008). The development of these structures was followed by the Late Miocene–Pliocene reactivation of the Main Central Thrust (Harrison et al., 1997; Catlos et al., 2001) or development of a duplex structures causing folding of the MCT (Robinson et al., 2003). The Main Boundary thrust in the central and western Himalaya was initiated at 11–5 Ma (Megis et al., 1995; DeCelles et al., 2001).

Estimates of the total amount of Cenozoic shortening across the Himalayan orogen vary along strike. In the west, the magnitude of shortening is in the range of ~200 km or less (DiPietro and Pogue, 2004). In the central Himalaya (i.e., Nepal Himalaya and south-central Tibet), the minimum amount of shortening is on the order of 500–760 km and could reach to 900–1000 km if the shortening in the Tethyan Himalaya Sequence is included (DeCelles et al., 2001; Murphy and Yin, 2003; Robinson et al., 2003, 2006; Robinson, 2008). In the eastern Himalaya, the estimated minimum amount of shortening varies from 350 km to 500 km (Yin et al., 2006; McQuarrie et al., 2008).

2.2. Tibetan plateau

Cenozoic tectonics of Asia is most dramatically expressed by the development of the Tibetan plateau resulting from the India–Asia collision (Dewey and Burke, 1973; Chang and Zheng, 1973; Le Fort, 1975; Allègre et al., 1984; Dewey et al., 1988; Le Fort, 1996; Yin and Harrison, 2000; DeCelles et al., 2002; Yin, 2006) (Fig. 1). The exact onset age of the collision remains debated, with estimates mostly ranging from 65 Ma (e.g., Yin and Harrison, 2000) to ~50 Ma (e.g., Rowley, 1996, 1998; Najman et al., 2001; Zhu et al., 2005). Aitchison et al. (2007) recently suggested that the India–Asia collision did not start until about 35 Ma, a proposal not consistent with most of the geologic observations (see discussion by Garzanti, 2008). Regardless of this uncertainty, a wide range of geologic observations indicate that the southern Tibetan plateau (i.e., the Lhasa block) has been largely devoid of Cenozoic crustal shortening, with its current elevation mostly established in the late Cretaceous through the development of arc-related or collision-related thrust-belt development (England and Searle, 1986; Yin et al., 1994; Murphy et al., 1997; Kapp et al., 2005, 2007; DeCelles et al., 2007a; Volkmer et al., 2007; Leier et al., 2007; Pullen et al., 2008). As a result, my review on the Cenozoic deformation of Tibet below focuses on the geology of northern and central Tibet.

In contrast to southern Tibet, central and northern Tibet experienced significant crustal shortening expressed as localized thrust belts within 10 m.y. from the onset of India–Asia collision. The Fenghuo Shan–Nangqian thrust belt was developed in the Eocene and Early Oligocene (55–30 Ma), producing a large foreland basin in the Hoh Xil region (Leeder et al., 1988; Liu and Wang, 2001; Horton et al., 2002; Spurlin et al., 2005; Zhu et al., 2006). Coeval with Fenghuo Shan–Nangqian thrust belt was the development of the Qilian Shan thrust belt that was reactivated from a Paleozoic orogen (e.g., Yin et al., 2007a) and bounds the current northern edge of the Qaidam basin. The Eocene–Early Oligocene shortening event was also associated with significant crustal rotation and possible strike-slip faulting in the northeastern margin of the Tibetan plateau (Dupont-Nivet et al., 2004).

Associated with the Eocene–Oligocene development of the Qilian Shan thrust belt in the north and the Fenghuo Shan–Nangqian thrust belt in the south was the formation of a large intermontane basin, the Paleo-Qaidam basin of Yin et al. (2007b, 2008a,b). It has a north–south width of >800 km in the Eocene and Oligocene (50–25 Ma) and was later partitioned into the Hoh Xil and modern Qaidam basins in the early Miocene at ~20 Ma. The basin partitioning was induced by the formation of the left-slip Kunlun fault and the kinematically linked northwest-striking Qimen Tagh and Baryanhar thrust belts to the north and south (Yin et al., 2007b; Yin et al., 2008b) (Fig. 1). The Kunlun transpressional system was initiated at ~15 Ma or earlier, with a GPS and Quaternary slip rate of 11–16 mm/yr along its central segment (Kidd and Molnar, 1988; van der Woerd et al., 2002; Jolivet et al., 2003; Zhang et al., 2004; Fu and Awata, 2007) and ~5 mm/yr along its eastern segment (Kirby et al., 2007; Harkins and Kirby, 2008). The fault may die out completely before reaching to the north-trending Longmen Shan thrust belt in the east (Kirby et al., 2007; cf. Chen et al., 1994). A prominent Cenozoic cooling event at 30–10 Ma occurred in the eastern Kunlun region (Mock et al., 1999; Jolivet et al., 2001; Wang et al., 2004; Liu et al., 2005; Yuan et al., 2006), coeval with initiation of crustal shortening along the southern margin of the Qaidam basin as recorded in growth-strata relationships (Yin et al., 2008b).

The Cenozoic Altyn Tagh fault system lies along the northern edge of the Tibetan plateau, linking the Qilian Shan thrust belt in the northeast and the western Kunlun thrust belt in the southwest (Fig. 1) (Burchfiel et al., 1989; Tapponnier et al., 1990; Wang, 1997; Meyer et al., 1998). The fault system initiated in the Paleocene to Middle Eocene (60–45 Ma) during the initial stage of India–Asia collision (Bally et al., 1986; Yin et al., 2002, 2008a) has a total slip of ~470 km at its western end (Cowgill et al., 2003), decreasing to ~360 km along its central segment (Ritts and Biffi, 2000; Yang et al., 2001; Gehrels et al., 2003a,b), and finally reaching ~230 km at its eastern end (Yin and Harrison, 2000). The fault system consists of several branches, some of which are no long active. The fault system exhibits complex internal geometry consisting of strike-slip duplexes and thrust-related folds (Cowgill et al., 2000, 2004a,b; Yin et al., 2002; Gold et al., 2006). Although Meriaux et al. (2005) originally determined the Quaternary slip rate on the Altyn Tagh fault to be 25–30 mm/yr, additional dating and reinterpretation of Meriaux et al.'s (2005) field relationships have led to revision of the fault slip rate to ~10 mm/yr (Cowgill, 2007; Zhang et al., 2007). This lower value is consistent with both the GPS-determined slip rate at decadal scale and long-term geologic rate over tens of million years on the fault (Shen et al., 2001; Yin et al., 2002).

The western Kunlun thrust belt at the western end of the Altyn Tagh fault is kinematically linked with the Main Pamirs Thrust to the northwest (Fig. 1) (Pan et al., 1990; Brunel et al., 1994; Cowgill et al., 2003; Robinson et al., 2007). This structure links the north-striking left-slip fault system along the Afghanistan–Pakistan border accommodating northward penetration of India into Asia (Fig. 1) (Lawrence et al., 1981). An important branch of the Altyn Tagh fault is the left-slip Karakax fault south of the western Kunlun thrust belt (Matte et al., 1996), which terminates in the south at the Muji–Kongur Shan extensional system (Robinson et al., 2004, 2007).

The active right-slip Karakorum fault extending over 1000 km links the north-trending Muji–Kongar Shan extensional system in the northwest and the Gurla Mandhata–Pulan extensional system in the southeast (Ratschbacher et al., 1994; Murphy et al., 2002; Murphy and Copeland, 2005; Robinson et al., 2004, 2007; Robinson, 2009). Although the structure is conjugate to the left-slip Altyn fault, its total slip appears to be much smaller than that on the Altyn Tagh fault. Along its central portion, the fault may have moved ~120 km (Searle, 1996; Searle et al., 1998) whereas at its southern end the fault only slipped ~65 km (Murphy et al., 2000, 2002). As the Gurla Mandhata detachment fault and Muji–Kongur Shan extensional system started to develop at 9–10 Ma (Murphy et al., 2002; Robinson et al., 2004, 2007, the Karakorum fault should also have initiated at this time, which is significantly younger than the initiation age of the Altyn Tagh fault at about 55–45 Ma (Bally et al., 1986; Yin et al., 2002; Yin et al., 2008b). Even the Oligo-Miocene initiation of the Karakorum fault as proposed by Valli et al. (2008) would still make the Karakorum fault considerably younger than the age of the Altyn Tagh fault.

The Indus–Tsangpo and Bangong–Nujiang suture zones along the southern edge and inside Tibet were both reactivated by Cenozoic crustal shortening (Yin and Harrison, 2000). Along the Indus–Tsangpo suture zone is marked by the development of the Late Oligocene (27–23 Ma) Gangdese thrust system and the Miocene Renbu–Zedong thrust zone (Yin et al., 1994, 1999a; Harrison et al., 2000). The Oligocene–Miocene Shiquanhe–Gaize–Amdo thrust belt was developed along the Bangong–Nujiang suture zone (Yin and Harrison, 2000; Kapp et al., 2003, 2005).

A major change in Tibetan deformation pattern occurred in the Middle to Late Miocene (i.e., 18–8 Ma) when north–south contraction was replaced by coeval development of conjugate strike-slip faulting and east–west extension (Armijo et al., 1986, 1989; Yin et al., 1994; Coleman and Hodges, 1995; Harrison et al., 1995; Blisniuk et al., 2001; Williams et al., 2001; Taylor et al., 2003; Li and Yin, 2008; Taylor and Yin, 2009). This deformation pattern occurred under a constrictional strain field associated with north–south shortening and east–west extension (Mercier et al., 1987; Yin et al., 1999b; Yin, 2000). North-trending dikes, dike swarms, and normal faults accommodating east–west extension in Tibet began to develop between 18 and 8 Ma (Yin et al., 1994; Coleman and Hodges, 1995; Harrison et al., 1995; Blisniuk et al., 2001; Williams et al., 2001). The northern limit of the east–west extension is defined by the left-slip Kunlun fault (Fig. 1).

Crustal shortening in eastern Tibet is expressed by the development of an Eocene–Oligocene thrust belt extending from the Fenghuo Shan–Nangqian region in the north to the Red River region in the south (i.e., the Simao–Langpin thrust belt of Wang and Burchfiel, 1997; the Gongjo thrust belt of Studnicki-Gizbert et al., 2008; also see J.H. Wang et al., 2001) (Fig. 1). This crustal thickening event was followed by the Late Miocene to Early Pliocene initiation of the Longmen Shan thrust belt that lies along the eastern edge of the Tibetan plateau directly against the Sichuan basin and north-trending left-slip faulting north of the Red River fault zone and west of the Longmen Shan thrust belt "after" the Sichuan basin (Fig. 1) (Burchfiel et al., 1995, 2008a; Ratschbacher et al., 1996; Kirby et al., 2002a,b). Based on growth-strata relationship, Jia et al. (2006) argued that the Longmen Shan thrust belt may have started to develop in the Paleocene. Whether the uplift of the Longmen Shan thrust belt was completely induced by upper crustal shortening or mainly by lower crustal flow that has inflated the crust remains debated (Burchfiel et al., 2008a; Hubbard and Shaw, 2009).

2.3. Southeast Asia extrusion system

The Southeast Asia extrusion system is bounded by the left-slip Ailao Shan–Red River shear zone in the northeast and the right-slip Gaoligong shear zone and the Wang Cao and Three Pagodas faults of its southern extensions in the southwest (Lacassin et al., 1997; Morley, 2001; Morley et al., 2001; Kornsawan and Morley, 2002; Akciz et al., 2008) (Fig. 1). The Ailao Shan–Red River shear zone and the conjugate right-slip Wang Cao and Three Pagodas faults were active between 32 and 17 Ma (Leloup et al., 1995, 2001; Harrison et al., 1996). Motion on the Red River fault reversed to right-slip in the Late Miocene and Early Pliocene (Leloup et al., 1995, 2001). North-trending left-slip faults such as the Xiangshuihe–Xiaojiang fault system also began to develop at this time or slightly younger (Ratschbacher et al., 1996; Wang et al., 1998; Wang and Burchfiel, 2000). Small north-trending normal faults are present at the northwestern end of the right-slip Red River fault zone to accommodate clockwise rotation of eastern Tibet around the eastern Himalayan syntaxis (Wang et al., 1998). Late Miocene to present northwest-striking right-slip faults linking with north-striking normal faults are also common in Vietnam south of the Red River fault (Rangin et al., 1995). The occurrence of extrusion tectonics in southeast Asia could either be the cause of marginal-sea opening (i.e., the South China Sea, Tapponnier et al., 1982; Peltzer and Tapponnier, 1988) or a result of trench retreat causing opening of the marginal seas and eastward extrusion of continental blocks from the collision zone (e.g., Royden et al., 2008).

2.4. Central Asia deformation domain

Cenozoic deformation of central Asia represents the far-field effects of the India–Asia collision (Molnar and Tapponnier, 1975; Tapponnier and Molnar, 1979).

The central Asia deformation domain extends from the northern edge of the Tibetan plateau (i.e., the Altyn Tagh fault) to the northern tip of the Baikal rift system. The deformation domain is characterized by large intracontinental basins including the Tarim and Junggar basins, widely spaced intracontinental orogens such as the Tian Shan and Altai Shan extending over 2500 km along strike, and intracontinental rift zones exemplified by the Baikal rift system (Fig. 1). The central Asia deformation comprises the northern and southern parts divided along the northern edge of the Altai Shan and having contrasting deformation styles. In the south, the deformation is characterized by a series of northwest-striking right-slip faults such as the Talas–Fergana and Junggar faults originating in the stable Kazakhstan region in the north and terminating in the south at the east-trending fold–thrust belts in the Tian Shan and the Altai Shan (Fig. 1). The formation and linkage between the right-slip faults and the thrust belts are interpreted to have resulted from distributed left-slip shear deformation across a northeast-trending strip in central Asia, and the left-slip shear zone was developed to accommodate northward indentation of India into Asia (e.g., Davy and Cobbold, 1988; Cobbold and Davy, 1988). The inferred left-slip shear deformation may have caused domino-style rotation of blocks about the vertical axes bounded by the right-slip faults (Thomas et al., 1993, 1994, 1996 a,b, 1999; Cobbold et al., 1996; Bourgeois et al., 1997). Although the block-rotation model explains the observed paleomagnetic data, they do not explain well why the northwest-striking right-slip faults die out to the north into the stable interior of Asia. It seems that main role of the right-slip faults was to accommodate differential north-south shortening across the fault. Specifically, the magnitude of the north-south shortening decreases eastward, which could have been related to the clockwise rotation of the Tian Shan (Avouac et al., 1993) or a far-field effect of the Arabia–Asia collision as suggested in this study (see Discussion below).

The inferred left-slip shear zone by Davy and Cobbold (1988) and Cobbold and Davy (1988) extends from the Gulf of Oman in the south to the Sea of Japan in the north. These workers proposed that the shear-zone development may have contributed to the opening of the back-arc basins in the northwestern Pacific (also see Jolivet et al., 1990; Yue and Liou, 1999). The east-trending fold-thrust belts in the Tian Shan are spaced from tens of kilometers to >100 km and bound several large intermountain basins, including the Chu and Issyk–Kul basins in the western Tian Shan and the Turpan basin in the eastern Tian Shan (Fig. 1) (e.g., Windley et al., 1993; Cobbold et al., 1996). In the north, deformation in central Asia is expressed by the east-trending left-slip faults such as the Bulnay and Bogd faults that link with both the discrete rifts in southern Siberia and northwestern Mongolia and distributed normal faults in central and eastern Mongolia (Figs. 1 and 2). The eastern half of the Bogd fault comprises a series of the left-slip Riedel shears that eventually link with the northern tip of the Shanxi rift zone in northern China (Fig. 1). Prior to ~25 Ma, Cenozoic intracontinental deformation in Asia was confined mostly inside Tibet. It was likely that the present Tarim and Junggar basins were linked as one unified basin at this time before the formation of the Tian Shan range, forming a broad topographic depression north of the Tibetan plateau (Yin et al., 2008b). After this time, the northern front of the contractional deformation had jumped across the Tarim block creating the prominent Tian Shan range in the continental interior (Avouac et al., 1993; Burchfiel et al., 1999; Allen et al., 1999). Abdrakhmatov et al. (1996) used the total amount of shortening across the Tian Shan (~200 km) and the GPS determined shortening rate (~11–13 m/year) to suggest that the uplift of the Tian Shan began at ~8 Ma as a result of the Tibetan plateau reaching its maximum elevation (Molnar et al., 1993). This assertion is inconsistent with thermochronologic data and the age of foreland basin sedimentation that indicates the uplift of the Tian Shan began at 24–20 Ma (Hendrix et al., 1994; Yin et al., 1998). It is possible that shortening and uplift of the Tian Shan may have accelerated since 11 Ma as suggested by neotectonic studies and magnetostratigraphic analyses of foreland basin strata (Bullen et al., 2001; Abdrakhmatov et al., 2001; Chen et al., 2002; Heermance et al., 2007).

North of the Tian Shan are the Altai Mountains that straddle across southeast Kazakhstan, southwest Siberia, northwestern China and western Mongolia. Cenozoic deformation of the Chinese Altai started in the early Miocene and is characterized by east-trending thrusts and folds that terminate northwest-striking right-slip Fuyun fault system on the east side and the left-slip Daebut fault system on the west side of the Junggar basin (Fig. 1) (Xinjiang BGMR, 1993). Both the Fuyun and Daebut faults terminate into east-striking thrusts along the north flank of the Tian Shan (Fig. 1). Based on the offset geologic units (Xinjiang BGMR, 1993), slip on the two faults bounding the Junggar Basin appear to be 20–40 km (also see Briggs et al., 2007), which is significantly less than the Cenozoic displacement on the Altyn Tagh fault bounding the Tarim Basin.

In Russia, Cenozoic deformation of the Gorny Altai is best expressed by the development of the east-trending Chuya thrust system (Fig. 1) (Buslov et al., 1999). Detailed geological and morphological mapping show that sedimentation in the basin began in the middle Miocene to early Pleistocene apparently in an extensional graben that was later inverted by a north–south compressional structure since the latest Early Pliocene (Buslov et al., 1999). In south-central Siberia immediately north of the Gorny Altai, Cenozoic deformation is expressed by formation of the north-trending Teletskoye graben at the termination of two conjugate strike-slip faults (Fig. 1). This geometry is very similar to that observed in central Tibet by Taylor et al. (2003) and indicate a constrictional strain field. The initiation age of the Teletskoye graben is

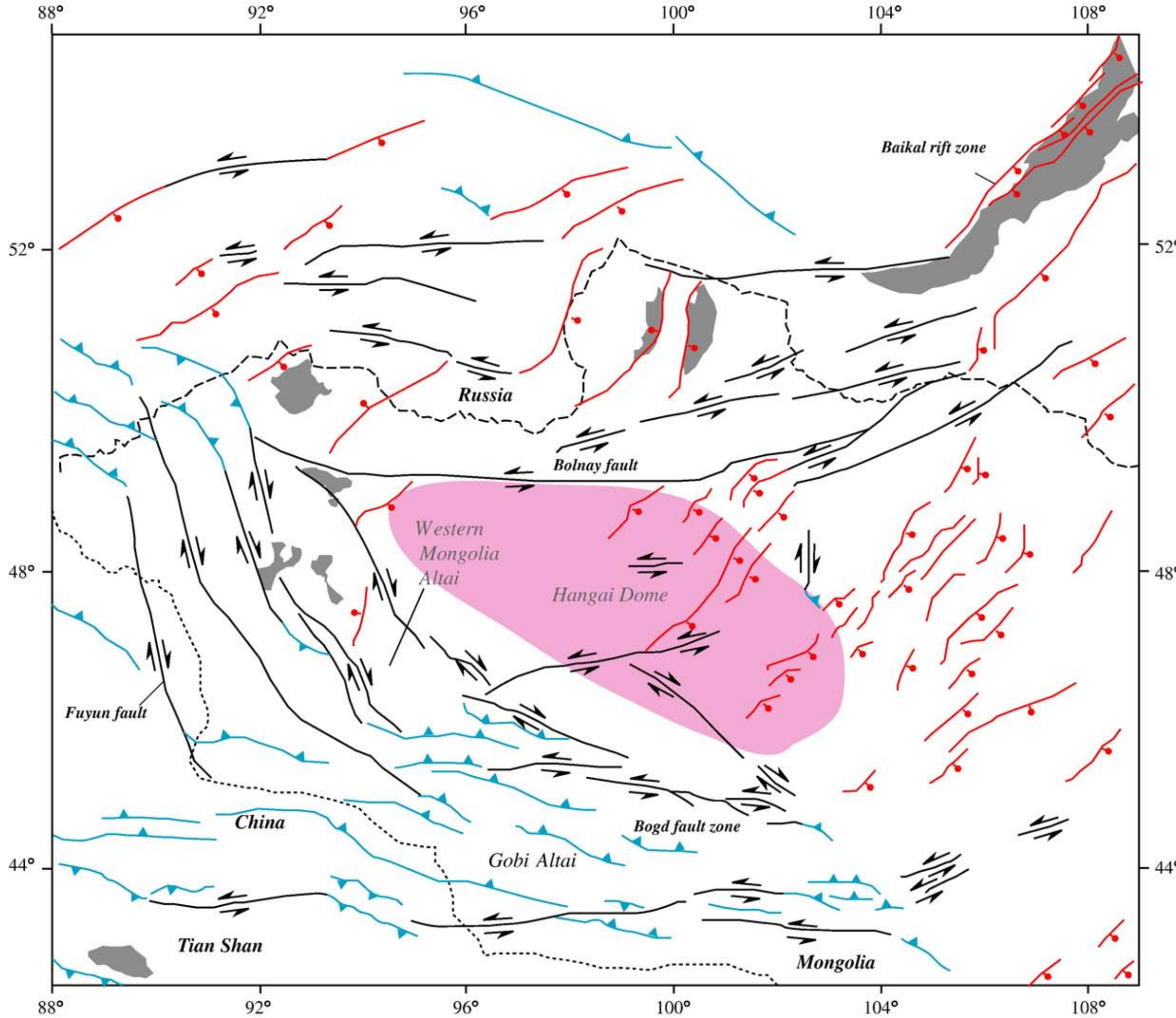

Fig. 2. Active fault map of western Mongolia and its neighboring regions. Sources of information are mainly from Abdrakhmatov et al. (1996), Cunningham (1998, 2005), Webb and Johnson (2006), Walker et al. (2006, 2007, 2008) and Briggs et al. (2007).

Late Miocene to Pliocene (10–5 Ma), estimated by apatite fission-track cooling ages (De Grave and van den Haute, 2002; De Grave et al., 2007).

Mongolia and its neighboring western China and southern Siberia expose complex fault systems typically expressed by large north- and east-trending strike-slip faults linking with either northeast-trending normal faults or northwest-trending thrusts (Fig. 2). North-trending strike-slip faults are right-lateral whereas the east-trending faults are left-lateral (Fig. 2) (Abdrakhmatov et al., 1996; Cunningham, 1998, 2005; Webb and Johnson, 2006; Walker et al., 2006, 2007, 2008). The southern tip of the Altai range and the eastern tip of the Tian Shan merge in southwestern Mongolia, where active deformation is dominated by several major east-trending left-slip fault systems (Cunningham et al., 1996). In west-central Mongolia Cenozoic deformation is expressed by late Cenozoic normal faulting in the Hangay Dome region, which covers an area of ~200,000 km^2 with numerous flat-topped peaks over 3000 m with widespread Neogene–Quaternary basaltic flows (Cunningham, 2001) (Fig. 2). Despite its high elevation, the region has experienced little erosion sine 150 Ma (Jolivet et al., 2007). Some of the normal faults of Cunningham (2001) in the Hangay region have been reinterpreted as east-trending left-slip faults (Walker et al., 2007). Although the high topography of the Hangay dome was initially related to mantle plume activities (Windley and Allen, 1993), geochemical and geophysical investigations indicate that the thermal anomaly is shallow and the late Cenozoic volcanism was largely related to partially melting in the asthenosphere of the upper mantle in the region that argue against the proposal (Petit et al., 2002, Barry et al., 2007). GPS slip rates on the major east-trending left-slip faults are typically in the range of 1–3 mm/yr (Calais et al., 1998, 2003).

The 1500-km long Baikal rift system is the most dominant Cenozoic structure in southeast Siberia (Fig. 2). GPS measurements across the rift zone show extension at a rate of 4.5 ± 1.2 mm/yr in a WNW–ESE direction (Lesne et al., 1998, 2000). Logatchev and Zorin (1987) and Logatchev (1994) proposed a two-stage model for the development of the Baikal rift: (1) a slow rifting phase between the Late Oligocene and Early Pliocene and (2) a fast rifting phase since the Late Pliocene. This traditional division of rift evolution in the Lake Baikal region was mainly based on sedimentation patterns (e.g., Kashik and Mazilov, 1994) and has been later revised by Delvaux et al. (1997) based on detailed structural analysis of large populations of Cenozoic faults (also see a recent synthesis by Petit and Deverchere, 2006). They showed that the Lake Baikal region has experienced three stages of deformation. (1) Between Oligocene to Late Miocene, the region underwent north–south contraction and transpressional deformation along northeast-striking shear zones. (2) Transpressional deformation evolved into transtensional tectonics in the Late Miocene to Early Pliocene. (3) Since the Late Pliocene, the Baikal rift began to open by northwest–southeast extension. The inferred extensional deformation starting since the Late Miocene in the Baikal region is generally consistent with the Quaternary extensional rate of $3.2 \pm$

0.5 mm/yr (San'kov et al., 2000) and the timing of partitioning between the central and north Baikal basins (Mats et al., 2000). The later pure rifting event in the Baikal region was coeval with the initiation of rifting and emplacement of north-trending dikes in the Tibetan plateau (Yin et al., 1994; Coleman and Hodges, 1995; Harrison et al., 1995; Blisniuk et al., 2001; Williams et al., 2001), suggesting that east–west extension in Asia shares the common cause and was most likely induced by plate boundary processes along the eastern margin of Asia (Yin, 2000).

2.5. North China deformation domain

Cenozoic deformation in north China between latitudes 31°N and 42°N and longitudes 100°E and 122°E is expressed by the development of the Paleogene Huabei basin in the east and Neogene rift systems around the Ordos block in the west (Ye et al., 1987; Zhang et al., 1998). The Paleogene Huabei Basin formed by early Tertiary back-arc extension associated with Late Paleocene and Eocene basaltic eruptions (Ye et al., 1987). Due to post-rifting thermal subsidence, the Huabei basin is largely covered by Neogene to Quaternary sediments.

There are three major graben systems around the Ordos block: the Yinchuan rift along its western margin, the Hetao rift along its northern margin, and the Shan rift along its eastern margin (Fig. 1). The Yinchuan rift has been assigned to initiate in the Oligocene because of the presence of Oligocene red beds in the rift basin (Ye et al., 1987). However, a closer examination of seismic reflection profiles across the rift (Ningxia, 1989) suggests that the syn-rift sediments are late Miocene and Pliocene in age, which thicken towards the rift-bounding faults whereas the Oligocene strata are pre-rift deposits that maintain constant thickness and are present on both sides of the rift-bounding faults. Similarly, the southernmost part of the Shanxi rift basin (i.e., the Weihe graben of Ye et al., 1987) contains Oligocene and possible Late Eocene strata. Because Paleogene strata are also widely distributed outside the southern segment of the Shanxi rift (Wang et al., 1996), it is possible that the inferred Paleogene initiation of rifting by Zhang et al. (1999) was due to assigning the pre-rift sequence to syn-rift sequence.

Wang et al. (2002) showed that Pn velocity is low below the Shanxi rift, indicating relatively hot upper mantle. GPS studies by Shen et al. (2000) show that contrary to the early inferences that the Shanxi rift has accommodated significant right-slip motion (e.g., Wang et al., 1996), at least the northern part of the rift is under ESE–WNW extension at a rate of 4 ± 2 mm/yr.

The Yinchuan and Shanxi rifts terminate in the south at the left-slip Haiyuan and Qinling fault zones (Burchfiel et al., 1991; P. Zhang et al., 1991; Y.Q. Zhang et al., 1998; Lasserre et al., 2002) that extend eastward to the Dabei Shan region (Ratschbacher et al., 2000) and westward to central Tibet linking with the left-slip Kunlun fault system via several large-scale step-over structures marked by a series of basins (e.g., the Gonghe basin, Linxia basin, etc.) (Pares et al., 2003; Fang et al., 2003, 2005; Garzione et al., 2005). "Because the Kunlun fault is the northern boundary of the north-trending late Cenozoic rifts in central Tibet, the tectonic linkage between the rifts in Tibet and north China implies a common dynamic cause such as deep mantle flow or a change in boundary along the eastern margin of Asia (Yin, 2000; Yin and Harrison, 2000). The Hetao graben is the northern extension of the Yinchuan graben (Ye et al., 1987). However, how the Hetao and Shanxi rifts terminate in the north is not clear. GPS studies in this region show a broad left-slip shear zone trending east–west separating the Hetao and Shanxi grabens in the south and the stable Amurian plate to the north (Shen et al., 2000).

The Tanlu fault zone bounds the eastern edge of the modern Huabei Basin and is a first order Cenozoic tectonic feature in east Asia (Fig. 1). Its Cenozoic development may have been related to opening of Bohai Bay during Paleogene back-arc extension (e.g., Allen et al., 1997; Ren et al., 2002a). Zhang et al. (1999) showed that the southern segment of the Tanlu fault experienced three phases of deformation: (1) normal-slip, and (2) left-slip with a normal component, and (3) right-slip with a minor normal component. GPS survey shows that the fault is extensional accommodating east–west extension (Shen et al., 2000).

2.6. Eastern Asia margin deformation domain

The eastern margin domain is comprised of several large back-arc basins in the western Pacific (Sea of Okhotsk, Japan Sea/East Korea Sea, and South China Sea) and a highly extended (and thus thinned) continental margin (Bohai Bay, East China Sea, and Bay of Siam) (Fig. 1) (Ren et al., 2002a; Zheng et al., 2006; Chen et al., 2008). The back-arc basins and the extended domain of continental margin are characterized by (1) rhombic geometry on map view and (2) high-angle relationship between the long dimension of the basins and the trend of the nearby trench (Fig. 1).

Motion of two right-slip fault zones and coeval Eocene to Early Miocene extensional faults created the Sea of Okhotsk and the Kurile Basin floored by oceanic crust along its eastern margin (Fig. 1) (Worrall et al., 1996). Worrall et al. (1996) proposed that Eocene–Oligocene extension in the Sea of Okhotsk was linked with development of the Baikal rift via a linking east-trending left-slip fault system. This interpretation is problematic because Eocene extension in the Sea of Okhotsk predates even the earliest estimated initiation age for the Baikal rift (Logatchev and Zorin, 1987). The magnitude of extension over the Baikal rift (<30 km, Mats et al., 2000) is also too small in comparison to the amount of extension required to open the Sea of Okhotsk.

Extension in Japan Sea (also known as the East Korea Sea) initiated in the Late Paleocene and Eocene (Lallemand and Jolivet, 1986; Celeya and McCabe, 1987). However, the oceanic crust in the basin was not created until the Late Oligocene and lasted to the end of the Early Miocene (30–15 Ma). The Japan Sea is bounded in the northeast by the right-slip Sakhalin fault that may have accommodated about 400 km of displacement during the opening of the basin (Jolivet et al., 1994). There are three models for the origin of the Japan Sea. Lallemand and Jolivet (1986) suggested that the Japan Sea was developed as a pull-apart basin between two right-slip fault. Jolivet et al. (1990) and Yue and Liou (1999) proposed that the opening of the Japan Sea was related to the development of the left-slip Altyn Tagh fault in northern Tibet. Both of these models consider the development of the Japan Sea to have been associated with the Indo–Asian collision. Alternatively, the opening of the Japan Sea may be caused by back-arc extension (Jurdy, 1979; Celeya and McCabe, 1987) during slow convergence between Eurasia and Pacific plate (Northrup et al., 1995).

The Bohai Bay extensional system is separated from Japan Sea by the Korea peninsular that appears to have experienced little Cenozoic extension (Fig. 1). This extensional system extends southward to the East China Sea and westward to the Huabei Basin (Zhao and Windley, 1990; Allen et al., 1997; Ren et al., 2002a). Extension in the Bohai Bay region started in the Paleocene and was most active between the Eocene and latest Oligocene (Allen et al., 1997; Ren et al., 2002a). Rift-related structures and sedimentary sequences are overprinted by Quaternary dextral transpressional deformation, causing inversion of some earlier normal faults (Allen et al., 1997). The Triassic Tanlu fault is reactivated in the Cenozoic in the Bohai Bay area with the following deformation history: (1) Paleocene–Eocene transtensional tectonics, (2) Oligocene–Early Miocene transpressional tectonics, and (3) Middle Miocene to present-day uniform subsidence (Yu et al., 2008; Tong et al., 2008).

The continental margin of the East China Sea consists of three tectonic zones: (1) the East China Sea extended continental shelf, (2) the Taiwan–Sinzi thrust-fold zone, and (3) the Okinawa Trough (Fig. 1). The tectonic evolution of the East China Sea was best summarized by Zhou et al. (1989) and Kong et al. (2000). Between the latest Cretaceous and earliest Paleocene, extension occurred in the East China Sea as expressed by the development of detachment faults. This extensional event is part of widely distributed extension in east Asia (Ren et al., 2002a). Between the Late Paleocene and Early Oligocene, extension was focused in the East China Sea region. Significant crustal thinning during this period was manifested by the development of normal faults, development of a narrow basin, and rapid subsidence associated with the basin formation (Kong et al., 2000). Contraction began in the central and northern Taiwan–Sinzi thrust-fold zone in the Middle Oligocene (Kong et al., 2000) and was significantly intensified in the late Middle Miocene (Ren et al., 2002b). This event may have been associated with the subduction of the Palau–Kyushu ridge on the Philippine plate (Kong et al., 2000). Due to the presence of a very thick sequence of syn- and post-rift sediments, the age of the southern Taiwan–Sinzi thrust-fold zone is poorly constrained. Kong et al. (2000) suggested that the southern Taiwan–Sinzi folded zone initiated in the Late Miocene, possibly related to collision between the Luzon arc and Eurasia. In contrast, Sibuet et al. (2002) proposed that the Taiwan–Sinzi thrust-fold belt terminates its development at ~15 Ma during a major plate reorganization at the junction of the Philippine Sea plate and South China Sea. The development of the Okinawa Trough is the youngest deformation event in the East China Sea. Its opening may have started in the Late Miocene associated with clockwise rotation of the Ryukyu arc (Sibuet et al., 1998).

The South China Sea Basin is a northeast-trending rhombic basin floored by oceanic crust in its central part (Fig. 1). Geology of the South China Sea is summarized by Zhou et al. (1995) who proposed three stages of rifting in the region. (1) During the Late Cretaceous to Early Eocene (~87–50 Ma), distributed extension created half grabens that are filled by continental red beds with considerable amount of volcanic and metamorphic clasts. Zhou et al. (1995) attributed the extension to slab retreat of the Pacific plate. (2) In the Middle Eocene, rifting started again resulting in deposition up to 1000-m thick lacustrine deposits. (3) Between Late Eocene and Late Oligocene (38–23 Ma), intense extensional deformation renewed. This event is restricted to the South China Sea region that eventually led to the breakup of the continental margin and the opening of the South China Sea (Briais et al., 1993). Since the Late Miocene to Early Pliocene, contractional deformation prevailed in the South China Sea as expressed by folding and thrusting of the continental shelf sequences (Zhou et al., 1995; Sibuet et al., 2002). This compressional event marks the collapse of the South China as indicated by subduction of the South China Sea beneath Philippine plate and collision of the Luzon arc with Asia that created the Taiwan orogenic belt (Teng, 1990; Sibuet et al., 2002). The northernmost part of the South China Sea is floored by late Cretaceous–Paleocene oceanic crust that has been interpreted to be part of the proto-South China Sea (Sibuet et al., 2002).

Because of the presence of Late Cretaceous rifts in the South China Sea region, it has been suggested that the South China Sea began opening since this time (e.g., Hall, 2002). This interpretation should be treated with caution, because the first and second rifting events also occurred over much of east Asia and were not restricted to South China Sea region only. Late Cretaceous extension covers an area from the continental margin westward as far as to the Lake Baikal region (Ren et al., 2002a). Such widely distributed extension may be regarded as diffuse intracontinental extension instead of a back-arc extensional event causing the breakup of the continental margin. However, the subsequent Late Paleocene–Early Eocene extension was restricted to the eastern margin of Asia and was responsible for major crustal thinning of the region (Fig. 1). It appears that the Oligocene–Middle Miocene extensional event is rather localized along the eastern margin of Asia, only occurring in the South China Sea and Japan Sea (Fig. 1).

The Philippine Sea plate is the largest back-arc basin against the eastern margin of Asia. It consists of four tectonic domains with different ages of oceanic crust: the late Cretaceous Huatung basin, the Eocene–Oligocene west Philippine basin, the Oligocene–Miocene Shikoku–Parece Vela basins, and the latest Miocene–present Mariana basin (Fig. 1). The Gagua ridge that separates the Huatung basin from the western Philippine basin is a major plate boundary, possibly representing an inactive transform fault (Deschamps et al., 2000; Sibuet et al., 2002). The oceanic floor of the Huatung basin formed in the Late Cretaceous whereas the rest of the Philippine Sea plate was created in the Cenozoic in three separate stages. (1) The west Philippine Sea basin experienced sea-floor spreading between 57 and 34 Ma (Hilde and Lee, 1984). (2) A slower spreading may have lasted until ~30 Ma when the Shikoku and Parece Vela basins began to form by back-arc extension (Hall, 2002). Opening of the Shikoku and Parece Vela basins terminated at about 15 Ma (Okino et al., 1998, 1999). (3) The Mariana back-arc basin started opening since 6 Ma (Hall, 2002).

Although interpretation of Cenozoic deformation of Asia in a regional pate-tectonics context depends critically on the kinematic history of the Philippine Sea plate, its origin remains poorly understood because its movement history cannot be linked with the global plate circuit as it is surrounded by subduction zones (Hall, 2002). Efforts of constraining the movement history of the plate were made by using paleomagnetic data within and around the plate boundaries and by studying the deformation history around the plate margins (Hall et al., 1995a,b; Lewis and Byrne, 2001). Paleomagnetic data indicate that the crust in the northwestern part of the Philippine Sea moved across the equator from the south at about 60 Ma (Kinoshita, 1980). According to Hall (2002), the Philippine Sea plate also experienced clockwise rotation of about 50° between 55 and 45 Ma, no rotation from 40 to 25 Ma, and finally a minor amount of clockwise rotation from 25 to 5 Ma. The present motion of the Philippine Sea plate continues to rotate clockwise with respect to the Eurasian plate. That is, the convergent rate is at 30 mm/yr in its northwest corner and at 92 mm/yr in its southwest corner (Seno et al., 1993).

There are several contrasting models for the origin of the western Philippine basin. (1) It formed by trapping of a part of the Kula–Pacific spreading center (Lewis et al., 2002). (2) It was originated by back-arc spreading above the postulated North New Guinea plate that is now completely consumed by subduction. This plate was located between the Pacific plate in the northeast and the Philippine plate in the southwest and was subducting southward beneath the western Philippine basin (Seno and Muruyama, 1984). (3) The west Philippine basin was generated by back-arc spreading above the subducting Australian plate (Hall et al., 1995a). Because of the uncertainties about the movement history and tectonic setting for the formation of the Philippine Sea plate (mostly its western basin), the configuration of plate boundaries in the western Pacific during the Tertiary remains an important and unresolved problem (Northrup et al., 1995).

2.7. Turkish–Iranian–Caucasus orogen

Equally important to the India–Asia collision in the Cenozoic tectonic history of Asia is the Arabia–Asia collision (Berberian and King, 1981) (Fig. 1). The two plates are currently converging at a rate of 20–30 mm/yr in the north–south direction (McClusky et al., 2000; Jackson et al., 2002; Reillinger et al., 2006). The decadal convergence rate determined by the GPS studies is similar to the rate over the past 65 Ma that varied little with time (McQuarrie et al., 2003). The oblique alignment of the Arabia–Asia convergence front caused right-slip tranpressional deformation (Fig. 1) (e.g., Jackson et al., 2002). The active tectonics of the Turkish–Iranian–Caucasus orogen is expressed by north–south contraction along the northern and southern margins of the orogen (around the Caspian Sea including the Greater Caucasus in the north and the Zagros belt in the south) and strike-slip faulting in the core of the orogen across the Turkish–Iranian plateau (Philip et al., 1989; Jackson et al., 2002; Sengor et al., 2003; Allen et al., 2004; Copley and Jackson, 2006; Guest et al., 2006a). Strike-slip faulting also dominates the western and eastern flanks of the orogen (Fig. 1). For example, eastern Iran is dominated by north-trending right-slip faulting and northeast-striking left-slip faulting (Alavi, 1994). The current active fault pattern in the Turkish–Iranian–Caucasus orogen may have been established since 3–7 Ma. This is drastically different from the deformation pattern dominated by crustal thickening during the early stage of the Arabia–Asia collision (Jackson et al., 2002; Allen et al., 2004; Copley and Jackson, 2006), despite a constant convergence velocity in the past 65 Ma between Arabia and Asia (McQuarrie et al., 2003).

Estimates on the age of the initial collision between Arabia and Asia vary widely from prior to ~80 Ma to ~12 Ma (Sengör and Kidd, 1979; Berberian and King, 1981; Dewey et al., 1986; Robertson et al., 2006). The older collision age of >80 Ma was determined by the timing of Cretaceous melange emplaced over the continental shelf of Arabia and unconformably overlain by Maastrichtian (78–65 Ma) shallow marine deposits (e.g., Alavi, 1994). Relating the age of deformation in the Turkish–Iranian–Caucasus orogen to the initial contact of Arabia with Asia, the estimated age of collision ranges from Eocene to Oligocene (Hempton, 1987; Yilmaz, 1993), and even as young as 6–7 Ma or ~12 Ma corresponding to rapid exhumation events in Iran (Axen et al., 2001; Guest et al., 2006b). Finally, using the timing of transition from marine to terrestrial sedimentation in eastern Turkey, Sengör and Kidd (1979) and Sengör et al. (1985) considered that the collision between Arabia and Asia started in the Middle Miocene at ~13 Ma. The end of marine sedimentation should be regarded as a minimum age for continental collision, as the continental shelves of the two continents could have been in contact while marine sedimentation was on-going. A modern example of the process is the development of the Persian Gulf, where marine sediments are actively deposited although collision between the Arabian and Asian continents had already occurred (see extensive discussion in Yin and Harrison, 2000). The onset age of the Arabia–Asia collision has also been evaluated from the history of foreland sedimentation along the collision zone. Reinterpreting the age of thrust-related coarse-grained sediments and the analysis of the timing of syn-orogenic transport systems support an interpretation that the initial collision occurred prior to the early Miocene and possibly in the Oligocene or even in the Eocene (Fakhari et al., 2008; Horton et al., 2008).

The inference of latest Cretaceous collision between Asia and Arabia presents several problems. First, despite such an early collision, marine sedimentation continued in the Zagros belt until the Early Miocene, which would imply an unusually wide continental shelf for the Arabian continent. Second, at about 70 Ma, the poles between the united Africa–Arabia plate and the Eurasia plate were still 800–1000 km apart (Besse and Coutillot, 1991). This means that that the northern edge of Arabia was probably >1000 km farther south of its current position at ~80 Ma when the initial collision between Arabia and Eurasia occurred. However, there has been no documentation of such large an amount of crustal shortening across the Turkish–Iranian–Caucasus orogenic belt since the Late Cretaceous. The estimated shortening across the Zagros fold-thrust belt is on the order of only 70 km (McQuarrie, 2004; Allen et al., 2004). Third, a large volume of calc-alkline volcanic extrusions occurred continuously from Late Eocene to Early Miocene time (40–20 Ma) in Iran, suggesting continuous subduction of oceanic crust at this time (e.g., Stoneley, 1981). If Eocene–Miocene volcanism was related to slab breakoff after the initial continental collision (Horton et al., 2008), similar to the eruption of the Linzizong volcanic sequence in southern Tibet immediately after the India–Asia collision (Yin and Harrison, 2000), then collision could have occurred at ~40 Ma, which is still much younger than the inferred older collision age of ~80 Ma. Fourth, Babaie et al. (2001) showed that the Neyriz ophiolite complex in southwest Iran is tectonically juxtaposed against an island arc volcanic sequence. The juxtaposition occurred during the subduction of the Neo-Tethyan oceanic crust rather than continental collision. Regardless of the exact timing of collision between Arabia and Asia, the major crustal thickening process appears to have initiated at the end of the Eocene at about 37 Ma in the region (Berberian and King, 1981). This event is expressed by the local development of angular unconformity and subsequent southward migration of foreland-basin depositional centers throughout the rest of the Cenozoic (Berberian and King, 1981; Hessami et al., 2001; Guest et al., 2007).

The most striking active tectonic feature of the Turkish–Iranian–Caucasus orogen is the presence of a series of northwest-striking right-slip transpressional fault zones best exemplified by the Kopeh Dagh fault zone, the Ashgabat fault zone, and the Zagros fault zone from north to south (Fig. 1) (Sengör et al., 1985; Allen et al., 2004). These faults are parallel to a series of right-slip transpressional fault zones farther to the north in central Asia that link to the Tian Shan and Altai Shan thrust belts and large-scale left-slip faults in western Mongolia. The remarkably uniform spacing of these faults (500–600 km) (Fig. 1) and their continuous distribution from the Zagros to southern Siberia suggest the initiation of the large right-slip fault zone was related to collision between Arabia and Asia. Because the right-slip faults in central Asia are linked with the Tian Shan thrust belt, it raises the possibility that the development of the Tian Shan thrust belt was related to continental collision to the west, not to the south along the Himalayan front (cf., Cobbold and Davy, 1988). Collision between Arabia and Asia as the cause for the development of the Tian Shan may also explain why the amount of shortening decreases systematically from west to east (Avouac et al., 1993) and a late initiation of the Tian Shan orogen (i.e., ~24–20 Ma, see, for example, Yin et al., 1998) relative to the onset of the India–Asia collision at about 60–50 Ma (Yin and Harrison, 2000).

2.8. Western Turkey extrusion system

The western Turkey extrusion system (also known as the Anatolian extrusion system) is bounded in the north by the right-slip North Anatolian fault zone, in the east by the left-slip East Anatolian fault zone, to the west by the Aegean–Cretan extensional domain, and to the south by the south-facing Hellenic arc (Fig. 1) (McKenzie, 1972, 1978; LePichon and Angelier, 1981; Sengör et al., 1985, 2005; Armijo et al., 1999; Dhont et al., 2006; Mantovani et al., 2006; Piper et al., 2006; Baraganzi et al., 2006). The right-slip North Anatolian fault has a total displacement of 80–85 km and locally could be as high as 100–110 km (Sengör and Kidd, 1979; Westaway, 1994; Hubert-Ferrari et al., 2002). Its Quaternary slip rate is ~18–20 mm/yr (Hubert-Ferrari et al., 2002; Kozaci et al., 2007) and the GPS-determined slip rate is ~22 mm/yr (Reillinger et al., 2006). The left-slip East Anatolian fault is the northern extension of the Dead Sea fault system that accommodates relative plate motion between the African and Arabian plates (e.g., Westaway, 2004a,b, 2006 and references therein). The Dead Sea fault has a total slip of ~105 km in the south and 50–80 km in the north and was initiated at 19–15 Ma (Garfunkel, 1981; Dewey et al., 1986; Westaway, 2004a). The East Anatolian fault zone serving as a transform fault between the Arabian and Eurasian plates consists of several strands, some of which are currently inactive (Lyberis et al., 1992; Westaway, 2004a). Its Quaternary and GPS-determined slip rates are 8–9 mm/yr with a total slip of ~37 km (McClusky et al., 2000; Westaway, 2004a,b).

The initiation age of the North and East Anatotlian faults were estimated by two methods. The first was to use the ages of strike-slip related basins, which constrain the initiation ages of the North and East Anatolian faults at 15–11 Ma (Sengör et al., 1985, 2005). The second method was to use the total slip on each fault divided by their current slip rates, which yields an initiation age at about 5–7 Ma (Hubert-Ferrari et al., 2002; Westaway, 2004a). The differences in the estimated fault ages may reflect progressively accelerated rates of westward extrusion of the Turkey block.

2.9. Aegean–Cretan extensional system in Western Asia

The Aegean Sea, Cretan Sea, western Turkey and southern Balkan constitute a large Cenozoic extensional system accommodating north–south and northeast–southwest extension at the western end of the Arabia–Asia collision zone (Fig. 1) (McKenzie, 1970, 1972; LePichon and Angelier, 1979, 1981; Sengor et al., 2005; e.g., Burchfiel et al., 2008b). Large magnitude extension in the Cretan Sea was accommodated by the development of low-angle extensional detachment systems (Lister et al., 1984; Buick 1991; Jolivet et al., 1996; Forster and Lister 1999; Ring et al., 2001; Rosenbaum et al., 2007;

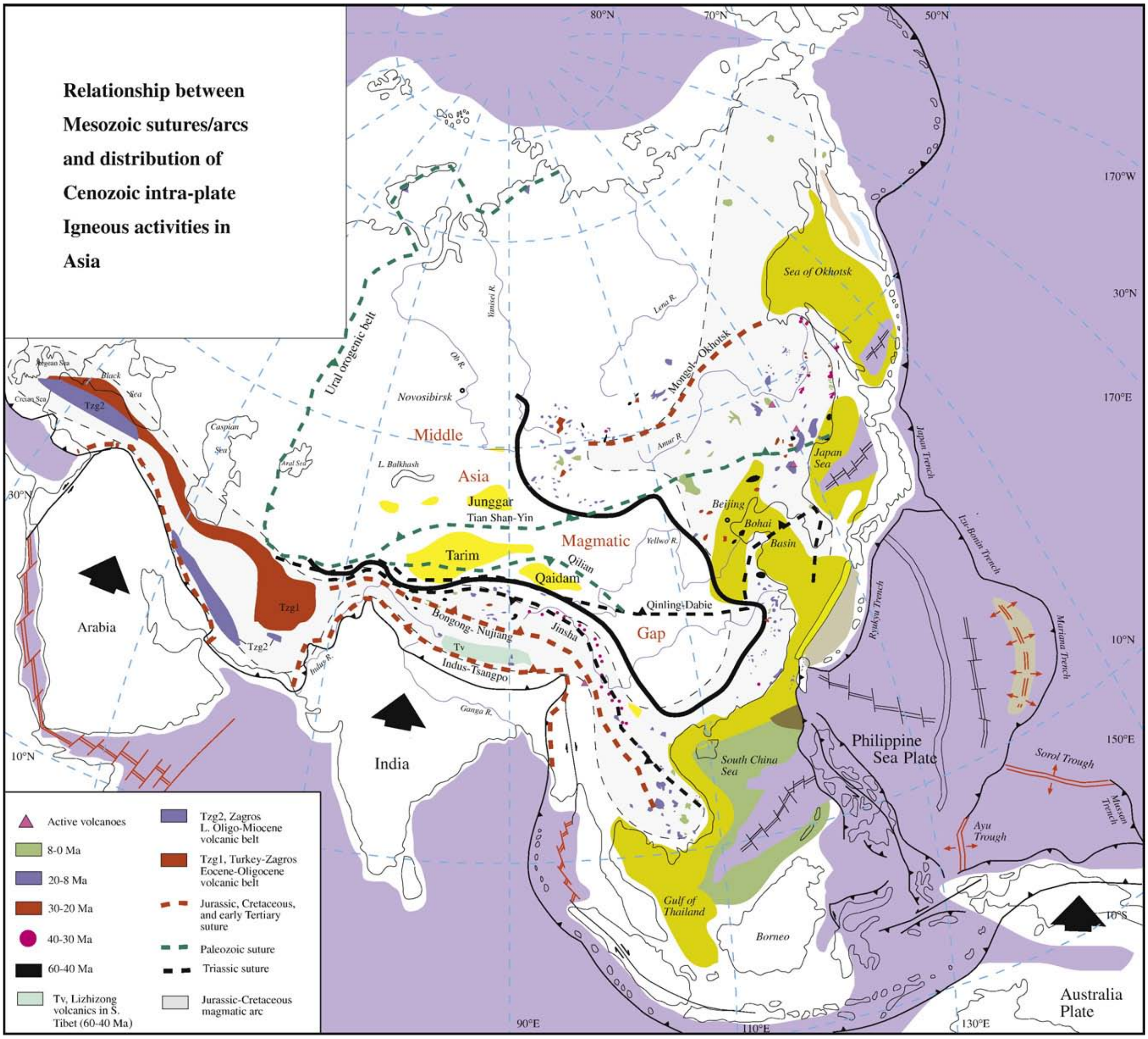

Fig. 3. Relationship between Cretaceous sutures and arcs and distribution of Cenozoic volcanic rocks in Asia. In central Asia and eastern China, the age of Cretaceous igneous rocks is from Ren et al. (2002a). In Tibet, the age of Cretaceous igneous rocks is based on Xu et al. (1985), Coulon et al. (1986), Liu (1988), and Pan et al. (2004).

Marsellos and Kidd, 2008). Extension in the region was originally thought to have started at 13–5 Ma (LePichon and Angelier, 1979), but subsequent dating indicates that at least in the Cretan Sea the onset of extension occurred at 25–21 Ma (Lister et al., 1984; Ring et al., 2001).

LePichon and Angelier (1979) attributed Cenozoic extension across the Aegean–Cretan–western Turkey system to a southward retreat of the Hellenic arc during which the arc had also rotated about vertical axes in a clockwise sense with its curvature tightened considerably (also see Marsellos and Kidd, 2008). The retreat of the Hellenic arc was related to sinking of the slab and gravitational spreading of the thicker Aegean crust towards the Hellenic trench (LePichon and Angelier, 1979). Although the kinematic and geometric connections between the Aegean extensional domain and the right-slip North Anatolia fault have long been recognized (see review by Sengor et al., 2005), the causal relationship between the latest Oligocene–early Miocene (25–20 Ma) onset of Hellenic arc retreat (e.g., Lister et al., 1984; Ring et al., 2001) and westward extrusion of the Anatolia block starting at 15–5 Ma (Sengör et al., 1985; Hubert-Ferrari et al., 2002; Sengor et al., 2005; Westaway, 2004a) remains unclear. On one hand, arc retreat and its induced back-arc extension in the Aegean and Cretan Seas could have facilitated lateral extrusion of Turkey in a manner suggested by Royden et al. (2008) for the development of the India–Asia collision zone. On the other hand, northward penetration of the Arabia plate into Asia may have pushed the asthenosphere below the Turkish–Iranian–Caucasus orogen to have a counterclockwise flow pattern below Turkey and the Aegean and Cretan Seas, which forced the Hellenic subducting slab to retreat. The strength of the flow (i.e., its velocity) may decrease westward along the Hellenic arc away from the Arabia–Asia collision front at the Zagros, resulting in the observed clockwise rotation of the Hellenic arc.

2.10. Laptev–Moma rift in Northernmost Asia

Although it has long been noted that the mid-ocean ridge of the Atlantic ocean extends across the Arctic ocean into the northern margin of Asia (Chapman and Solomon, 1976; Michael et al., 2003), the nature and location of the projected continental extension are not understood due to lack of detailed studies. Current knowledge in the western literature is mostly based on a compilation of early work by Russian scientists. Cook et al. (1986) and Fujita et al. (1990) considered the Arctic mid-ocean ridge (also known as the Gakkel Ridge, see Fig. 1) into the

Laptev Sea as a continental rift and correlated it farther inland with the Moma rift that was developed mainly in the Miocene and Pliocene associated with volcanism. The Moma rift system switched to become a transpressional system at ~3 Ma (Cook et al., 1986; Fujita et al., 1990; Gaina et al., 2002).

3. Cenozoic igneous activities

3.1. Spatial distribution and relationships to deformation

Cenozoic igneous activity is widespread in Asia, extending from Lake Baikal in the north, Turkey in the west, Korean peninsular and Russia Far East in the east, and southeast Asia in the south (Fig. 3). A prominent east–west trending magmatic gap exists in central Asia that occupies the Junggar basin, the Tian Shan, the northern Altai, the Tarim basin, the Qilian Shan–Nan Shan region, the Qaidam basin, and the western part of South China. A narrow corridor lies along the eastern margin of South China that connects the northern igneous province with the southern igneous province. Cenozoic igneous activities are spatially correlated with the Late Jurassic–Cretaceous arcs. In the north, Cenozoic volcanism mainly follows the volcanic arc related to the closure of the Mongol–Okhotsk Ocean. Along the eastern margin of Asia, Cenozoic volcanism follows closely the Cretaceous arc produced by subduction of the Pacific plate. Similarly, from the Black Sea to Indochina across the Tibetan plateau, Cenozoic igneous activities follow the Jurassic–Cretaceous suture zones and related arcs within the Tethyan orogenic system (Fig. 3). It should also be noted that basaltic eruptions in Huabei and Songliao Basins of eastern China, the Tibetan plateau, Mongolia and Turkey occurred over regions previously occupied by Mesozoic arcs (Zhou et al., 1988;

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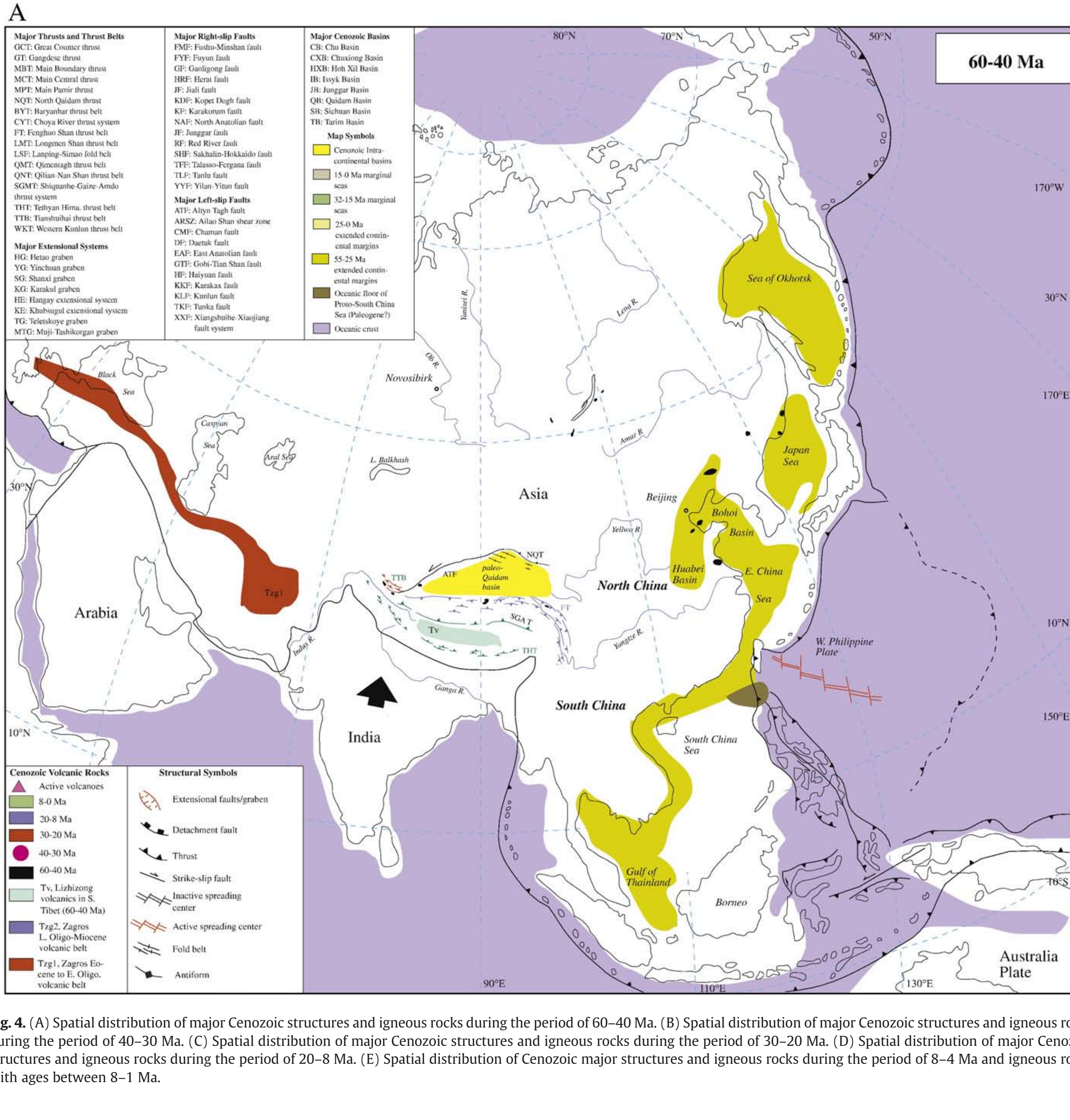
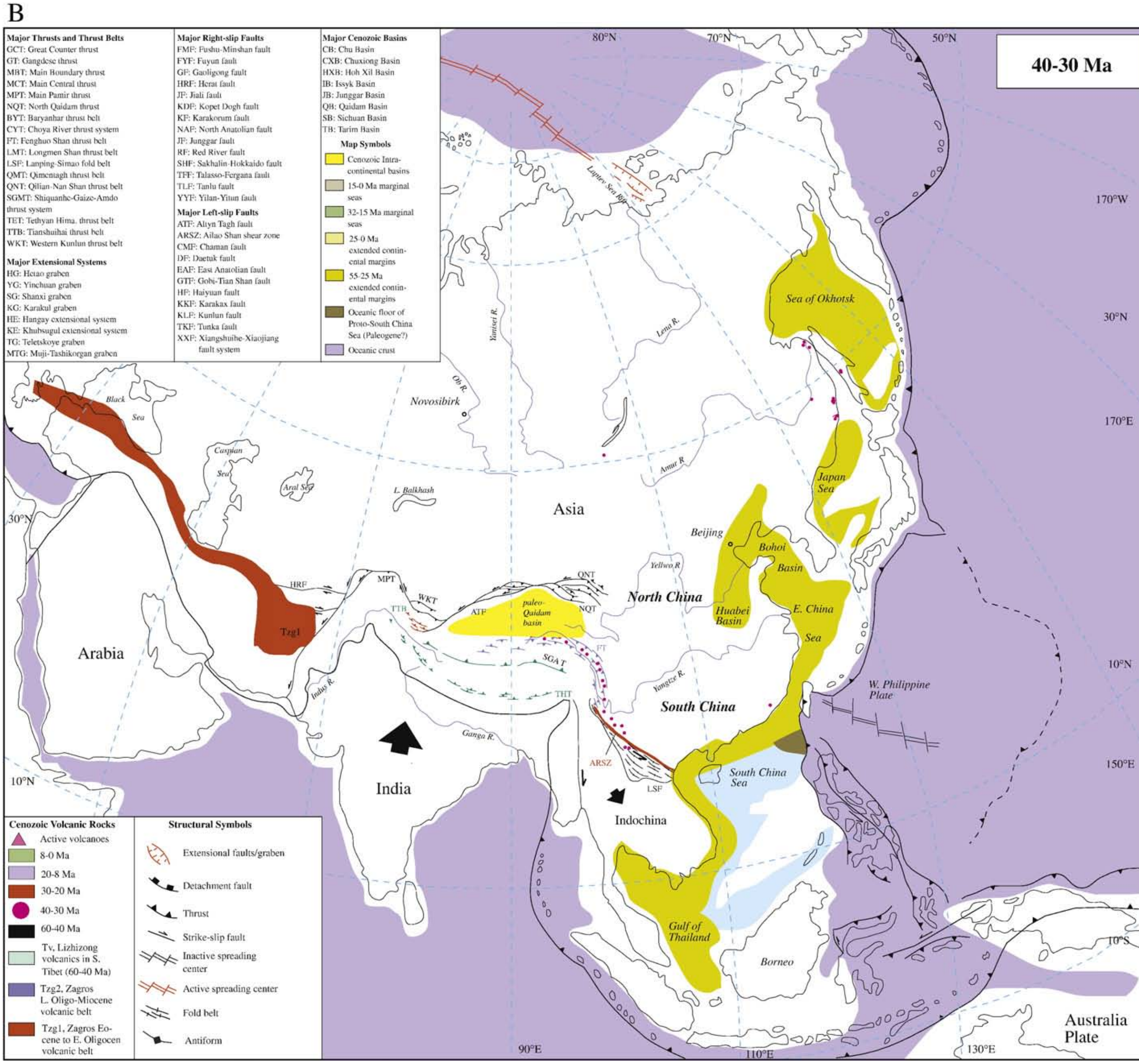


Fig. 4. (A) Spatial distribution of major Cenozoic structures and igneous rocks during the period of 60–40 Ma. (B) Spatial distribution of major Cenozoic structures and igneous rocks during the period of 40–30 Ma. (C) Spatial distribution of major Cenozoic structures and igneous rocks during the period of 30–20 Ma. (D) Spatial distribution of major Cenozoic structures and igneous rocks during the period of 20–8 Ma. (E) Spatial distribution of Cenozoic major structures and igneous rocks during the period of 8–4 Ma and igneous rocks with ages between 8–1 Ma.



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R.X. Liu et al., 1992; Barry and Kent, 1998; Li, 2000; Barry et al., 2003, 2007; Altunkaynak and Genc, 2008; Zhang et al., 2008).

Except in the Himalaya (Le Fort, 1996, Yin, 2006), Cenozoic igneous activity in Asia is expressed mostly by volcanism. In eastern Tibet (Chung et al., 1998, 2005; Wang et al., 2001), and Indochina (Lee et al., 1998), the Cenozoic igneous rocks were dated by the $^{40}Ar/^{39}Ar$ method. However, in the rest of Asia the rocks were mostly dated by the K–Ar method. The K–Ar method has the potential to overestimate the true age of a rock because it does not detect existence of access argon in the sample. As pointed out by Liu et al. (1992), the ages of some northern China volcanic rocks were overestimated some 15–100% higher than their true ages using the K–Ar method.

The age distribution of Cenozoic igneous rocks and coeval deformation across Asia are summarized in Fig. 4. The Miocene Himalayan leucogranites are not shown in the figure because they are volumetrically small (Yin, 2006). During the Paleogene and Eocene (60–40 Ma) (Fig. 4A), Cenozoic igneous activities are mainly concentrated in northern Iran (Berberian and King, 1981), northern Turkey (Okay and Satir, 2006; Dilek and Altunkaynak, 2007), and southern Tibet (i.e., the Lizizong volcanics; Xizang BGMR, 1992; Liu, 1988; Yin and Harrison, 2000). The igneous activities during this period in Iran are typically expressed by intrusion of intermediate plutons related to subduction of the Neo-Tethyan oceanic plate (Berberian and King, 1981). The Lizizong volcanic rocks erupted at 65–45 Ma mostly as ignimbrite sheets (Coulon et al., 1986; Pan et al., 2004; Mo et al., 2007). Plutons as young as ~32 Ma are emplaced in the southern Gangdese batholith belt in southeast Tibet (Harrison et al., 2000; Mo et al., 2007). The occurrence of the Lizizong volcanic rocks was attributed to arc magmatism related to the final phase of Neo-Tethyan subduction (e.g., Coulon et al., 1986; Yin et al., 1994). Because the initial India–Asia collision could have occurred at ~65 Ma (see summary by Yin and Harrison, 2000), the Lizizong volcanism could have been related to slab breakoff after initial continental collision (Yin and Harrison, 2000; Chemenda et al., 2000). It is also

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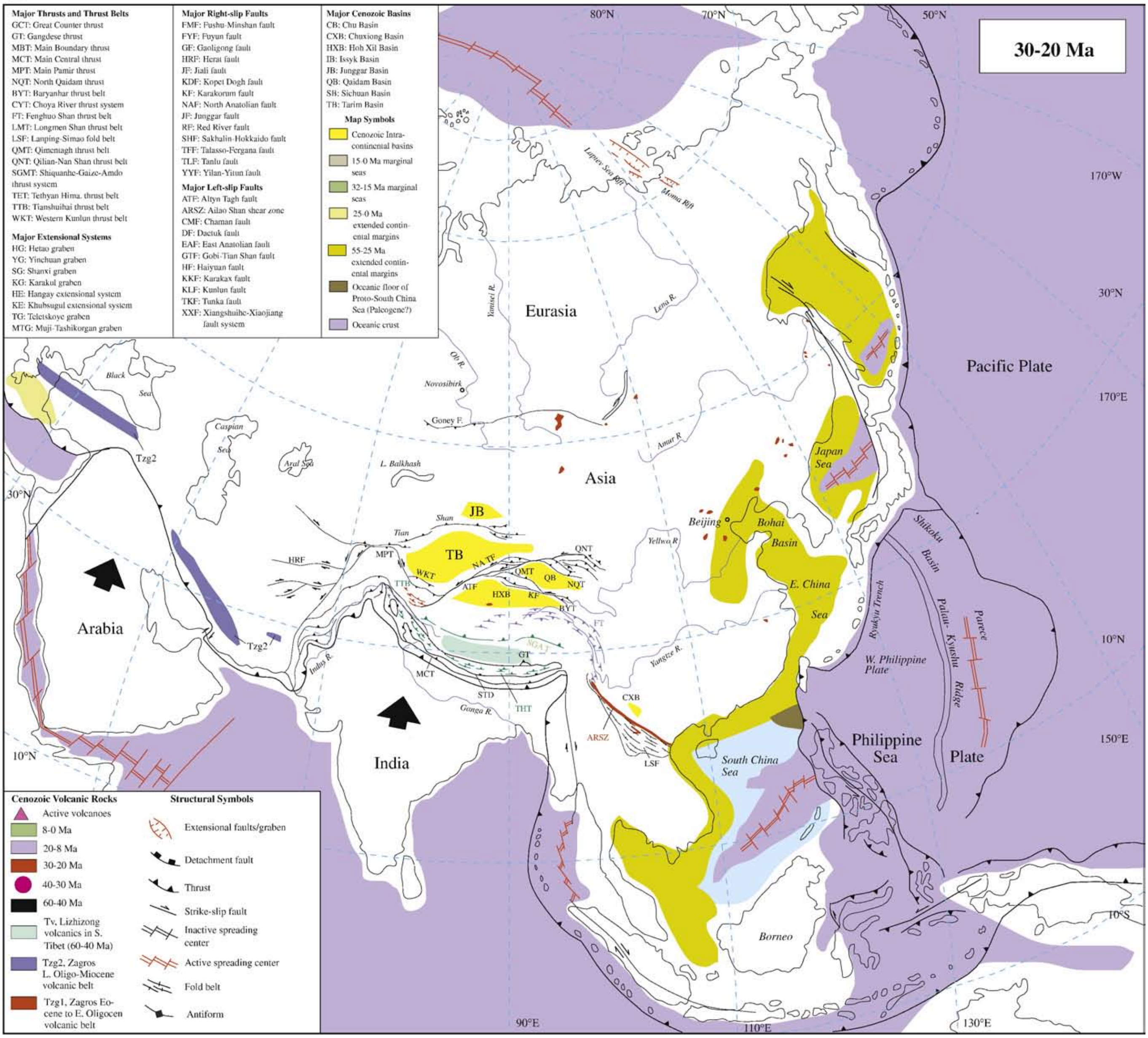

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possible that the Lizizong volcanic activity was related to slab rollback at the terminal stage of Tethyan subduction (Ding et al., 2003).

In the Huabei Basin, Bohai Bay, and Songliao Basin of northern China, basaltic eruption dominated between 60 and 40 Ma associated with back-arc extension (J.Q. Liu et al., 2001). Sparse basaltic eruptions also occurred in Bohai Bay and Huabei Basin of northern China (R.X. Liu et al., 1992, 2001), Far East Russia (Okamura et al., 1998), the Lake Baikal region of southeast Siberia (Rasskazov, 1994), and in Mongolia (Barry and Kent, 1998; Barry et al., 2003, 2007) (Fig. 3A). The Paleocene–Eocene volcanic rocks in the Baikal region were part of the continuous basaltic eruption starting in the latest Cretaceous at about 72 Ma (Rasskazov, 1994). Although Paleocene–Eocene volcanism along the eastern margin of Asia and in southern Tibet can clearly be associated with back-arc extension (Basu et al., 1991; Ren et al., 2002a; Ho et al., 2003) and consumption of the Pacific and Neo-Tethyan oceanic crust (Yin et al., 1994; Yin and Harrison, 2000), mechanisms for the coeval occurrence of basaltic rocks in Mongolia, southeast Siberia, and western China are still debated (Barry and Kent, 1998; Zhang et al., 2008). This is because there is no extensional tectonics associated with this igneous event. The earliest estimated age for extension in central Asia is Late Oligocene (Delvaux et al., 1997). The long distance between Paleocene–Eocene central Asia volcanic fields and the west Pacific trench systems also makes it difficult to attribute the volcanism to back-arc extension. We will return to this problem in Section 5.1 in the Discussion.

In the western Pacific, the most important event at this time is the formation of the western Philippine basin. The relationship between seafloor spreading in the Philippine Sea plate and the widely distributed extension along the eastern margin of Asia remains unclear. According to Hall (2002), Philippine plate was located near the equator at this time and may have only interacted with the southeastern tip of the Asian continental margin. If this reconstruction is correct, Paleocene–Eocene back-arc extension along the eastern margin of Asia must have been of limited magnitude or distributed in a wide zone, because no back-arc basins floored by oceanic crust were generated at this time. Widely distributed extension at this time may be due to the inherited thermal state of the lithosphere (i.e., hot and weak) from the development of the

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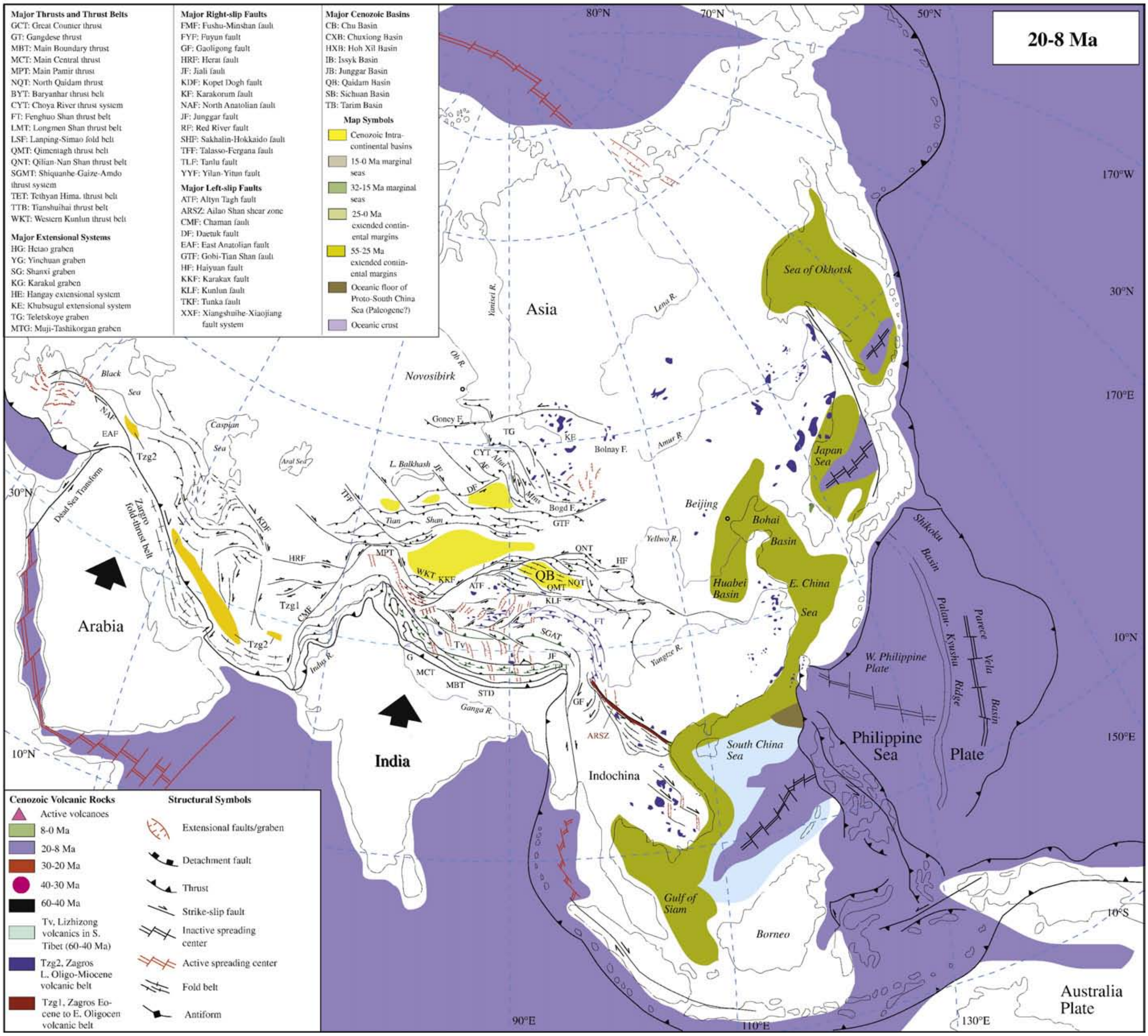

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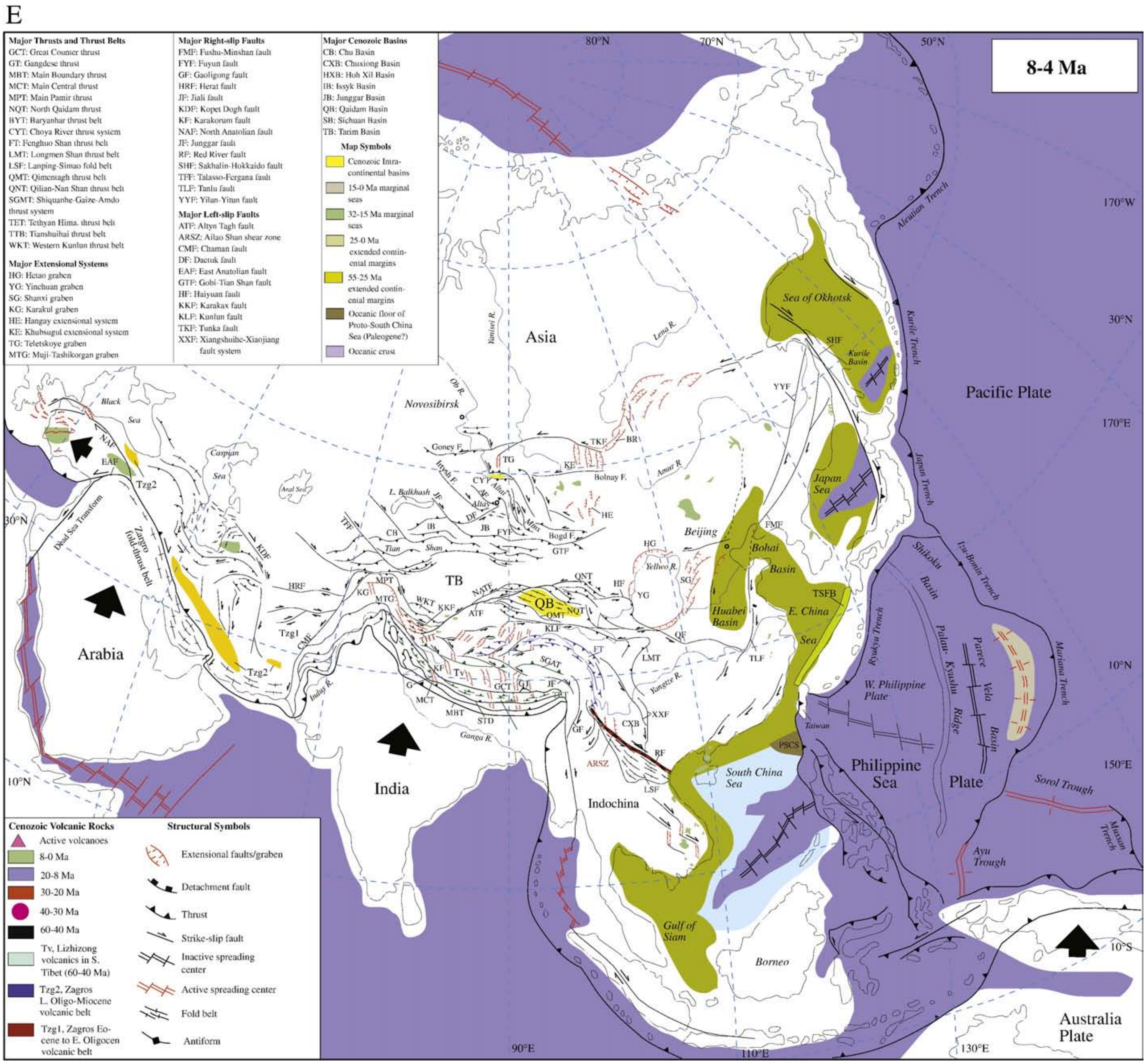
Late Cretaceous arc, upon which the Paleocene–Eocene back-arc extension occurred. The distributed mode of extension due to thermal weakening by an early event of magmatism and crustal thickening in east Asia is similar to the processes associated with mid-Tertiary to present extension in the western United States (Coney and Harms, 1984).

During the period of 65–40 Ma, intracontinental deformation of Asia is concentrated in southern and central Tibet, as expressed by the development of the Tethyan Himalayan fold-thrust belt, the Fenghuo Shan thrust belt in central Tibet, the Tianshuihai thrust belt of Cowgill et al. (2003) in western Tibet, and the southern Qilian Shan thrust belt in northern Tibet to accommodate north–south shortening. Intra-arc shortening may have also occurred across the future Turkish–Iranian–Caucasus orogen.

Between 40 Ma and 30 Ma (Fig. 4B), extension along the continental margin continues while the Philippine sea plate expands as a result of back-arc extension caused by subduction the Pacific plate. Within Asia, deformation is concentrated in central Tibet as expressed by the continuous development of the Fenghuo Shan thrust belt, the Lanping–Simao thrust belt, the Shiquanhe–Gaize–Amdo thrust system, and thrust belts associated with the Altyn Tagh fault (i.e., the western Kunlun, Qilian Shan–Nan Shan, North Qaidam, and Qimentagh belts). Igneous activities in Asia are mainly concentrated in central and eastern Tibet (Chung et al., 1998; Wang et al., 2001; Roger et al., 2001, 2003; Ding et al., 2003; Chung et al.,2005; Mo et al., 2007), Iran (Berberian and King, 1981), and western Turkey (e.g., Yilmaz et al., 2001; Dilek and Altunkaynak, 2007; Altunkaynak and Genc, 2008). Isolated basaltic eruption also occurred directly south of Lake Baikal in Mongolia (Rasskazov, 1994; Barry and Kent, 1998) and in Far East Russia (Okamura et al., 1998). The 40–30 Ma igneous rocks in central and eastern Tibet consist of syenite, trachyte, shoshonitic lamprophyre, and basaltic trachy andesite. Because of their spatial and temporal association with the Fenghuo Shan thrust belt and the subduction-related geochemical characteristics, Wang et al. (2001) proposed the occurrence of igneous rocks in the central and eastern parts of the Tibetan plateau to be related to intracontinental subduction and transpressional deformation, following the early suggestion of Deng (1989, 1998) and Arnaud et al. (1992).



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Roger et al. (2001) and Tapponnier et al. (2001) suggested a similar model for the occurrence of Eocene calc-alkaline plutonic rocks in central Tibet along the Jinsha suture (Fig. 2).

Volumetrically, 30–20 Ma is a relatively quiet period for Cenozoic igneous activity in Asia. In North China basaltic volcanism occurred in the northern end of the Shanxi and Hetao grabens and along the Tanlu, Yilan–Yitung, and Fushu–Minshan faults in northeast Asia (Fig. 4C) (R.X. Liu et al., 1992; Okamura et al., 1998; J.Q. Liu et al., 2001). Basaltic volcanic eruptions also occurred in Mongolia and the Lake Baikal region at this time. Sparse high-K volcanism continued in central and western Tibet (Deng, 1998). Igneous activity at this time completely stopped in eastern Tibet and northern Indochina, coeval with a change from transpressional to transtensional deformation along the Ailao Shan–Red River shear zone (Wang et al., 2001). Syn-collisional magmatism continued in western Turkey at this time (Aldanmaz et al., 2000; Yilmaz et al., 2001; Dilek and Altunkaynak, 2007; Altunkaynak and Genc, 2008), which was associated with the development of extensional core complexes and associated detachment faults from the end of the Oligocene to Early Pliocene (~25–23 Ma) (Okay and Satir, 2000; Isik et al., 2003; Seyitoglu et al., 2004; Catlos and Cemen, 2005).

Intraplate volcanism after 20 Ma spreads over much of east Asia (Figs. 1 and 4D). The most prominent new volcanic provinces are located along the eastern margin of Asia, extending from the mouth of the Yangtze River region in the north to the southern tip of the Indochina peninsular. This igneous activity also expands eastward into the western Pacific ocean in northern part of the South China Sea and the Taiwan region (R.X. Liu et al., 1992; Flower et al., 1998; Lee et al., 1998; Ho et al., 2003). Volcanism in Tibet is concentrated in its western part, mainly in the western Kunlun region along the Altyn Tagh fault and the Karakax fault (Fig. 1). It also occurred at the western termination of the Kunlun fault. Small volcanic fields also exist in south-central Tibet (Fig. 1) (Coulon et al., 1986; Deng, 1998; Zhao et al., 2001; Chung et al., 2005; Mo et al., 2007). In contrast, eastern Tibet (i.e., the Longmen Shan region) had no volcanic activity.

During 8-0 Ma, intracontinental volcanism in central and east Asia is distributed in the northern Sino–Korea craton (e.g., Datong volcanic field in the northern Shanxi rift; Changbai Shan volcanic center along border between China and North Korea), western Tibet, central Mongolia, and SE Asia (Fig. 4E). The above volcanic fields are without exception related spatially to active rifts, indicating strongly the causal relationship between extension and local volcanism. Sporadic volcanism also occurred across the Turkish–Iranian–Caucasus orogen in west Asia. There the volcanic fields are superposed over the earlier collision belt and their occurrence has been attributed to the breakoff of the subducted Neo-Tethyan oceanic slab (Sengor et al., 2005).

3.2. Geochemistry

Many workers noted that Cenozoic volcanic rocks in central and eastern Asia have chemical and isotopic compositions indistinguishable from those of ocean island basalts (OIBs) (e.g., Zhou et al., 1988; Basu et al., 1991; Barry and Kent, 1998). However, an important complication of such a simple observation exists in Tibet where the Paleogene magmatism appears to be associated with continental subduction while the Neogene igneous activity may have been associated with asthenospheric partial melting (Wang et al., 2001). Basu et al. (1991) examined major- and rare-earth-element (REE) concentrations and U–Th–Pb, Sm–Nd, and Rb–Sr isotope systematics for Cenozoic volcanic rocks from northeastern China. Their geochemical analyses show that these rocks are characteristically lacking the calc-alkaline suite and may have been emplaced in a rift setting. They interpreted the Sr–Nd–Pb isotopic data in terms of mixing of an EMI (enriched mantle I) component (i.e., ancient enriched mantle component, see Zindler and Hart, 1986) derived from the sub-continental mantle and a MORB (mid-ocean ridge basalt) component introduced by subduction of the Kula–Pacific Ridge beneath Asian plate in the Late Cretaceous. In southeast China, Sr–Nd–Pb data indicate an EM2-type lead isotopic signature (Flower et al., 1998). Zou et al. (2000) suggested that the basalts from southeast China are a result of mixing between asthenosphere and EM2 whereas basalts from northeast China are a

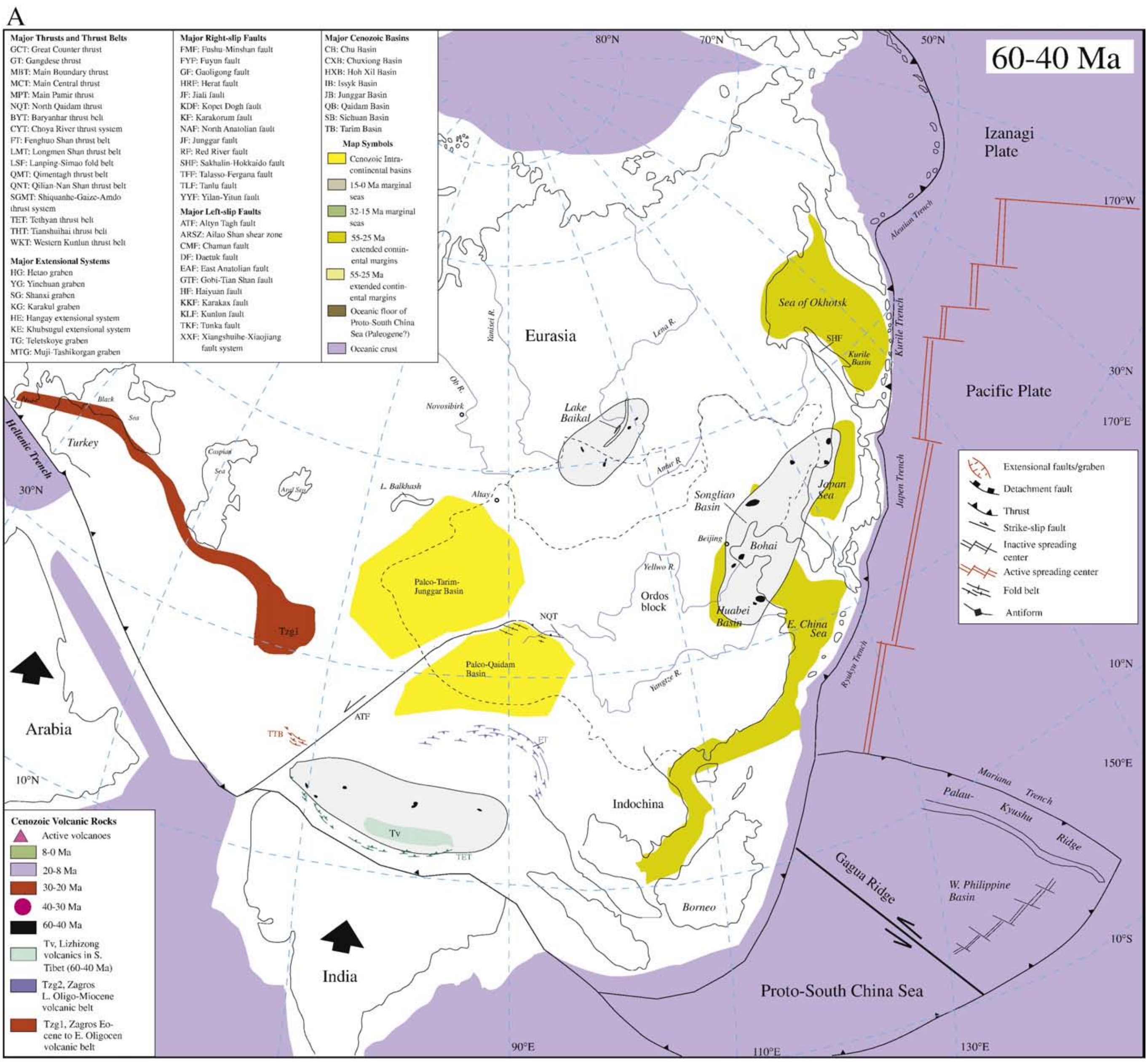
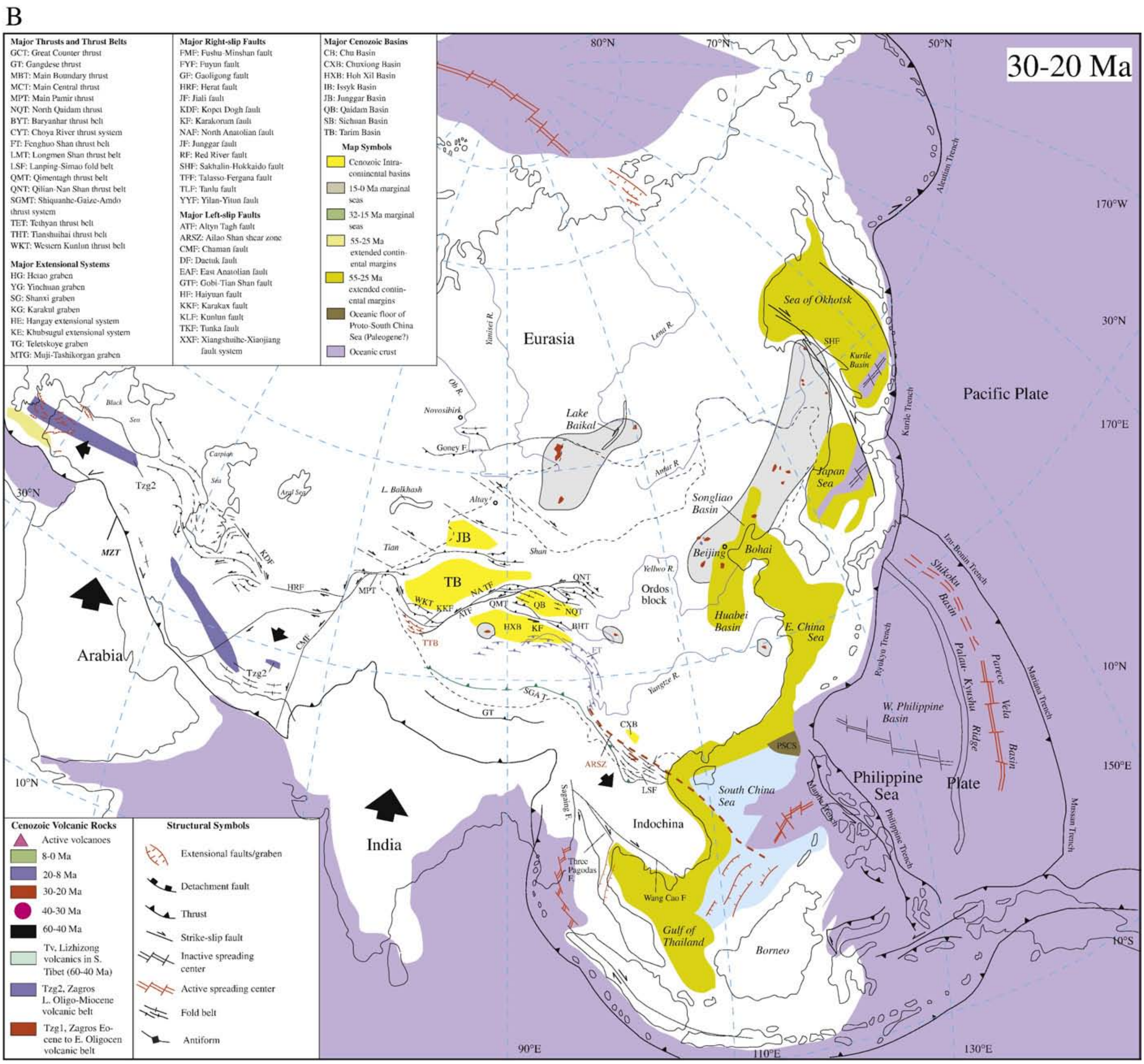


Fig. 5. Preliminary palinspastic reconstruction of Asia during the periods of (A) 60–40 Ma, (B) 30–20 Ma, and (C) 10–5 Ma.



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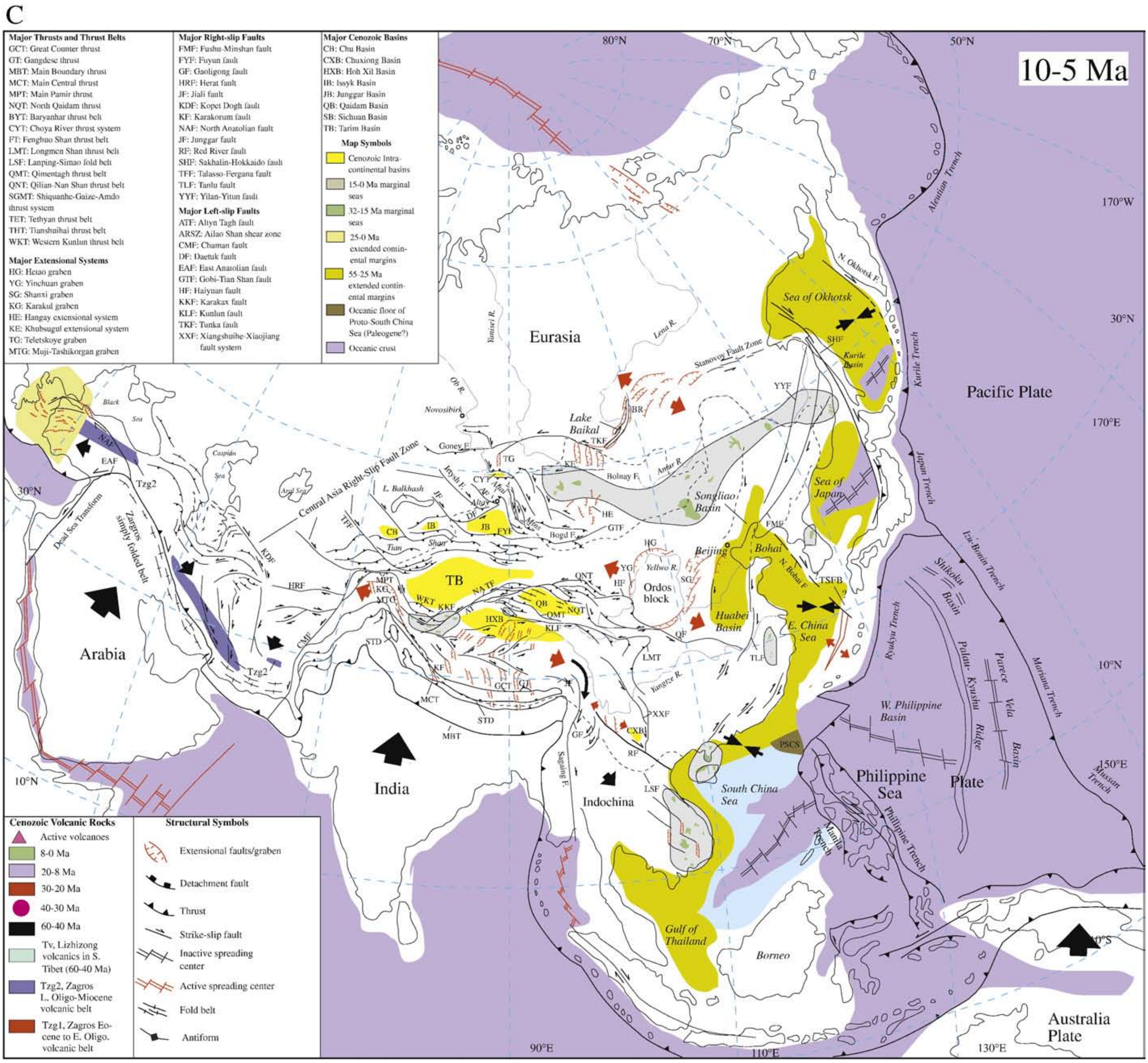
result of mixing between asthenosphere and EM1. More recently, Barry et al. (2003) used helium isotopes to determine if widespread alkali basaltic volcanism that occurred over a vast area of central Asia from Siberia to North China in the past 30 m.y. was related to mantle plumes as proposed by Windley and Allen (1993). Barry et al. (2003) showed that the maximum $^3He/^4He$ ratios in Siberia and Mongolia are in the range of 8.1–9.5, overlapping with $^3He/^4He$ ratios for MORB. This similarity argues against significant involvement of lower mantle sources for the generation of the volcanism. Barry et al. (2003) proposed that the basaltic magmatism in central Asia could have been driven from (1) replacement of delaminated mantle lithosphere by shallow asthenosphere, (2) thermal blanketing effect in the interior of a large continental mass, and (3) large scale mantle disturbance caused by protracted Mesozoic subduction below Asia. The above results and inferences are consistent with the lack of deep-rooted low seismic-wave velocity anomalies in the upper mantle beneath eastern and central Asia (Boschi and Ekstrom, 2002; Zhao, 2001).

Cenozoic volcanism in Turkey exhibits similar geochemical characteristics to those in Tibet and indicates a transition from a continental-subduction dominated setting from the Eocene to the Late Miocene to an asthenospheric melting setting in the Pliocene–Quaternary (Dilek and Altunkaynak, 2007; Ersoy et al., 2008). The temporal change in petrology and geochemical characteristics of Cenozoic volcanism in Turkey has been attributed to slab breakoff, delamination of mantle lithosphere, and asthenospheric upwelling (Sengor et al., 2005; Dilek and Altunkaynak, 2007).

4. Preliminary palinspastic reconstruction of Cenozoic Tectonic history of Asia

Since the publication of the landmark paper of Molnar and Tapponnier (1975), eastern and central Asia has become the focal point of research on continental dynamics (e.g., Tapponnier et al., 1982; England and Houseman, 1986; Kang and Bird, 1996). However,



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the past research has mostly focused on the Cenozoic geologic history of the Tibetan plateau and its surrounding regions (e.g., Burchfiel et al., 1991; Yin and Harrison, 2000; Hodges, 2000; Tapponnier et al., 2001; Yin, 2006), with no attempt in examining possible interactions between Arabia–Asia and India–Asia collision in creating the deformation pattern of whole Asia. This subject was only briefly mentioned in Berberian and King (1981).

Using the data shown in Fig. 4 as constraints, three stages of Cenozoic deformation of Asia at 50–40 Ma, 30–20 Ma, and 10–5 Ma are shown (Fig. 5). These time periods are chosen to illustrate how the deformation style in Asia has evolved rather than providing an exact reconstruction that would strongly depend on the assumptions about how fault-bounded blocks have deformed during the course of India–Asia collision (e.g., Replumaz and Tapponnier, 2003). In the following reconstructions, the work of Dewey et al. (1989) is used for the kinematic history of the India plate and that of Besse and Coutillot (1991) and McQuarrie et al. (2003) for the kinematic history of the Arabian plate. The plate evolution of the Pacific Basin in the Cenozoic follows the work of Hall (2002), Smith (2007) and Whittaker et al. (2007).

4.1. 60–40 Ma

Between 60 Ma and 40 Ma, deformation in Asia is mainly concentrated in front of the northward indenting Indian continent (Fig. 5A). Deformation is characterized by the development of the Tethyan and Fenghuo Shan thrust belts and thrust systems around the Qaidam and Hoh Xil basins. The jump of deformation front across the Qaidam and Hoh Xil basins at the onset of the India-Asia collision zone created the Altyn Tagh fault that terminates in the southwest at the Tianshuihai thrust belt and in the northeast at the Qilian Shan thrust belt (Yin et al., 2002; Cowgill et al., 2003; Yin et al., 2007a; Yin et al., 2008a,b). Along the eastern margin of Asia, widely distributed extension caused thinning of the continental crust and synchronous eruption of basalts in northeast Asia. This

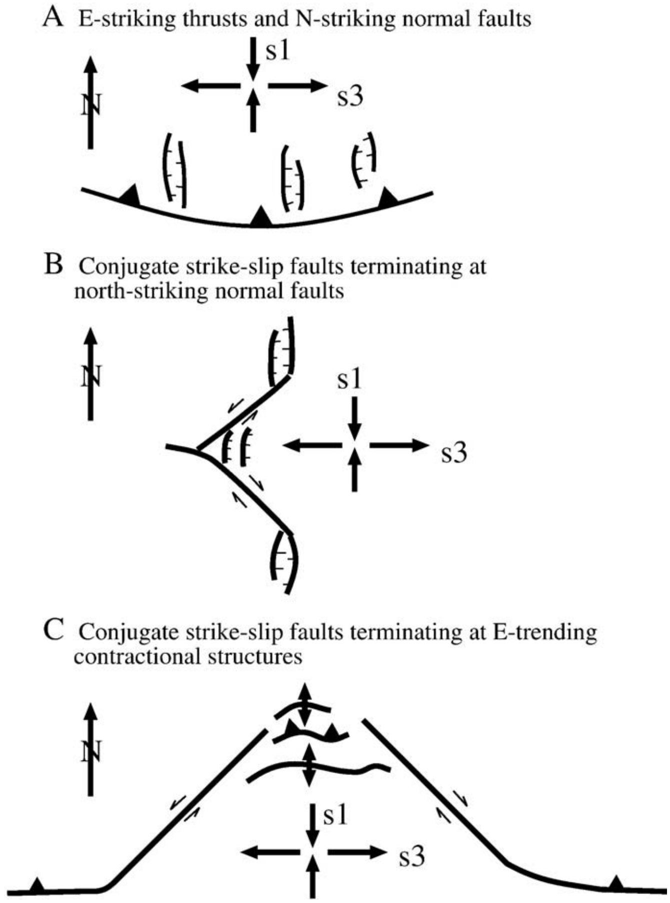


Fig. 6. Structural associations accommodating constrictional strain field in the Neogene. (A) Simultaneous development of crustal-scale thrusts and lithospheric scale normal faults modified from Yin (2000), (B) Conjugate strike-slip faults terminate at extensional fault systems, and (C) conjugate strike-slip fault systems terminate at contractional faults systems.

extensional event was probably related to eastward migration of the Western Pacific trench system as indicated by eastward younging of volcanic arcs from Late Cretaceous to early Tertiary time (e.g., Ren et al., 2002a). As a result of rapid eastward migration of the trenches, broken oceanic slabs of the Pacific plate may have been left behind beneath east Asia; the slabs may have been sinking downward, inducing upwelling of the asthenosphere and volcanism at the surface. The remnants of the subducted Pacific slab fragments may be expressed by the northeast-trending high P-wave velocity anomalies in the western Pacific at depths between 100 km and 400 km (Zhao, 2001).

In the southwestern Pacific, subduction of the India–Australia plate caused back arc spreading and the formation of the western Philippine basin. Subduction of the North New Guinea plate of (Seno and Muruyama, 1984) and the mid-ocean ridge that separates it from the Pacific plate. Ridge subduction may have created voluminous eruption of boninite along the now Palau–Kyushu ridge (Stern and Bloomer, 1992). The opening of the western Philippine basin may have occurred on top of a Late Cretaceous oceanic crust (i.e., the Huatung basin east of Taiwan). The Gagua ridge may have served as a transform fault to separate the Cretaceous oceanic crust from the Paleogene Philippine back arc basin. In the northwestern Pacific, the Izanagi–Pacific ridge was being subducted below the Aleutian and Kurile trenches, followed by rapid counterclockwise rotation of the Pacific plate relative to the Eurasia plate (Whittaker et al., 2007; Smith, 2007).

4.2. 30–20 Ma

The period of 30–20 Ma marked two important events in the inland of Asia. First is the development of large strike-slip faults that assist extrusion of the Afghanistan and Indochina blocks on both sides of indenting Indian continent (Fig. 5B). Second, the contractional deformation front in the India-Asia collision zone migrates northward across the relatively rigid Tarim block, which is expressed by the formation of the Tian Shan orogen. In the eastern part of the India–Asia collision zone, the left-slip Ailao Shan–Red River shear zone and its associated Lanping–Simao fold belt and the Three Pagodas fault were all active (Leloup et al., 1995, 2001; Wang et al., 1998). Along the western edge of the Indochina block, the Wang Cao and Three Pagodas faults changed its slip direction from left-slip to right-slip at about 30 Ma during progressive penetration of India into Asia (Lacassin et al., 1997). Right-slip motion on these faults together with left-slip motion on the Ailao Shan shear zone caused southeastward extrusion of the Indochina block. Extrusion tectonics resulted in localized high-magnitude extension in the South China Sea and creation of oceanic floor underneath. Because coeval extension also occurred in front of the extruding Indochina block, southeastward motion of Indochina and extension in southeast Asia were also possibly facilitated by eastward migration of the trench system in the western Pacific due to a decrease in plate convergence rate (Northrup et al., 1995; Royden et al., 2008). At ~30 Ma, deformation at the western termination of the Altyn Tagh fault expanded northward to incorporate the western Kunlun thrust belt (Yin et al., 2002; Cowgill et al., 2003). This structure linked to the Main Pamir thrust to the east, which in turn was connected with the left-slip Chaman fault at the western end of the Main Pamir thrust system. Farther to the west, motion on the left-slip Chaman fault and the right-slip Herat fault (Fig. 1) caused south-westward extrusion of the Afghanistan block (Tapponnier et al., 1981).

The left-slip Kunlun transpressional system was also initiated at the time of 30–20 Ma, causing the partitioning of the Paleo–Qaidam between the Qilian Shan and the Fenghuo Shan thrust belts into the Hoh Xil basin to the south and the modern Qaidam basin to the north (Yin et al., 2008b). The Hoh Xil basin with an average elevation of ~5 km is about 2 km higher than the Qaidam basins that has an average elevation of ~2900 m. As the Hoh Xil region did not experience any significant upper crustal shortening since the two basins began separated at 30–20 Ma, the elevation gain for the Hoh Xil basin must have occurred either in the lower crust or the mantle. Yin et al. (2008b) suggested the Neogene–Quaternary Hoh Xil uplift was induced by the development of a flake tectonic system across the Kunlun transpressional system, with the Qaidam lower crust and mantle lithosphere subducted below the Hoh Xil upper crust. Alternatively, the Hoh Xil area could have been raised by convective removal of the mantle lithosphere, which had created a dynamic support to raise the region (Molnar et al., 1993). This argument is currently favored by the author as recent seismic imaging below southern Tibet strongly suggests that the Hoh Xil mantle lithosphere had been delaminated in the Early Miocene that is currently trapped at the base of the upper mantle at a depth of ~660 km below central India (Tseng and Chen, 2008).

The northern deformation front of the India–Asia collision zone may have propagated across the Tarim block and created the southern Tian Shan thrust belt at about 24–20 Ma (e.g., Yin et al., 1998) or later at about 8–11 Ma (e.g., Molnar et al., 1993; Bullen et al., 2001). Meanwhile, the thrust system along the northern edge of the Qaidam basin also expanded northward to form the Qilian Shan–Nan Shan thrust belt (Yin et al., 2002). It is possible that the contractional deformation had propagated all the way to the Lake Baikal region at this time interval, causing transpressional deformation in the Late Oligocene and Early Miocene across southern Siberia (Delvaux et al., 1997).

In west Asia, the initial collision at 40–20 Ma between Arabia and Eurasia resulted in the development of the Zagro thrust belt and compression in and around the southern Caspian Sea (e.g., Sengör et al., 1985; Sengör and Natalin, 1996). Within the Himalaya, the Main Central Thrust and the South Tibet Detachment began to develop (see review by Yin, 2006). The Gangdese thrust in southern Tibet and the Main Central Thrust in the Himalaya are the main contractional structures to accommodate north–south contraction between India and Asia in the Late Oligocene and Miocene (e.g., Yin et al., 1999a; Murphy and Yin, 2003; Robinson et al., 2003).

In the west Pacific, the North New Guinea plate is completely consumed by subduction at this time. Subduction of the Pacific plate beneath the Philippine Sea plate created Shikoku and Parece Vela Basins. Because the Shikoku–Parece Vela ridge is partially subducted, its original size could have been much larger across the northern Pacific ocean and may have even intersected the Izu–Bonin trench to form a triple junction near the present Sea of Okhotsk. The counter-clockwise rotation of the Pacific plate relative to Asia continued at 30–20 Ma. Whether this relative rotation had caused left-slip shear deformation along the eastern margin of Asia is not clear, as the position of the Phillipine plate was uncertain and it could have been located between the Asia and Pacific plates at the time. As noted by previous workers, the spreading centers of these back-arc basins, where oceanic crust was created, are oblique to the trenches (Fig. 1) (Ren et al., 2002a). This relationship suggests that slab retreat alone cannot adequately explain the segmented nature and rectangular shapes of the back-arc basins. We will discuss this point in Section 5.3 by suggesting that the back arc basins were most likely produced by a margin-parallel right-slip shear deformation. This inference contradicts the common believe that the Pacific and Asia plates had an oblique right-slip convergence in the middle Tertiary (e.g., Ren, 2002a).

4.3. 10–5 Ma

Intracontinental deformation at this time is characterized by widespread east–west extension associated with north–south contraction (Fig. 5C). This mode of deformation is commonly described as constrictional strain field and is expressed in the following three types of fault association. (1) East–west extension accommodated by north-striking normal faults occurs in the same region with north–south shortening accommodated by east-striking thrusts. This kind of fault association is best developed in the Himalaya where contractional deformation along the Main Frontal and Main Boundary thrusts was coeval with the north-striking rifts (Mercier et al., 1987; Yin, 2000) (Figs. 1 and 6A). (2) Coeval north–south contraction and east–west extension are accommodated by strike–slip faults terminating at north-trending rifts. This mode of deformation is best developed in central Tibet north and south of the Bangong–Nujiang suture (Taylor et al., 2003) and in western Mongolia (Figs. 1, 2, and 6B). The third type of fault association accommodating constrictional strain field is by the development of conjugate strike–slip faults that terminate at north-trending contractional structures. This type of structures are best developed in the Junggar basin where the left-slip Daertuk fault

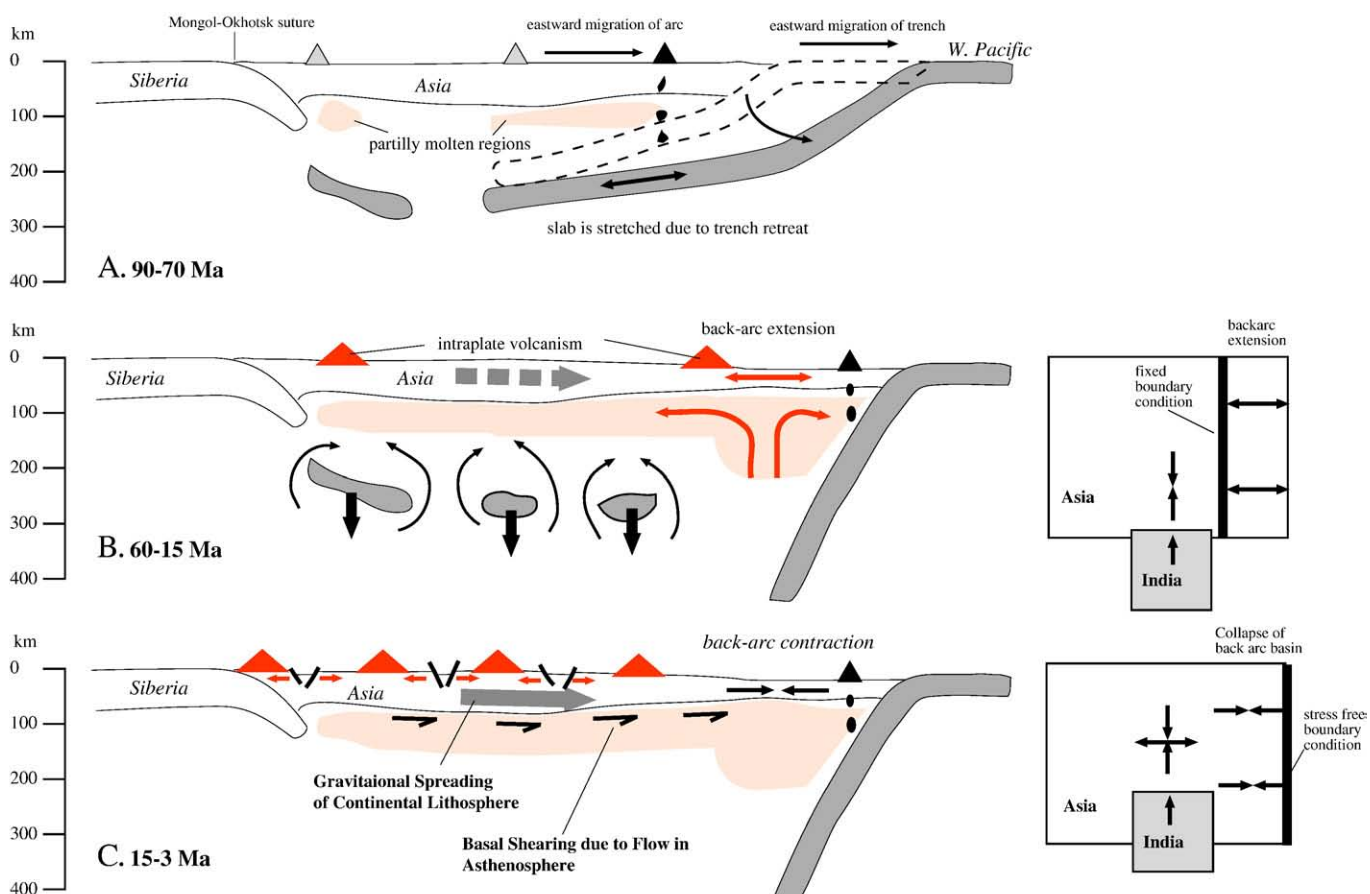


Fig. 7. Possible tectonic processes that have led to the occurrence of Cenozoic volcanism in Asia and westward migration of extensional domains from Paleogene along the continental margin to the Neogene in the interior of Asia including Tibet. (a) 90–70 Ma. Slab breakoff during closure of the Mongol–Okhotsk ocean; beginning of upwelling and partially melting of the upper mantle that generated volcanism. Trench system in the western Pacific retreated rapidly, causing the collapse of subducted Pacific plate and generation of volcanism and extension along continental margin. (B) 60–15 Ma. Continuing back-arc extension along continental margin and north–south contraction in the continental interior. The upwelling of mantle flow along the continental margin may have prevented eastward gravitational spreading of the Asian lithosphere. (C) 15–3 Ma. Extension and associated mantle upwelling ceased along the eastern edge of Asia, allowing Asia to move towards causing widespread contraction along the continental margin.

and the right-slip Fuyun fault terminate at a series of north-trending folds in the northern Junggar Basin and within the western Altai Shan (Figs. 1 and 6C).

Within inland Asia, east–west extension is accommodated by the following extension systems. From south to north, they are the Tibetan rift system, the Ordos rift systems that include the Yichuan, Hetao, and Shan graben systems in northern China, the Hangay extensional system in Mongolia, and the Baikal rift system in southeast Siberia. Although all these extensional structures are currently active, it is not clear whether they were initiated synchronously or diachronously. However, this information may provide a key test to whether east–west extension was related to the India–Asia collision or induced by subduction along the western Pacific. For example, if initiation of north-trending rift systems had propagated northward, it would strongly suggest that northward motion of the indenting Indian plate has been the main driving force for their development. Alternatively, if the rifts were all initiated simultaneously, it is more likely that their formation was induced by a change in boundary condition along the eastern margin of the continent (Yin, 2000; cf., Davy and Cobbold, 1988).

5. Discussion

5.1. Cenozoic Volcanism

From above descriptions and the available geologic data, the following generation of the Cenozoic tectonics of Asia can be made. (1) Extension is concentrated along the continental margin of Asia in the Paleogene and early Neogene. This event was closely associated with basaltic eruptions and resulted in considerable crustal thinning of the eastern margin of Asia and the eventual formation of several back arc basins floored by oceanic crust. (2) Since the mid-Miocene, extension has shifted to the inland region of Asia from the Baikal rift system to the north to the Tibetan rift system in the south. Although Neogene rifts in Asia are generally correlated with the Cenozoic igneous provinces, the initiation of extension is at least 10–30 Ma younger than the earliest volcanic eruptions near the rift systems. (3) Coeval with the Neogene to present intracontinental east-west extension in the interior of Asia is the widespread east-west contraction in the western Pacific along the eastern edge of Asia; the latter is expressed by initiation of new subduction zones and consumption of existing back arc basins (e.g., subduction of Japan Sea and South China Sea below nearby arcs). (4) Intraplate volcanic eruptions occurred nearly continuously in all major igneous provinces since the beginning of the Cenozoic, regardless of variation of plate boundary conditions along the margins of Asia. (5) Cenozoic intraplate igneous provinces are spatially closely associated with the Jurassic and Cretaceous arcs and their corresponding suture zones. (6) Immediately after the initial collision between India and Asia and between Arabia and Asia, a large amount of volcanic materials were erupted over a wide region of the arcs immediately predating the collision.

In Tibet, the origin of syn-collisional volcanism has been related to (1) continental subduction (Deng, 1989; Arnoud et al., 1992; Meyer et al., 1998; J.H. Wang et al., 2001), (2) convective removal of thickened mantle lithosphere (England and Houseman, 1989; Turner

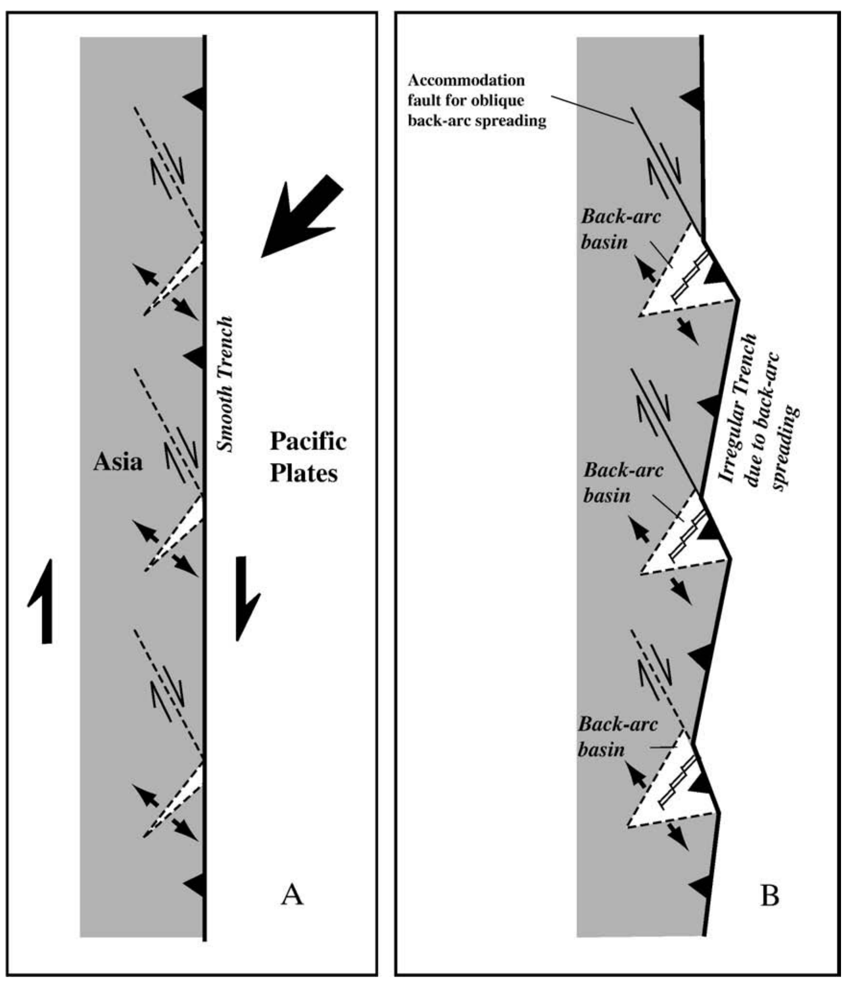


Fig. 8. A Riedel shear model for the formation of back-arc basins in Asia during oblique subduction of the Pacific plate beneath Asian plate between 30 Ma and 10 Ma. Note that this model explains the spacing and shaped of the back arc basins and oblique angles between the spreading ridges in the back-arc basins and the trenches.

et al., 1993; Chung et al., 2005), (3) extension induced by development of pull-apart basins along major strike-slip faults (Cooper et al., 2002), (4) rollback of subducted Tethyan oceanic slab (Ding et al., 2003), and (5) lithospheric-scale rifting (Yin, 2000). Southern Tibet volcanism occurred continuously since the beginning of the Tertiary and consistently exhibits geochemical signatures of subduction-related magmatism. Thus, the protracted volcanism may be related to upwelling of upper mantle flows induced by the breakoff of Indian oceanic lithosphere (Yin and Harrison, 2000; Kohn and Parkinson, 2002).

In contrast to diverse hypotheses for Tibetan igneous activity, intraplate volcanism in central and east Asia has been consistently related to rifting (Ye et al., 1987; Zhou et al., 1988; J.Q. Liu et al., 2001; Ho et al., 2003). The rifting in turn is interpreted to be induced by asthenospheric flows either caused by the India–Asia collision or produced by subduction of the Pacific plate (Flower et al., 1998; Basu et al., 1991). The proposal that Cenozoic volcanism in Asia was caused by lateral extrusion of Asian asthenosphere due to indention and subduction of the Indian continent (Flower et al., 1998) encounters the difficulty of explaining the existence of a broad zone of magmatic gap in central Asia (Fig. 3). However, this difficulty may be overcome by the possible lack of permeability structure in central Asia, which has prohibited the magma to surface. Relating Cenozoic volcanism in central Asia to rifting alone also faces the some prominent problems. Volcanism in regions where rifts are present initiated at least some 20–30 Ma earlier than the start of extension. For example, the Fanshi volcanic rocks in the northernmost segment of the Shanxi rift system were erupted at 40–38 Ma (R.X. Liu et al., 1992). However, rifting in the region did not initiate until the Pliocene at about 5–3 Ma (Wang et al., 1996). (3) If rifting caused volcanism, partial melting of mantle lithosphere due to decompression is required and should have occurred after the initiation of the rift. One possible explanation is that widespread crustal thinning in the Paleogene across eastern and central Asia may have induced partial melting below east Asia (Ye et al., 1987). The subsequent small-magnitude extension (<10 s km) across the Shanxi and Baikal rifts (Zhang et al., 1998; Mats et al., 2000) could have served as conduits to guide the migration of the partial melts to the surface along the rift zones.

Cenozoic volcanism in Asia is spatially correlative to the Jurassic–Cretaceous suture zones and magmatic arcs immediately predating the India–Asia and Arabia–Asia collision (Fig. 3). It is possible that asthenospheric upwelling was induced by downward sinking motion of the subducted Pacific, Neo-Tethyan and Mongolo–Okhotsk ocean slabs in the late Jurassic and Cretaceous. The asthenospheric upwelling caused heating and thus partial melting of the overlying mantle lithosphere, which is independent of lithospheric deformation. This mechanism not only explains the nearly continuous igneous activities throughout the Cenozoic in Mongolia, North China, South

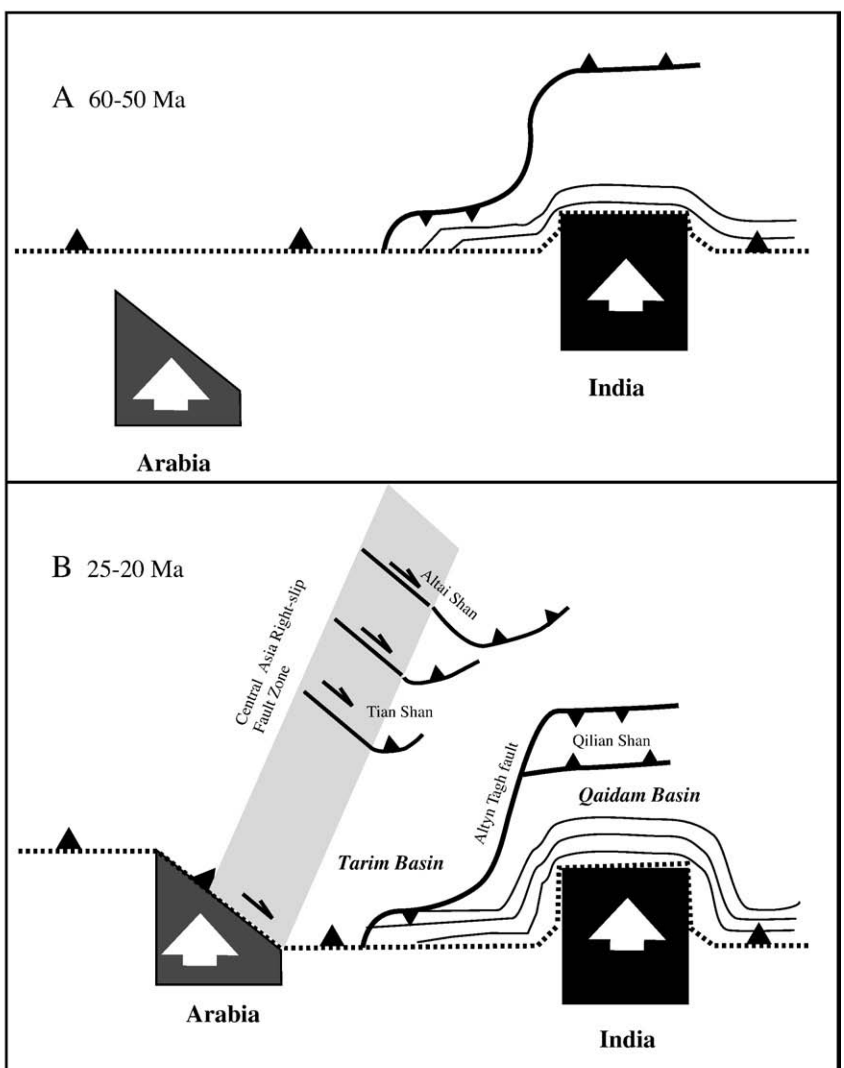
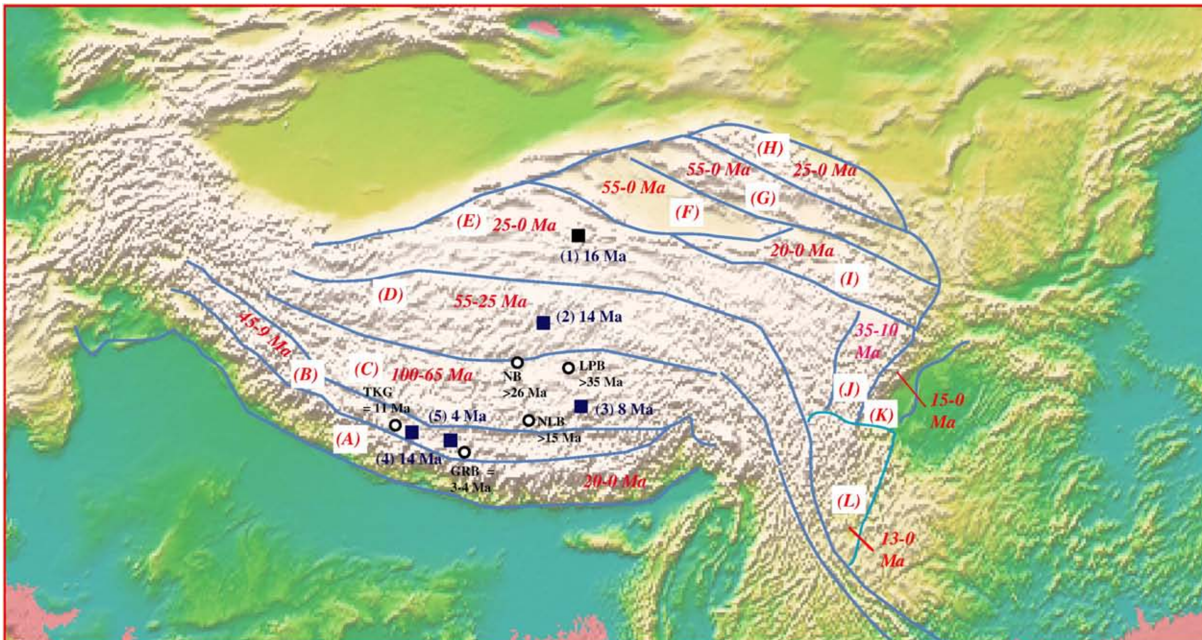


Fig. 9. A model for the development of the Central Asia right-slip Fault Zone induced by oblique collision between Arabia and Asia. The formation of the right-slip faults is interpreted to be the cause for the formation of the Tian Shan and Altai Shan intracontinental orogens in central Asia. There, the east-trending thrusts acted as the termination structures of the right-slip faults.

China, and Tibet, but also accounts for spatial correlation between the Mesozoic arcs and the Cenozoic volcanic provinces in Asia. Specifically, the following history is proposed. (1) During the Late Cretaceous (90–65 Ma), closure of the Mongol–Okhotsk ocean (e.g., Sengör and Natalin, 1996; Yin and Nie, 1996) caused the breakoff of an oceanic slab. The sinking and attenuation of the slab caused upwelling and partially melting of the upper mantle that generated volcanism in Mongolia, southern Siberia and Chinese Altai. Along the eastern margin of Asia, the Late Cretaceous trench system retreated rapidly eastward, leaving broken fragments of subducted Pacific plate behind beneath east Asia. The broken slabs sank slowly and induced an upward flow to fill the space left behind the sinking slabs (Fig. 7A). (2) In the Paleogene and Early Miocene (65–15 Ma), the eastern margin of Asia was dominated by widely distributed extension and development of back-arc basins (Fig. 7B). (3) At about 15 Ma, extension ceased throughout the eastern margin of Asia, as expressed by the initiation of subduction along the Manila trench and the contraction and subduction of the South China Sea (Sibuet et al., 2002). Sea-floor spreading in the Japan Sea, South China Sea, and Shikoku–Parece Vela basin in the eastern Philippine



Names of the regions and main sources of information:

(A) Crestal and Northern Himalaya (also known as Tibetan or Tethyan Himalaya) (Ratschbacher et al., 1994; Harrison et al., 2000; Wiesmyr and Grassmann, 2002; Aikman et al., 2008).
(B) Southern Himalaya, including mostly Greater and Lesser Himalayan Units (see review by Yin, 2006)
(C) Lhasa Terrane (Yin et al., 1994; Murphy et al., 1997; Kapp et al., 2007; DeCelles et al., 2007a)
(D) Paleogene Fenghuo Shan-Nanqiang-Simao Thrust Belt (Spurlin et al., 2005; Horton et al., 2002)
(E) Hoh Xil Basin (Liu et al., 2001; Yin et al., 2008a)
(F) Qaidam Basin (Zhou et al., 2006; Yin et al., 2008a,b)
(G) Southern Qilian Shan-Nan Shan thrust belt (Yin et al., 2008a,b)
(H) Northern Qilian Shan-Nan Shan thrust belt (Yin et al., 2002 and references therein)
(I) Burhan Buda Mountains and Anyemaqen Shan (Yuan et al., 2006)
(J) Western Longmen Shan thrust belt (Hames and Burchfiel, 1993)
(K) Eastern Longmen Shan thrust belt (Kirby et al., 2002)
(L) Southeast Tibetan plateau (Clark et al, 2005)

Names of the basins and main sources of references:

LB: Lunpola Basin (Rowley and Currie, 2006)
NB: Nima Basin (DeCelles et al., 2007b)
NLB: Namling Basin (Spicer et al., 2003)
GRB: Gyirong Basin (Wang et al., 2006)
TKG: Thakkhola graben Garzione et al. (2000)

Main sources of information on the ages of rifts

1. Jolivet et al. (2003)
2. Blisniuk et al. (2001)
3. Harrison et al. (1995)
4. Coleman and Hodges (1995)
5. Maheo et al. (2007)

Map Legends

(D) 55-25 Ma — Deformation zone and duration of uplift as inferred from either crustal shortening events or mantle thermal events.

LB > 35 Ma — Study location and age of uplift as inferred from stable isotope analyses or paleo-botanical analysis.

(3) 8 Ma — Study location and initiation age of Tibetan rifts used as proxy for the time when the Tibetan plateau reached its current elevation.

Fig. 10. Summary of spatial distribution of uplift history of the Tibetan plateau (Clark et al., 2005; Hames and Burchfiel, 1993; and Zhou et al., 2006).

Sea plate also terminated at this time. The elimination of the mantle upwelling along the continental margin switched the eastern edge of Asia from a fixed boundary condition (i.e., a boundary condition that prevents the Asian lithosphere to spread eastward) to a stress-free boundary condition (i.e., a condition that allows the Asian lithosphere to spread eastward). It is the stress free boundary condition that has allowed the regions, which had already been thermally weakened and tectonically thickened by protracted arc magmatism and intra-arc contraction to spread eastward towards the trench due to its high gravitational energy. The eastward spreading of the Asian continent in conjunction with north-south contraction induced by the India–Asia collision produced the widespread Neogene constrictional deformation in Asia (Fig. 7C).

5.2. Neogene–Quaternary East–West Extension in Asia

East–west extension in Tibet has long been regarded as a result of either gravitational collapse of a thickened lithosphere or a sudden increase in gravitational energy due to a thermal event in the mantle induced by convective removal of the underlying mantle lithosphere or continental subduction (Molnar and Tapponnier, 1978; England and Houseman, 1989; Tapponnier et al., 2001). The similar timing, mode of extension, and direct kinematic linkages between rifts in Tibet and North China suggest that east–west extension occurred across whole Asia rather than restricted to Tibet alone. This suggests that a change in plate-boundary conditions along the eastern margin of Asia may have been the cause for the Neogene–Quaternary extension in Asia. The above explanation faces the problem of how stress being transmitted some 1000 km away from the western Pacific trench system. By simulating Quaternary fault kinematics and strain distribution in Asia, Kong and Bird (1996) found that the eastern and central Asian lithosphere is currently spreading eastward via distributed deformation, resulting from the differences in gravitational potentials between the continent and the trench. In the context of this model, lithospheric thickening due to India–Asia collision may have progressively built up the gravitational potentials in Asia that eventually overcame the resistance of the trenches for eastward spreading of the continent (Fig. 7). In addition, slab breakoffs, convective removal of the mantle lithosphere, and astehnosphereic flow could also increase local gravitational potential that contributes to the overall driving forces for lateral spreading and distribution extension of the Asian lithosphere (e.g., Dilek and Altunkaynak, 2007).

5.3. Primary shears and opening of back-arc basins in the Western Pacific

The opening of some of the back-arc basins in the western Pacific can be clearly attributed to the trench retreat, which has produced back-arc spreading centers parallel to the trenches. Examples of this type of basins include the Philippine Sea basin and Parece Vela basin behind the Marina Trench (e.g., Hall, 2002) (Fig. 1). However, this mechanism does not explain adequately the semi-rectangular shaped Japan Sea, the Sea of Okhotsk, and Bohai Bay basin, all of which have their long basin axes oriented at an oblique angle to the strike of the local subduction zones (Fig. 1). For the first two basins, the orientations of the spreading centers are also at an oblique angle to the nearby trenches and the basins are bounded by northwest-trending right-slip faults (Fig. 1). It is possible that the eastern margin of Asia experienced oblique right-slip shear, which had produced a series of synthetic primary shear zones (i.e., P shears) that were linked with spreading centers. (Fig. 8A). Motion on the strike-slip faults led

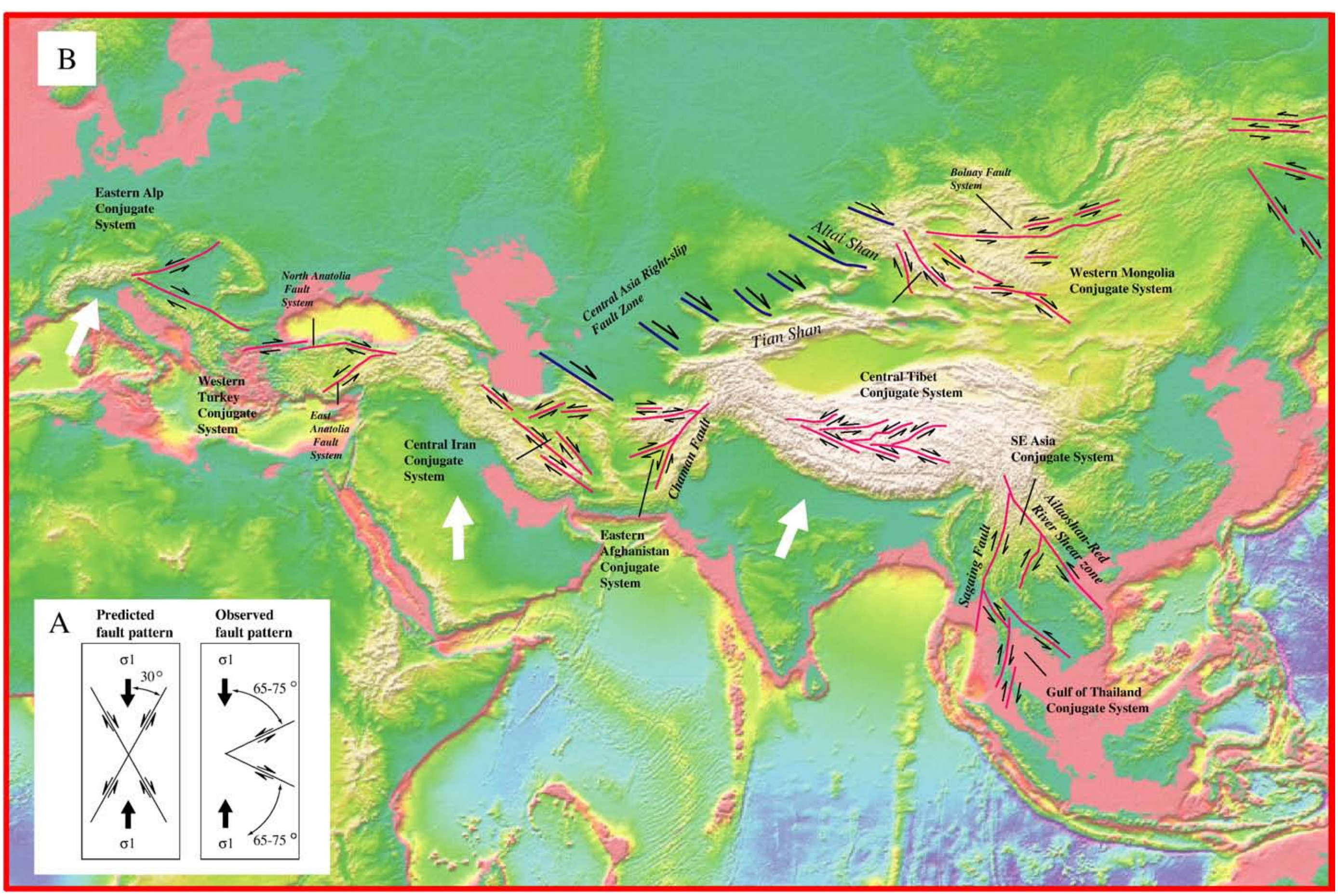


Fig. 11. Distribution of Cenozoic V-shaped conjugate strike-slip faults in Asia.

to the development of rectangular-shaped back-arc basins with their long axes obliquely oriented with respect to the strike of the nearby trench system (Fig. 1). This mechanism explains well the spacing, shape and orientation of the back-arc basins along the eastern Asia margin. However, it is not consistent with the inferred counter-clockwise rotation of the Pacific plate relative to the Asia plate in the mid-Tertiary during the opening of most western Pacific marginal seas. This consistency may imply more complicated possibilities in paleoplate reconstructions of the Pacific regions than the currently available models.

5.4. Effects of combined India–Asia and Arabia–Asia collision on Asian deformation

Cenozoic deformation of central and eastern Asia has been traditionally viewed as a result of India–Asia collision (Molnar and Tapponnier, 1975). However, as the western part of the central Asia deformation systems (e.g., the Tian Shan and Altai Shan) are spatially closer to the Arabia–Asia collisional front, it is possible that some Cenozoic strain in central Asia was induced by northward penetration of Arabia into Asia. The most convincing evidence to link deformation of central Asia with Arabia–.Asia collision is the development of series of right-slip fault zones that are approximately evenly spaced and trend in the northwest direction. This fault zone termed Central Asia Right–Slip Fault Zone in this study can be traced from the Main Zagro Thrust that has an oblique right-slip component to the Altai right-slip transpressional system in western Mongolia (Fig. 1). Delvaux et al. (1997), following the early proposal by Davy and Cobbold (1988), suggested that this fault zone resulted from development of a broad left-slip shear zone with the right-slip faults to accommodate counterclockwise vertical-axis rotation in a bookshelf fashion. Alternatively, it is also possible that the individual right-slip faults were induced by a broad right-slip shear caused by indentation of Arabia into Asia (Fig. 9). In the context of this model, the right-slip faults were induced by oblique convergence between Arabia and Asia; the shear zones are terminated at thrust belts or transpressional systems such as the Tian Shan and Altai Shan intracontinental orogens in central Asia (Fig. 1). In the context of this model, the right-slip faults serve as transform faults linking east-trending thrust belts that have increasing amounts of shortening southwestwards approaching the Zagro convergence front. The dying-out of the northwest-striking right-slip faults into the interior of Asia in Kazakhstan is similar to the dying-out oceanic transform faults away from the linked mid-ocean ridges. The above interpretation contrasts sharply to the traditional view that the Tian Shan range was created by the India–Asia collision alone (e.g., Molnar et al., 1993).

5.5. Uplift history of the Himalayan–Tibetan orogen

Although determining the elevation history of the Himalayas and Tibetan plateau has important implications for the dynamics of continental deformation and global climate change (e.g., Harrison et al., 1992; Molnar et al., 1993), achieving this goal is challenging as all the existing methods are unable to measure the past elevation of the plateau directly. Five approaches have been commonly adopted. The first is to relate lithospheric shortening or thermal events in the mantle to elevation gains. That is, the time of initiation and termination of a Cenozoic shortening event in a region can be regarded as the time of initial uplift and the time when the region reached its current elevation, assuming no thermal effects from the mantle and erosion did not outpace crustal thickening. For example, crustal thickening in the southern Qilian Shan thrust belt of northern Tibet started at about 55 Ma and remains on-going (Yin et al., 2008b). The region has an average elevation of about 3500–4000 m and the elevation will continue to increase. On the other hand, the Fenghuo Shan–Nangqian thrust belt in central Tibet was developed between 55 Ma and 20 Ma (e.g., Liu et al., 2001), which implies that the region had gained its current elevation by ~20 Ma if thermal effects from the mantle did not play a role. Ideally, if we know the mode of deformation (i.e., simple vs. pure shear), the magnitude of crustal shortening, and the magnitude of erosion, we can obtain the amount of elevation gain assuming Airy isostasy (e.g., Yin et al., 1998). In reality, the thickened crust is also supported by the elastic strength of the lithosphere and as this parameter is not well constrained, estimates of elevation gains via crustal shortening and magnitude of exhumation across a thrust belt can involve large uncertainties.

The second approach is to use syn-collisional volcanic events as proxies for possible occurrence of thermal events in the mantle, such as slab break-off (Yin and Harrison, 2000), convective removal of mantle lithosphere (Molnar et al., 1993), or continental subduction (Meyer et al., 1998; Tapponnier et al., 2001; Wang et al., 2001). This approach assumes that a thermal event in the mantle must have led to an increase in the elevation of the plateau. The absolute magnitude of elevation gain is difficult to determine, as it depends on the nature of the tectonic processes as mentioned above.

The third approach is to use the time of rift initiation in Tibet. In the context of a thin-viscous sheet model, the Tibetan lithosphere started to extend when it reached its current and the maximum possible elevation (e.g., England and Houseman, 1989). This approach assumes that the boundary conditions do not change with time and the cause of extension in Tibet is entirely due to an increase in gravitational potential. As pointed out by Yin (2000) and discussed above, east–west extension across Tibet appears to be synchronous and kinematically linked with the similar north-trending rifts across central and east Asia (Fig. 1).

The fourth approach is to use thermal dynamically based theoretical relationships between stable isotope ratios of oxygen and elevation (Rowley, 2007). This approach is sensitive to the past climatic conditions and an elevation estimate represents an average value of the paleo-drainage basin where the sedimentary sample was collected. An variation of this method is to use stable isotope methods to obtain paleo-temperatures at sampling sites and then calculate the elevation gain assuming a reasonable vertical temperature gradient (e.g., Wang et al., 2006).

Finally, foliar physiognomy has also been used to estimate paleo-altitude of large plateaus (Chase et al., 1998; Spicer et al., 2003). This method assumes that the correlation of modern climate with modern folia physiognomy holds in the geologic past, which was difficult to evaluate.

With the above caveats in mind, I outline the distribution of uplift history across the Tibetan plateau obtained from a combination of the above approaches (Fig. 10). The age range shown in each tectonic domain from (A) to (L) represents the duration of uplift in each region. Estimated age bounds for the time when the plateau reached its current elevation using stable isotope or folia physiognomy methods are also indicated. The age for the Lunpola basin came from Rowley and Currie (2006), the Nima basin from DeCelles et al. (2007b), the Namling Basin from Spicer et al. (2003), the Gyirong Basin from Wang et al. (2006), and Thakkhola graben from Garzione et al. (2000). For example, the map indicates that the Lhasa terrane shown as tectonic domain (C) had reached its current elevation at or prior to 65 Ma before the onset of the India collision. It also shows that the onset of crustal shortening and thus uplift started immediately after the beginning of the India–Asia collision in northern Tibet. The above observations suggest that part of the Tibetan topography was inherited from pre-collisional tectonics and the uplift pattern is rather complex, localizing along zones in the northern and central Tibet first and then moved laterally towards the east and the center.

5.6. Cenozoic V-shaped conjugate strike-slip faults in Asia

Taylor et al. (2003) showed that Cenozoic conjugate faults are widespread in central Tibet (Fig. 11). They bound a series of V-shaped wedge blocks facing to the east towards the Pacific trench systems

(Fig. 1). Cenozoic V-shaped conjugate faults have also been documented in the eastern Alps (Ratschbacher et al., 1991a,b), western Turkey (Sengör and Kidd, 1979; Jackson and McKenzie, 1984; Dhont et al., 2006), eastern Afghanistan (Tapponnier et al., 1981; Brookfield and Hashmat, 2001), western Mongolian (e.g., Cunningham, 2005; Walker et al., 2007), Southeast Asia (Leloup et al., 1995), and Gulf of Thailand (Morley, 2001; Morley et al., 2001; Kornsawan and Morley, 2002) (Fig. 1). The common features of the V-shaped conjugate faults are that (1) they lie at 60–75° from the maximum compressive stress direction, (2) the opening direction of the Vs parallels the topographic gradient, and (3) the two sets of conjugate faults terminate at their merging points and have similar magnitudes of slip where they have been studied in details (e.g., Taylor et al., 2003; Leloup et al., 1995) (Fig. 11). This type of conjugate faults contrasts strongly to those in southern California, where strike-slip faults are mostly parallel to the transform plate boundary and the few secondary antithetic faults such as the Garlock and Pinto Mountains faults have much less magnitude of slip and were initiated later than the San Andreas fault (McGill and Sieh, 1991, 1993; Guest et al., 2003).

V-shaped conjugate strike-slip faults in Asia are typically related to extrusion tectonics (Tapponnier et al., 1982; Peltzer and Tapponnier, 1988). In this hypothesis, the fault-bounded wedge-shaped blocks are rigid and were translated laterally away from the collision zone without significant internal deformation. However, as the geometry of the V-shaped strike-slip faults departs greatly from the predicted fault orientation by the Coulomb fracture criterion (Fig. 11), it requires that the faults either formed at their current conditions controlled by an unknown mechanism or have been rotated significantly after their formation (>40°). Given the significant lengths of these strike-slip faults (commonly >100 s km), the possibility of large rotation would require either large distributed strain in fault-bounded regions or large magnitudes of fault slip in groups of parallel faults to accommodate the required rotation. The above problem is a current research subject pursued by the author and M.H Taylor and the results will be published elsewhere.

6. Summary

A synthesis of the available geologic data allows the following generations of the Cenozoic tectonic history of Asia.

(1) Paleogene to mid-Miocene (65–15 Ma) extension is concentrated along the eastern margin of Asia while Oligocene–present extension is concentrated at the western end of Asia. In east Asia, the early phase of extension between 65 Ma and 30 Ma is distributed over a zone of several hundreds of kilometers, while the second phase of extension is relatively localized, resulting in formation of back-arc basins floored by oceanic crust (South China Sea, Japan Sea, and Kurile Basin). Extension at the eastern margin of Asia in the Paleogene was spatially and temporally associated with basaltic eruptions resulted from large-magnitude lithospheric thinning. In westernmost Asia, extension was expressed by the development of large-scale detachment faults across the South Balkan, western Turkey, Aegean Sea, and Cretan Sea associated with rapid southward retreat of the Hellenic arc.

(2) Since the mid-Miocene at ~15 Ma, extension has shifted to the inland region of Asia from the Baikal rift system in the north to the Tibetan rift system in the south. Although Neogene rifts in Asia are generally correlated with the Cenozoic igneous provinces of Asia, the initiation of extension is at least 10–30 Ma younger than the earliest eruptions of volcanic rocks adjacent to the rift systems. This observation suggests that partial melting may have been widespread below Asia and rifts simply served as conduits to facilitate the migration of the melts to the surface. In western Asia, the Turkey extrusion system began to develop, as expressed by the initiation and motion on the North and East Anatolia fault systems at this time interval.

(3) Coeval with the Neogene to present intracontinental east-west extension is the widespread east-west contraction in the western Pacific along the eastern edge of Asia as expressed by initiation of new subduction zones in the back arc regions and contraction across the back arc basins. This coeval extension and contraction may be explained by a change in boundary condition along the eastern edge of Asia. That is, the termination of back-arc spreading has allowed the Asian lithosphere to slide eastward towards the trench, causing its trailing edge to extend and leading edge to contract.

(4) Intraplate volcanic eruptions have occurred nearly continuously in all major cenozoic igneous provinces since the beginning of the Cenozoic, regardless of variation of plate boundary conditions along the margins of Asia. However, a prominent amagmatic zone about 600–800 km wide is present in central Asia that stretches from Junggar basin in the west to the Sichuan basin in the east. It is possible that the lack of Cenozoic volcanic across this zone was caused by the presence of older cratons that have thicker lithospheric lid to prevent the partially molter mantle to migrate to the surface.

(5) Immediately after the initial collision between India and Asia and between Arabia and Asia, widespread volcanic eruptions occurred over regions previously occupied by Mesozoic arcs. This spatial relationship suggests that the two processes are genetically related. The occurrence of most intraplate volcanism in Asia can therefore be associated with continuous upwelling of upper mantle flow induced by slab breakoff, delamination of thickened arc mantle lithosphere, and most importantly continuous sinking of subducted oceanic slabs into lower mantle that may have been replaced by hot mantle materials spreading below the continental lithosphere and caused widespread Cenozoic igneous activity.

(6) The best expression of the interaction of the India–Asia and Arabia–Asia collision in shaping the Cenozoic deformation pattern of Asia is the development of the northeast-trending 300–400-km wide, and >1500-km long zone of northwest-striking right-slip faults, which can be traced from the Zagro thrust belt in the south and western Mongolia in the north. As the right-slip faults terminate at the intracontinental thrust systems in the Tian Shan and Altai Shan of central Asia, the initiation of the right-slip fault zone is considered to be the main driving mechanism for the development of the above orogens in central Asia.

(7) Although the back-arc extension in the Aegean Sea, Cretan Sea and western Turkey was clearly caused by slab rollback, the formation of the back-arc basins in east Asia that are floored by the oceanic crust and associated with back-arc spreading centers may have been induced by oblique subduction. Specifically, the oblique subduction produced a series of nearly evenly spaced primary shears; their development caused the opening of the back arc basins with a rectangular shape in map view and the spreading centers trending obliquely to the strike of the nearby trench.

(8) In both Tibet and west Turkey, the geochemical evolution of the Cenozoic volcanic rocks is rather similar, with the early phase of igneous activity mostly related to partial melting of an older metasomatized mantle lithosphere and the late phased related to a source in the asthenosphere. This commonality may point to a shared dynamic cause for the evolution of the mantle thermal state below collisional orogens.

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