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Continental lithosphere folding in Central Asia (Part I): constraints from geological observations

A.M. Nikishin [a], S. Cloetingh [b*], L.I. Lobkovsky [c], E.B. Burov [d,e**] and A.C. Lankreijer [b]

[a] *Geological Faculty, Moscow State University, Russia*
[b] *Institute of Earth Sciences, Vrije Universiteit, 1081 HV Amsterdam, The Netherlands*
[c] *Institute of Oceanology, Russian Academy of Sciences, Moscow, Russia*
[d] *Department of Earth Sciences, University of Leeds, Leeds LS29JT, England*
[e] *Institute of Physics of the Earth, Russian Academy of Sciences, Moscow, Russia* *

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ABSTRACT

Geological observations support the occurrence of long-wavelength surface undulations in Central Asia, which can be treated in terms of folding of the whole lithospheric plate. The characteristic wavelength of the whole lithosphere folding is 360 km in the western part of the Gobi sub-plate. Superimposed on the long wavelength deformation, two shorter wavelength scales of spatially periodical deformation are also observed. An intermediate wavelength of 30–50 km of the intraplate deformation reflects the response of the mechanically strong upper-crustal part of the Central Asian lithosphere. Small wavelength (4–9 km) deformation corresponds to faulting in the uppermost crustal layer and folding and faulting in the sedimentary cover. Folding of the lithosphere controls the location and geometry of compressional basin formation in the western part of the Gobi sub-plate. A number of folding and faulting characteristics of continental intraplate deformation in Central Asia presented in this study are strikingly similar to observed styles of intraplate deformation in oceanic lithosphere in the northeastern Indian Ocean.

Introduction

Over the last few years a number of studies have revealed the occurrence of large scale lithospheric folds in oceanic and continental lithosphere (Zuber, 1987; Stephenson and Cloetingh, 1991; Nikishin, 1992). Following the discoveries of folding in the Northeastern Indian Ocean (McAdoo and Sandwell, 1985) and the continental lithosphere of central Australia (Lambeck, 1983; Stephenson and Lambeck, 1985) examples of whole lithospheric folding have been reported from the Canadian Arctic region (Stephenson et al., 1990), the North Sea region (Huyghe, 1992; Cloetingh and Kooi, 1992) and the ocean floor offshore the Japanese islands (Chamot-Rooke and LePichon, 1991). From these studies a close genetic link has been inferred between the occurrence of high-level compressional intraplate stresses induced by collisional processes and the timing and spatial distribution of lithospheric folds in the plates. The Indian Ocean intraplate deformation (Fig. 1) is of particular importance in this respect. In this area the fold axes observed on seismic lines (e.g., Geller et al., 1983; Neprochnov et al., 1988; Levchenko, 1990) and SEASAT derived gravity maps are oriented (Stein et al., 1989, 1990) roughly perpendicular to the main axes of compression inferred from stress field modelling (Cloetingh and Wortel, 1985, 1986) and earthquake focal mechanism determinations (Bergman, 1986; Petroy and Wiens, 1989). The distance between neighbouring axes of subparallel whole-lithospheric anticlines is of the order of 150–250 km (McAdoo and Sandwell, 1985; Zuber, 1987), the crust layer being additionally deformed into smaller folds, complicated by

* Corresponding author.
** *Present address*: Institute de Physique du Globe de Paris, Paris 75252, France.

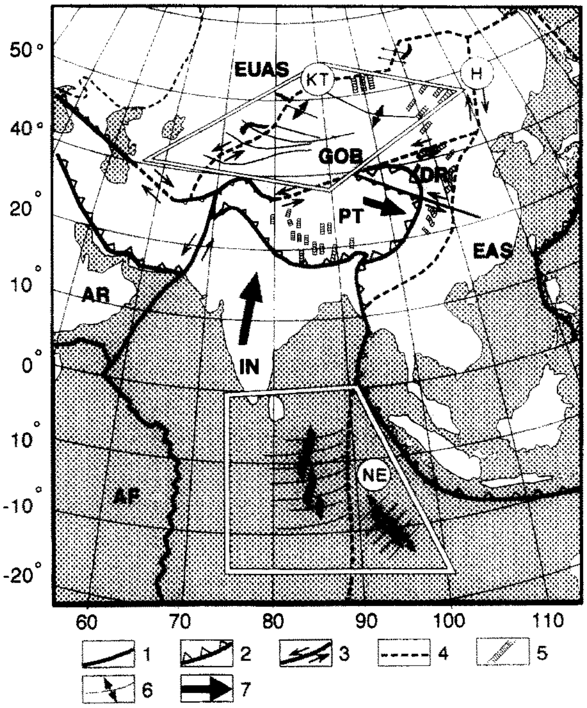

Fig. 1. Location map showing position of two areas of intraplate lithosphere folding associated with the collision of the Indian plate with the Eurasian plate (generalized from Tapponnier et al., 1986). Boxes indicate the northeastern Indian region and the Gobi area, respectively. Also shown are names of plates, plate boundaries and sub-plates discussed in this paper. 1 = Mid-ocean ridges; 2 = Subduction zones; 3 = Whole-lithosphere shear zones at plate boundaries; 4 = Location of Ninety-east ridge (*NE*), Kugitang–Tunka shear-zone (*KT*) and the Hushan–Hingan zone (*H*); 5 = Locations of late Cenozoic grabens; 6 = Axes of lithospheric folds; 7 = Directions of plate motions. *AF* = African plate; *AR* = Arabian plate; *EUAS* = Eurasian plate; *IN* = Indian plate; *PT* = Pamir–Tibet sub-plate; *GOB* = Gobi sub-plate; *EAS* = East Asian sub-plate.

thrusts and reverse faults (Bull and Scrutton, 1990).

The Indian plate is moving NNE at a rate of 7–8 cm/yr, whereas the Indian continent is moving northward at a rate of around 5 cm/yr (Molnar and Tapponnier, 1975; Tapponnier and Molnar, 1979; Molnar and Denq Qidong, 1984; Armijo et al., 1986; Chen et al., 1991). The collision between the Indian plate and the Eurasian plate was perhaps the most important tectonic event during the last 65 Ma of the Earth's history (Molnar and Tapponnier, 1975). The observed intraplate deformation in the northeastern Indian Ocean is the more spectacular as it occurs in relatively strong oceanic lithosphere with an age of 60–80 Ma (Cloetingh et al., 1989). Rheological arguments, based on the extrapolation of rock mechanics data, suggest a lower level of intraplate stress required to induce folding in continental lithosphere (Stephenson and Cloetingh, 1991). Therefore, dependent on the actual stress level in the Eurasian lithosphere, the occurrence of folding on the Central Asian side of the plate boundary between the Indian plate and the Eurasian plate could accompany the oceanic lithosphere deformation to the south of the Indian subcontinent. In the present paper we discuss observations supporting the occurrence of whole continental lithosphere folding in Central Asia (Fig. 1). In the companion paper (Burov et al., 1993-this volume) we present results from quantitative modelling of compressional intraplate deformation for a rheologically stratified lithosphere.

Tectonic setting of Central Asia

Division into sub-plates and microplates

Within Central Asia a number of regions of different structural style can be distinguished from south to north: the Pamir–Tibet, the Tien-Shan–Altai and the Baikal–East Mongolian regions (Milanovsky, 1989). In most of the western literature this area is commonly divided into the Pamir–Tibet, Tien Shan, Dzungaria (Junggar), Tarim, Gobi–Altai, Altai–Sayan and Baikal regions (e.g., Tapponnier and Molnar, 1979). Overall crustal thickening and eastward tectonic escape of the lithosphere occurs within the Pamir–Tibet region (Molnar and Tapponnier, 1975; Tapponnier and Molnar, 1979). The complexity of the area is expressed in the large set of tectonic models proposed to describe the tectonic structure and evolution of the area (e.g., Molnar and Tapponier, 1975; Zonenshain and Savostin, 1979; Peltzer and Tapponier, 1988; Khain, 1986). A prominent feature of these models for Central and Eastern Asia is the large spectrum of divisions of the lithosphere in terms of plates, microplates and sub-plates. We favour a rough division of this region into two large Central Asian and Eastern Asian domains (see Fig. 1). Each of these domains, or plates, can be subsequently divided into numerous sub-plates and microplates. The structures of the Central Asian and Eastern Asian domains are obviously controled by the interaction of the Indian, Eurasian and Pacific plates (e.g., Molnar and Tapponnier, 1975). The Central Asian plate is subjected to considerable compression in the west (Peltzer and Tapponier, 1988), and escapes in an eastward direction. The Eastern Asian plate undergoes extension associated with roll back of Pacific subduction zones (Tapponnier et al., 1986; Nikishin, 1992). The structure and make-up of the Central Asian plate (also known in the Russian literature as the High Asia area) is mainly determined by its collision with the Indian continent (Tapponnier et al., 1986; Allen et al., 1991). The structure of the Eastern Asian plate is also controlled by the West Pacific mobile belt. The western boundary between the Central Asian plate and Eastern Asian plate follows approximatelly the so-called Kugitang–Tunka line. This prominent topographic boundary corresponds also to a sharp contrast in the seismic structure of the lithosphere (Zverev and Kosminskaya, 1980) and to the upper mantle transition between an area with very low attenuation of seismic surface waves in the west and an area with higher attenuation in the east (Kopnichev, 1987). Therefore, this boundary potentially divides different crustal and upper mantle provinces (Ryaboy, 1987).

The eastern border of the Central Asia plate boundary probably coincides with the N–S-striking Hushan–Hingan mountain belt, separating High Asia from eastern regions of lower topography (Fig. 1). This boundary separates regions with a different character of neotectonic development and corresponds to a change in the average level of gravity anomalies as well as to a gradient zone in the mapped density distribution of the upper mantle of China (Kunin et al., 1988). It also can be distinguished as a belt of increased seismicity (Simkin et al., 1989). The northern boundary of the Central Asia plate is determined by large sinistral strike-slip faults controlling the Baikal rift zone: those are the Tunka and the Muja–Chara (emerging as Stanovoy in the East) strike-slip faults. The boundary coincides with the Baikal rift linking these strike-slip faults.

The Central Asian plate can be divided into several sub-plates. The most important ones are the Pamir–Tibet plate and the Gobi plate. The Pamir–Tibet sub-plate is an area of very young plateau uplift with an amplitude of 5–6 km (Molnar and Tapponnier, 1975; Nikolaev, 1987). The boundaries of this sub-plate roughly correspond to contours of Phanerozoic folding regions of different ages. It is bordered by the Himalaya collision belt on the south and by zones of major thrusts coinciding with Paleozoic structures (Lyon-Caen and Molnar, 1983). The region is also bounded by the large sinistral Altyn–Tagh strike-slip fault along the Hercynian ophiolite suture in the north, and by zones where thrusting occurs over the edge of the Yangtze platform on the east (Tapponnier and Molnar, 1977; Milanovsky, 1991). The Pamir–Tibet sub-plate is distinctly outlined by seismic belts (Roecker, 1982; Simkin et al., 1989) and corresponds to the region of strong (up to 60–70 km) thickening of continental crust with significant thickness gradients on its edges (Zhang et al., 1984; Vinnik et al., 1986; Kunin et al., 1988).

The displacement along the Altyn–Tagh strike-slip zone amounts probably up to several hundred kilometers (Peltzer and Tapponier, 1988). An upper-crustal origin (Burchfiel et al., 1989) as well as a whole-lithospheric character (Peltzer and Tapponier, 1988; Hirn, 1988) has been proposed for this strike-slip fault. The zone of the strike-slip fault itself could be whole-lithospheric, whereas its structure is probably multiple layered. It is possible that the crustal thickening of the Pamir–Tibet sub-plate is not only due to general crustal compression and shortening (e.g., Tapponnier and Molnar, 1977; Molnar and Denq Qidong, 1984) but also at some extent due to autonomous accretion of lower-crustal plastic material (Lobkovsky and Kerchman, 1991).

A number of late Cenozoic submeridional grabens is observed in the Himalaya–Tibet region (Armijo et al., 1986; Tapponnier et al., 1986). Most of them are observed in the northern Himalaya and southern Tibet. The typical width of these grabens is 4–10 km, many of them being linked by strike-slip faults. Numerous, though not very big, manifestations of volcanism are associated with zones of Tibetian graben formation. The Himalaya–Tibet grabens are assumed to have been formed along extension zones due to considerable compression of the Himalaya–Tibet region and crustal material escaping eastward, caused by the Indo-Asian collision (Armijo et al., 1986).

Lithospheric folding in the Gobi sub-plate

The Gobi sub-plate is located to the north of the Pamir–Tibet plate (Fig. 2). The Gobi area was the site of a large collision belt in the Late Paleozoic–Early Mesozoic. In the Mesozoic, similar to the whole Ural–Mongolian collision belt, this region underwent a phase of continental extension. The extension, following the younger collision, preceded thermal subsidence in the Jurassic, Cretaceous and Tertiary. This subsidence was stopped by renewed movement on Palaezoic reverse faults, caused by the Mesozoic collisions at the southern margin of Asia. The India–Asia collison in the early Tertiary has produced widespread thrusting in the Tien Shan and Goby and formed active foreland basins (Allen et al., 1991). Rejuvenation by intracontinental tectonic activity in Oligocene–Miocene times was marked by intensive crustal shortening (up to 200–300 km) and crustal thickening (Burov et al., 1990). Fault plane solutions of earthquakes suggest continued crustal shortening (Nelson et al., 1987), whereas seismic refraction data demonstrate abnormally thick crust (Volvovsky and Volvovsky, 1974; Vinnik et al., 1986). This process has lead to formation of a system of subparallel ridges in the western part of the Gobi sub-plate (Fig. 2), apparently created by NE transpressional horizontal stresses induced by northward movement of the Pamir–Tibet sub-plate (Tapponnier and Molnar, 1977, 1979). In the area near the western boundary of the Gobi sub-plate the following subpararallel ridges can be distinguished: the southern Tien Shan, Karatau–middle Tien Shan, Chu-Ili mountains–northern Tien Shan, Junggar Alatau–Borohora, Tschingiz-Tarbagatai, Altai, and Khangai. The ridges of the southern, middle, and northern Tien Shan and Junggar Alatau merge to the east into the Tien Shan ridge (Fig. 2; see also Burov et al., 1990).

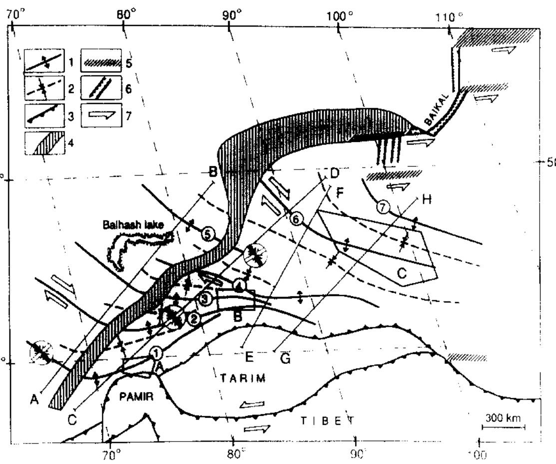

Fig. 2. Location of axes of late Cenozoic (0–10 Ma) zones of large scale uplift and subsidence in Central Asia (after Nikolaev, 1985, 1987; Tapponnier and Molnar, 1979). 1 = Axes of whole-lithospheric anticlines; 2 = Axes of whole-lithospheric subsidence; 3 = Thrusts; 4 = Kugitang–Tunka megashear zone; 5 = E–W-trending shear zones; 6 = Baikal rift; 7 = Strike-slip faults. Numbers in circles denote the system of main continental folded ridges: *1* = Southern Tien Shan; *2* = Mid Tien Shan; *3* = Northern Tien Shan; *4* = Junggar–Alatau–Borohoro; *5* = Chingiz–Tarabagatai; *6* = Altai; *7* = Khangai. Circles with arrows: dark arrows correspond to maximum horizontal compressional stress components, open arrows correspond to intermediate (transpressional) stress components. Areas *A*, *B*, and *C* described in Figs. 6–8.

Spacing of long-wavelength ridges and depressions

Neotectonic maps (Nikolaev, 1985, 1987; see also Figs. 2 and 4) show the expression of these ridges in the form of lineaments of 1–5 km elevation alternating with linear-shaped depressions. The magnitude of these depressions varies between 1 and 5 km below the reference level of the regional topography. The maximum observed peak to peak amplitude of the folds displayed in Figure 2 is of the order of 8 km (see Fig. 3). The system of subparallel ridges and linear depressions is overprinted by NE–SW-trending shear zones (Tapponnier and Molnar, 1979). To the west of the Kugitang–Tunka line, the distances between subparallel ridges are roughly the same, with values of 405 km for the spacing between the south Tien Shan–Karatau ridges, whereas the Karatau–Chu-Ili mountains are separated by a distance of 330 km. Observed spacings between the Chu-Ili mountains–Junggar Alatau, the Junggar Alatau–Tschingiz-Tarbagatai and the Tschingiz-Tarbagatai–Altai are 360, 316, and 390 km, respectively. Hence it appears that the linear structures are separated by an average distance of about 360 km (Figs. 2 and 3).

In the area of the Kugitang–Tunka line the orientation of the ridges changes from a NW-striking direction, observed in the area west of the Kugitang–Tunka line, to a more E–W trend of the ridges in the area at the eastern side of this major lineament. As mentioned above, northward of the Tarim block at a longitude of 82° the intra-ridge distances decrease dramatically due to the merging of the Tien Shan ridges and the Junggar Alatau into the joint eastern Tien Shan (Fig. 2).

Spatial characteristics of whole-lithospheric folding

We propose that the mountain ridges and intermountain linear depressions potentially correspond to whole-lithospheric anticlines and whole-lithospheric synclines, respectively, with an average wavelength of whole-lithospheric folding of 360 km in the West of the Gobi sub-plate. This idea is based on an analogy with the Indian Ocean intraplate deformation zone where the whole-lithospheric folding of 60–80 Ma old oceanic lithosphere is associated with wavelengths of about 200 km. The older continental lithosphere is relatively stronger and, therefore, its folding should be characterized by larger

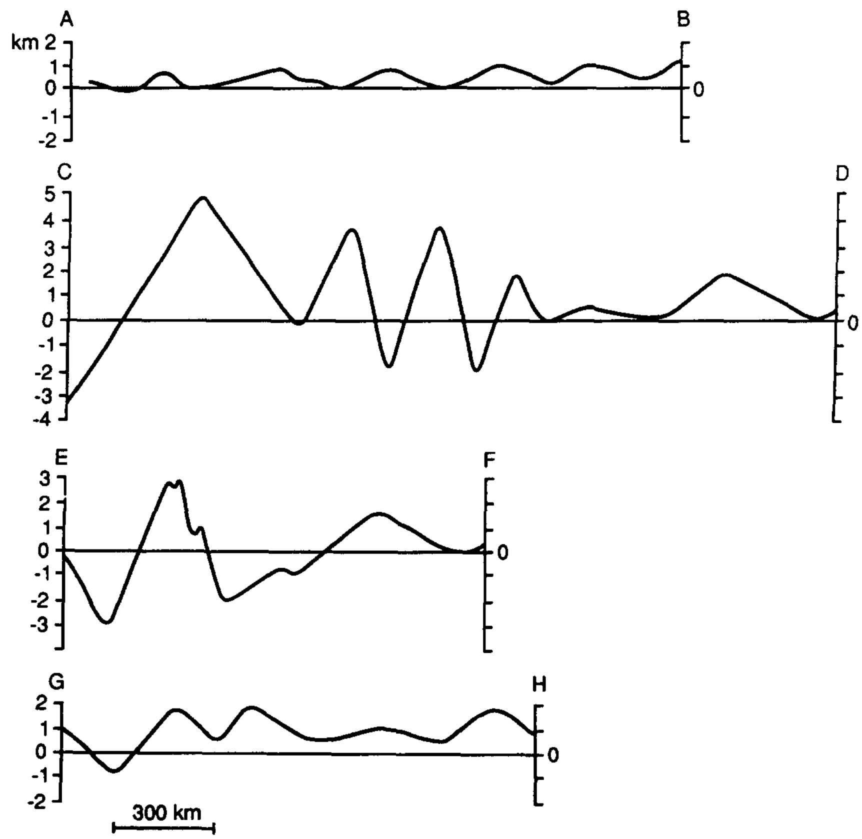

Fig. 3. Profiles of averaged values of late Cenozoic (0–10 Ma) long-wavelength uplift and subsidence of folded lithosphere in Central Asia. The location of the NE–SW-trending profile lines is given in Fig. 2. Amplitudes of whole-lithospheric folding are maximal in the area to the east of the Kugitang–Tunka lineament.

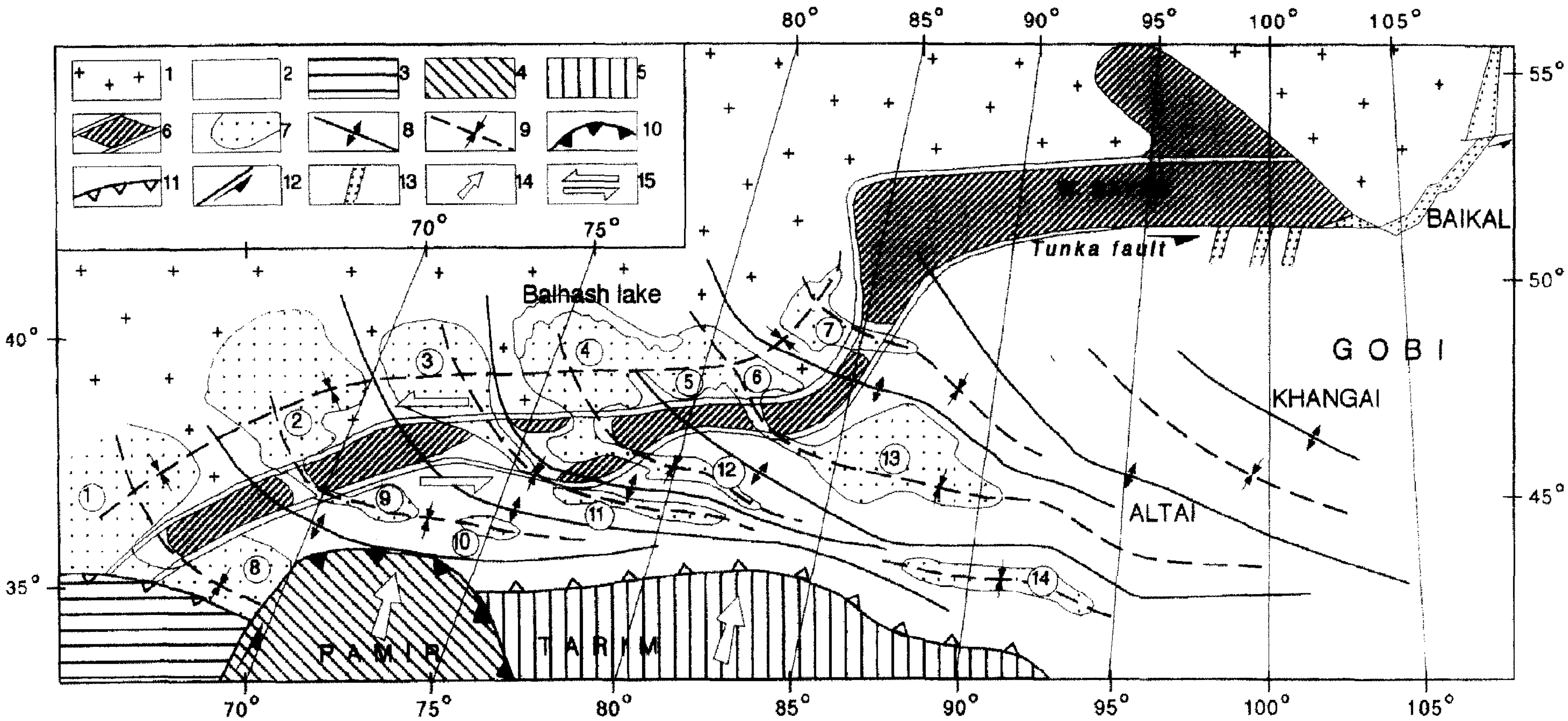

Fig. 4. Late Cenozoic structures of Central Asia; 1 = Eurasian plate; 2 = Gobi sub-plate; 3 = Iran–Afghan sub-plate; 4 = Pamir–Tibet sub-plate; 5 = Tarim massif of Gobi sub-plate; 6 = Transpressional sinistral Kugitang–Tunka lineament; 7 = Late Cenozoic sedimentary basins; 8 = Axes of proposed whole-lithospheric anticlinal folds; 9 = Axes of proposed whole-lithospheric synclinal folds; 10, 11 = Thrusts; 12 = Strike-slip faults; 13 = Baikal rift; 14 = Directions of motion of Pamir and Tarim sub-plates; 15 = Direction of motion along Kugitang–Tunka sinistral shear zone. Numbers in circles denote sedimentary basins with a location controled by the interplay of compressional structures and the Kugitang–Tunka shear zone in the area of continental intraplate deformation in the western part of the Gobi sub-plate. Note the differences in geometry and aspect ratio of the basins located to the east and west of the Kugitang–Tunka zone. *1* = Amu–Darya; *2* = Tashkent; *3* = Chu; *4* = Balhash; *5* = Lensi; *6* = Alakol; *7* = Zaisan; *8* = Tajik; *9* = Fergana; *10* = Naryn; *11* = Issyk-Kul–Tenes; *12* = Ili; *13* = Junggar; *14* = Turfan.

wavelengths (Marthinod and Davy, 1993). The observed decrease in wavelength of folding corresponds to the areas of localized compressional deformation situated northward of the Tarim massif (Figs. 2 and 3). The Kugitang–Tunka line itself forms a chain of blocky mountainous structures segmented by the system of linear folding axes crossing this lineament (Fig. 4). From southwest to northeast the Kugitang–Tunka structure can be divided into the Kugitang, Chatkal, Junggar Alatau, Barlak, Urkashar and west Sayan segments of high topography, respectively. We favour a transpressional origin of the Kugitang–Tunka megastructure, as a result of the superposition of sinistral shear motion and northward-directed compression induced by the NE movement of the Gobi sub-plate due to the Indo-Asian collision (Fig. 1; see also Chen et al., 1991).

Relationship between near-surface deformation and large-scale folding

The large-scale lithosphere folding is apparently accompanied by deformation on intermediate and small spatial scales. As discussed in the following, the observed spectrum for the spacings separating folding axes ranges between approximately 30 and 400 km. On a local scale, within the Tien Shan area, mountain ranges can be subdivided into systems of smaller subparallel ridges, separated by linear depressions, trending at a large angle to the whole-lithosphere folds (Fig. 5). As shown by Figure 5, the distances between neighbouring ridges are of the order of 30–50 km. Similar patterns of superimposed crustal and lithospheric deformation can be observed in the Mongolian Altai region (Fig. 6).

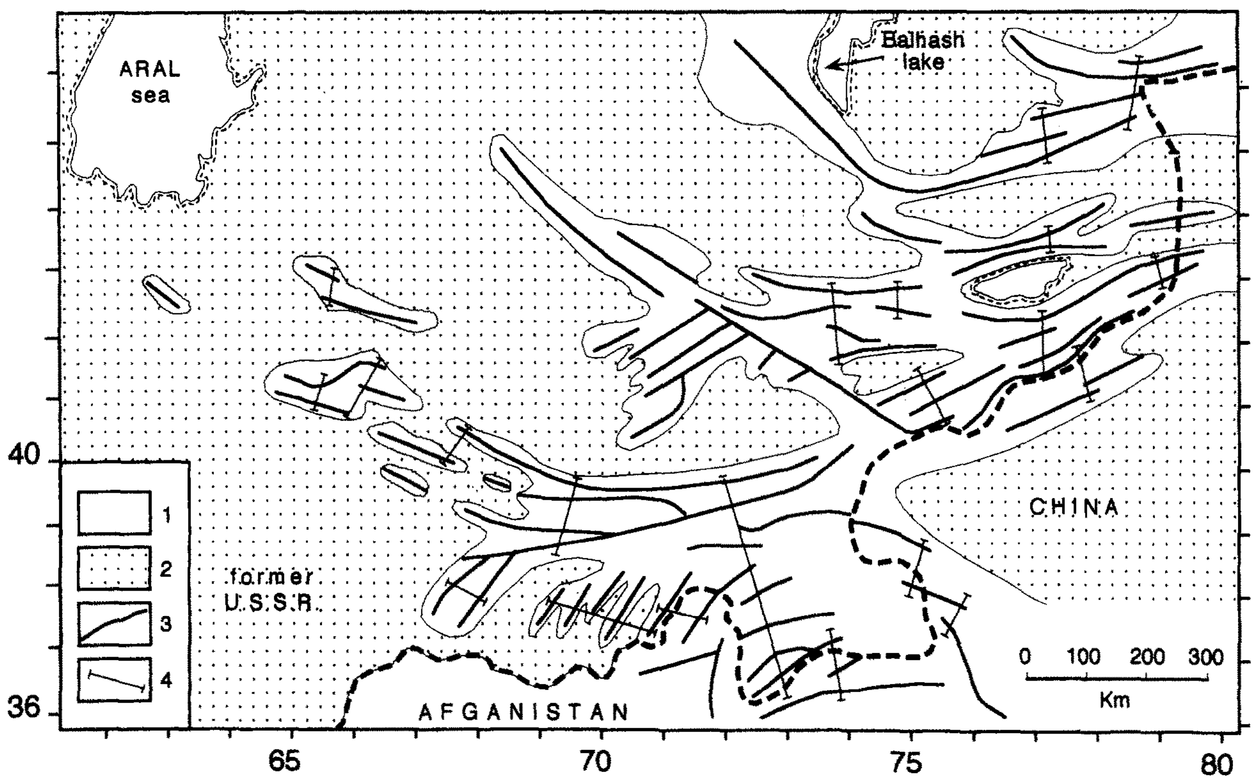

Fig. 5. Locations of mountain ridges in the Tien Shan and Pamir regions (after Milanovsky, 1989). 1 = Mountain regions associated with long-wavelength folding (characteristic wavelength of 360 km) of the entire lithosphere; 2 = Sedimentary basins; 3 = Axes of mountain ridges associated with intermediate wavelength (30–50 km) intraplate deformation; 4 = Lines of inter-ridge distance measurements used for histograms displayed in Fig. 9.

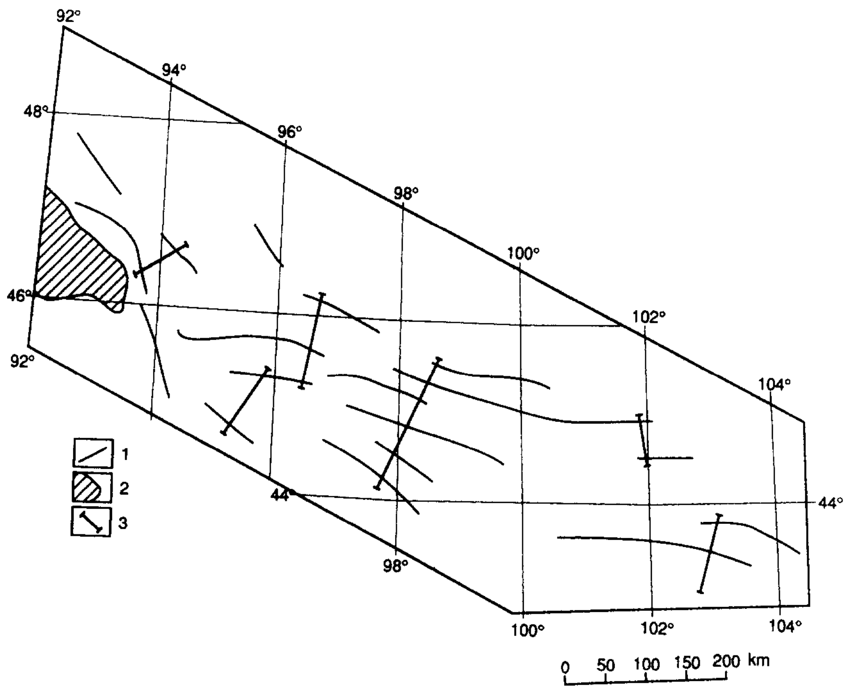

Fig. 6. Orientation of mountain ridges of intermediate wavelength (30–50 km) within the Mongolian Altai region (see for location area *C* in Fig. 2). 1 = Axes of ridges; 2 = Relief mountain area induced by long-wavelength folding of the entire lithosphere; 3 = Lines of inter-ridge distance measurements used for the construction of the histograms shown in Fig. 9.

Typical distances observed between subparallel neighbouring axes of neotectonic rises in northern Tien Shan are in the order of 6–9 km with some of the ridges spaced at distances as small as 3 km and as large as 17 km, respectively (Fig. 7). In the transitional area between Tien Shan and Pamir typical average distances between axes of subparallel neotectonic rises are 4–7 km, with extreme values of 3 to 10 km (Fig 8).

Figure 9 displays histograms of mapped distances between subparallel linear structures compiled by us for the areas discussed above. Hence it appears (see also Fig. 10) that the long-wavelength response of Central Asia to whole-lithospheric folding is superimposed on deformation characterized by intermediate and small wavelengths of 30–50 km and 4–9 km, respectively. The intermediate wavelength deformation proba-

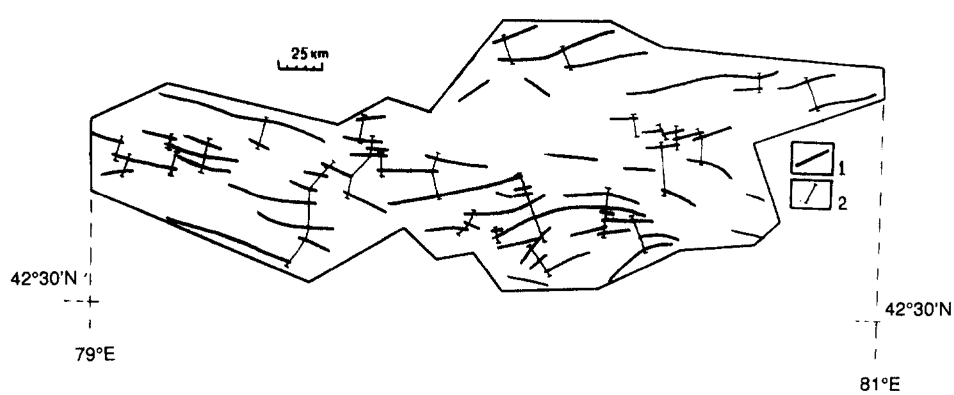

Fig. 7. Orientations of axes of neotectonic short-wavelength (4–9 km) structural uplifts in northern Tien Shan (after Kuchai, 1980). 1 = Axes of the rises; 2 = Lines of distance measurements between axes of uplifts employed for the construction of Fig. 9. Location denoted as area *B* in Fig. 2.

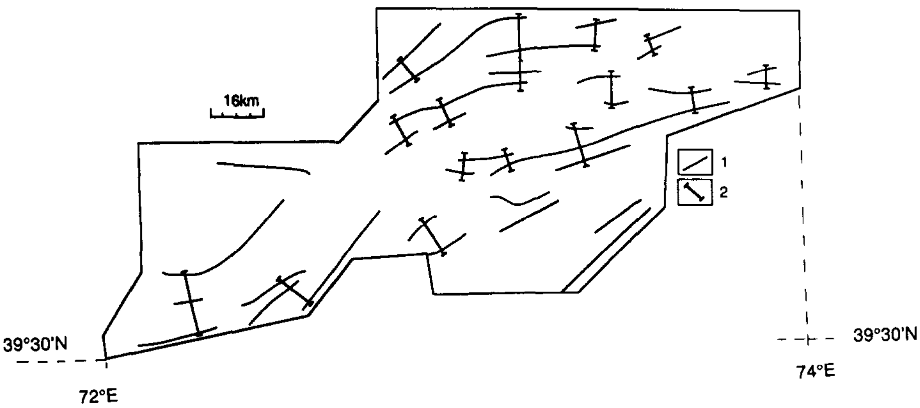

Fig. 8. Orientations of axes of neotectonic short-wavelength structural uplifts in the Pamir–Tien Shan transition zone (after Kuchai, 1980). Figure conventions as in Fig. 7. Location denoted as area *A* in Fig. 2.

bly reflects deformation in the mechanically strong upper part of the crust, whereas the small wavelength deformation could correspond to faulting in the uppermost part of the crust and folding and faulting of the sedimentary cover.

Basin formation in folded lithosphere

A large number of the intermediate wavelength structures appear to be inversion structures and ramp downwarps within the axial parts of whole-lithospheric folds (Fig. 11). The geometrical shape of interridge depressions (whole-lithospheric synclines) can be more complex along strike due to compression. Three principal modes of deformation occur (Fig. 11). Gentle large scale synclinal depressions are, for example, associated with the formation of the Syr Darya, Chu–Sarysh and Balhash basins (see Fig. 4). These basins underwent a subsidence of less than 1 km during the last 10 Ma. The late Cenozoic history of the area northwest of the Kugitang–Tunka line is of particular interest. A chain of Neogene sedimentary basins has been formed in this region (see Fig. 4), including the Tashkent, Chu, Balhash, Lensi, and Zaisan basins. The basins have an ellipsoid and isometric shape in planview and form marginal depressions associated with the Kugitang–Tunka transpression line. We propose that these Neogene basins have been formed in the crossing nodes of whole-lithospheric NW-striking folds with the NE-trending Tashkent–

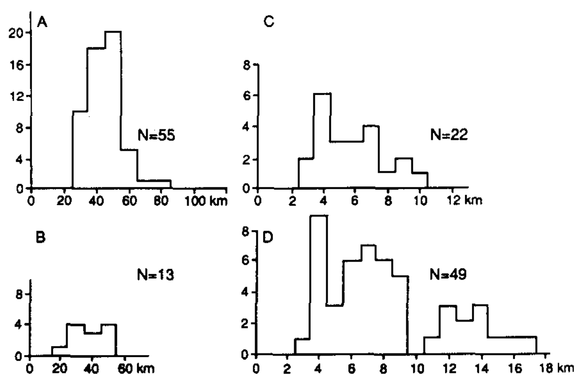

Fig. 9. (A,B) Histograms of distances (km) between subparallel ridges in the Pamir–Tien Shan (A) and Altai (B) regions. (C,D) Histograms of distances (km) between axes of neotectonic rises in the Pamir–Tien Shan transitional zone (C) and the northern Tien Shan region (D).

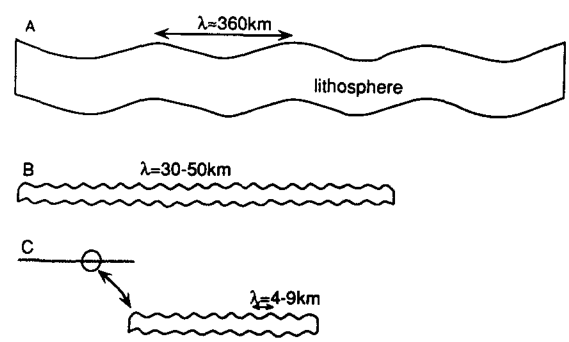

Fig. 10. Three levels of folding in Central Asian lithosphere induced by intraplate compression caused by the Indian–Asian collision. (A) Whole-lithospheric folding, supported by the strong mantle part of the mechanical lithosphere, with a characteristic wavelength of 360 km. (B) Folding in the upper (crustal) part of the mechanical lithosphere with a characteristic wavelength of 30–50 km. (C) Folding in the upper crust and sedimentary cover with a wavelength of 4–9 km.

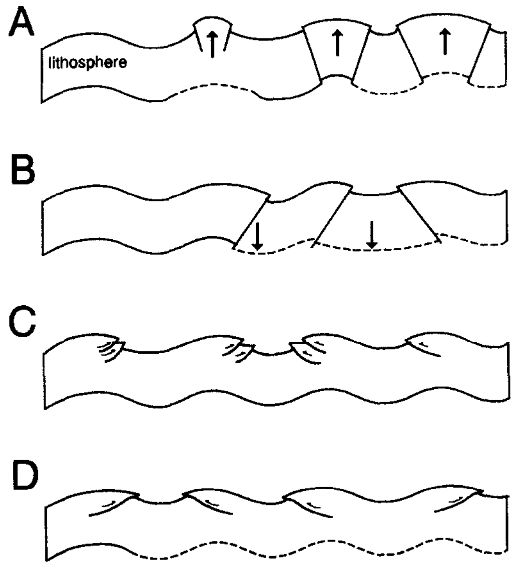

Fig. 11. Cartoons illustrating scenarios for basin formation and inversion superimposed on long-wavelength deflection of the lithosphere. (A) Inverted structures and horsts resulting from tectonic compression. (B) Formation of ramps and downfaulted basins during compression. (C) Formation of thin-skinned compressional thrusts. (D) Thick skinned basement involved foreland thrusting.

Zaisan lineament striking parallel to the Kugitang–Tunka line.

Discussion

The observations discussed above suggest the occurrence of a high level of intraplate deformation in the western Gobi area. The structural patterns and the geometry of the interplay of large scale lithospheric folding and intermediate scale faulting are strikingly similar to the characteristics observed for the zone of intraplate deformation in the northeastern Indian Ocean. The system of long wavelength folds is particularly well developed in the ocean floor southeast of Sri Lanka. The wavelengths of folds in the Indian Ocean (approximately 200 km) are smaller than the wavelengths in the area west of the Kugitang–Tunka lineament but are in the same range as wavelengths observed in the eastern part of the Gobi sub-plate. The wavelength of the folds reflects the interplay of the compressional stresses with the depth-dependent layered brittle–ductile rheology of the lithosphere. The observed spatial variations in the wavelengths point to important lateral variations in the mechanical properties of the continental lithosphere in Central Asia. This is partly the result of lateral variations in the level of stress caused by the presence of rigid blocks such as the Tarim block, causing a concentration of deformation in the indented lithosphere. An important part of the observed variations in the spacing of whole-lithospheric folds might reflect lateral differences in crustal thickness and lithospheric strength inhereted from preceeding extensional phases. The large wavelengths observed point to the presence of thick and relative strong continental lithosphere with a strength in excess of oceanic lithoshere. As discussed by several authors (e.g., Cloetingh and Banda, 1992) this can be the case when a thick mantle lid of continental lithosphere of thermotectonic ages in excess of 200 Ma is involved in intraplate deformation. The lithosphere in the western Gobi plate has probably had sufficient time (approximately 100–125 Ma) to undergo substantial cooling and strengthening after the phase of Jurassic–Early Cretaceous extension (rifting) which preceded the late Cenozoic folding phase.

It appears that the rheological make-up, controlled by its prefolding evolution, of the western Gobi sub-plate has played an important role in localizing the intraplate deformation in Central Asia in this area. The area is relatively weak (Burov and Kogan, 1990; Burov et al., 1990) compared to the adjacent Siberian platform region with a much larger thermomechanical age. For example, the Siberian craton lithosphere is as thick as about 200 km (Zorin et al., 1989), whereas the rejuvenated lithopsphere of the studied area is hardly thicker than 100 km (Burov and Diament, 1992). It appears that the area of thinned lithosphere in the western Gobi plate, located between strong lithosphere of the Indian and Siberian platforms was an optimal site for folding induced by the Indo-Eurasian collision.

Another main factor appears to be the anomalously high concentration of stress in the folded area in the western Gobi plate. An analogy for the controlling effect of the stress factor is provided by the Indian Ocean intraplate deformation area. Here the folding is concentrated in the area of oldest and strongest oceanic lithosphere, coinciding with the location of the stress maxima predicted by finite element stress modelling (Cloetingh and Wortel, 1985, 1986). The stress levels in the Indian Ocean are of exceptionally high magnitude as a result of geometrical focusing induced by the geometry of the plate boundaries and the age distribution of the subducted lithosphere. The absence of folding in the Indian platform can probably be explained both by a lower stress level resulting from its geometrical position in the Indian plate and by its relatively strong lithosphere of high thermo-tectonic age. Evidence for modest late Neogene compressional deformation has, however, been inferred from the sedimentary record and anomalous subsidence of the eastern Indian continental margin (Whiting, 1991).

The western Gobi intraplate compression is flanked by a zone of major extension in the east. It is interesting to note that a similar situation occurs in the Indian Ocean compressional area. The transition from the intraplate compressional area in the Indian Ocean to an area of major tensional stress concentration in the Chagos–Laccadive zone can be explained by lateral changes in stress caused by variations in plate boundary forces (Stein et al., 1990).

Another important similarity can be noted for the geometry of the interaction of intermediate structures and the axes of folding. As illustrated in Figure 12, the segmented structure of the basins and axial depressions in the area of the Kugitang–Tunka line is very similar to the make-up of the area where the Ninety-East Ridge is intersected by E–W-trending folds. As discussed by Stein et al. (1990) the blocky structures of the Ninety-East Ridge has been attributed in the past to its formation due to hot spot activity. Careful analysis of ocean floor topographic and SEASAT data, however, has shown that the segmentation of the Ninety-East Ridge is an expression of the uppermost Miocene to recent folding phase. Both the Kugitang–Tunka zone and the Ninety-East Ridge area mark the sites where the strikes of the lithosphere fold axes undergo an important rotation of at least 30°. Future analysis and tectonic modelling of the sedimentary basins in the Kugitang–Tunka area holds the prospect to obtain accurate constraints on the timing and distribution of the fine structure of intraplate deformation. At the same time, the recognition of intraplate continental folding offers a quantitative and predictive framework for future research on basin formation in this area.

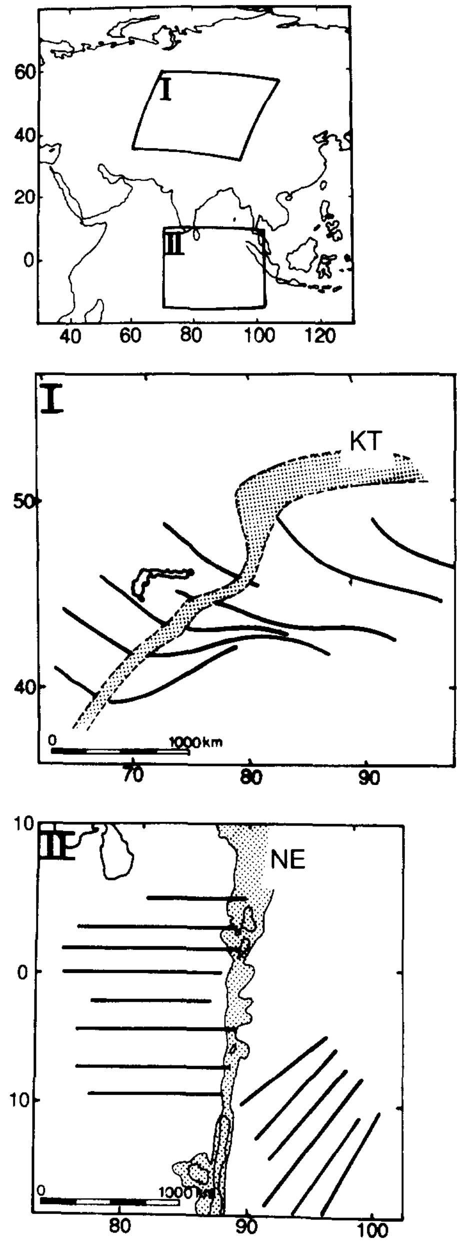

Fig. 12. Comparison between Central Asia and Indian Ocean. (I) Instersection of folding axes (solid lines) and Kugitang–Tunga megashear (stippled area). (II) Intersection of folding axes (solid lines) and Ninety-East Ridge (stippled area).

Conclusions

We have proposed that large-scale continental lithosphere folding occurs in Central Asia. The lateral variations in the observed wavelengths of the folds may contain interesting information on the mechanical properties of the lithosphere and the interplay of stresses and rheology. An important factor appears to be the structural grain of the lithosphere in the western Gobi region, dominated by the presence of the Kugitang–Tunka lineament. At the same time the focusing of stresses induced by the plate collision probably controls the occurrence of the folds in relatively weak lithosphere attenuated by previous extensional phases. The Indian–Eurasian plate collision, perhaps the most important tectonic event of Tertiary–Quaternary history of plate interaction, has led to a rather symmetrical picture in the distribution of intraplate deformation. The zones of folding in the Indian Ocean and the western Gobi area are found at considerable distances from the Himalayan and Java–Sumatra plate boundaries, separated by the relatively strong Indian continental lithosphere. Both areas are bounded by adjacent areas of extension. Superimposed on the folding in the western Gobi is an important concentration of brittle deformation on intermediate and small scales, with spacings of 30–50 km and 4–9 km, respectively. These structural patterns are strikingly similar to findings reported for the interplay of folding and faulting in the Indian Ocean. The segmentation of the sedimentary basins in the western Gobi region by the interplay of folding and the strike-slip zone of the Kugitang–Tunka has a strong analogy in the segmentation of the Ninety-East ridge by axes of compressional folding. The style of continental folding has also important implications for other aspects of the formation of sedimentary basins in Central Asia, including basin inversion and downwarping by ramping.

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