Rheology of the Indian and Tarim plates in the Karakoram continent-to-continent collision zone

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Abstract
Bouguer gravity anomalies in the region of Western Himalayas, Karakoram and Tien Shan show large negative values, but classical isostatic models are insufficient to account for the detailed pattern of the observed anomalies. In the past years the gravimetric surveys in the Karakoram done by Marussi, Caputo and others in 1954 have been extended and intensified. The full body of available gravimetric data, including the pendulum observations by De Filippi and Hedin at the beginning of this century, have been re-analyzed. Terrain corrections have been computed systematically for all available data using a unique algorithm and Digital Terrain Model. The isostatic anomalies along a profile from the Indo-Gangetic foredeep, across the Karakoram range and terminating in the Tarim basin show the oscillating values already noted by Marussi. It is here proposed that this oscillatory pattern can be explained by a model in which the convergent boundaries of the Indian and Tarim plates deform by elastic flexure, besides isostasy. The gravity data constrain the numerical values of the model parameters, particularly the flexural rigidity of the plates. For the Indian plate the best fitting value of the flexural rigidity is $D = 5\ 10^{24}$ N m, a value very similar to those reported in Central Himalaya. The flexural rigidity of the Tarim plate turns out to be considerably larger $D = 7\ 10^{25}$ N m, which makes the Tarim more rigid than the neighboring Central Tibet. Both plates are loaded by an estimated shear stress of $7\ 10^{12}$ N m^{-1} located in a region corresponding to the Nanga Parbat Haramosh syntaxis. It is concluded that the Indo-Asian continental collision in the Western Himalaya and Karakoram resulted in the development of flexural basins on both sides, unlike the Central Himalaya where the collision produced a flexural basin, the Ganga basin, to the south and, to the north, the indentation of an isostatically supported Tibetan block with possible rheological layering and eastward lateral extrusion.

Key words *gravity anomalies – rheology of the lithosphere – isostasy*

1. Introduction

The Indo-Asian collision zone is a major example of continental collision and compressive intracontinental deformation during the Cenozoic era and serves as an ideal laboratory for testing models of processes related to diffuse deformation at or near continental margins. According to Molnar and Tapponier (1975), in Central Tibet the Asian lithosphere behaves as a plastic deformable layer progressively indented by the rigid block of India as this has moved northwards by about 2500 km during the past 50 Ma (fig. 1). The major consequence of the collision has been the thickening of the crust beneath the Tibetan Plateau from 65 to 80 km, as implied by measurements of the phase velocities of surface waves (Brandon and Romanowicz, 1986; Molnar, 1988). According to the principle of isostasy, crustal thickening implies surface elevation in a proportional amount, and therefore mass defi-

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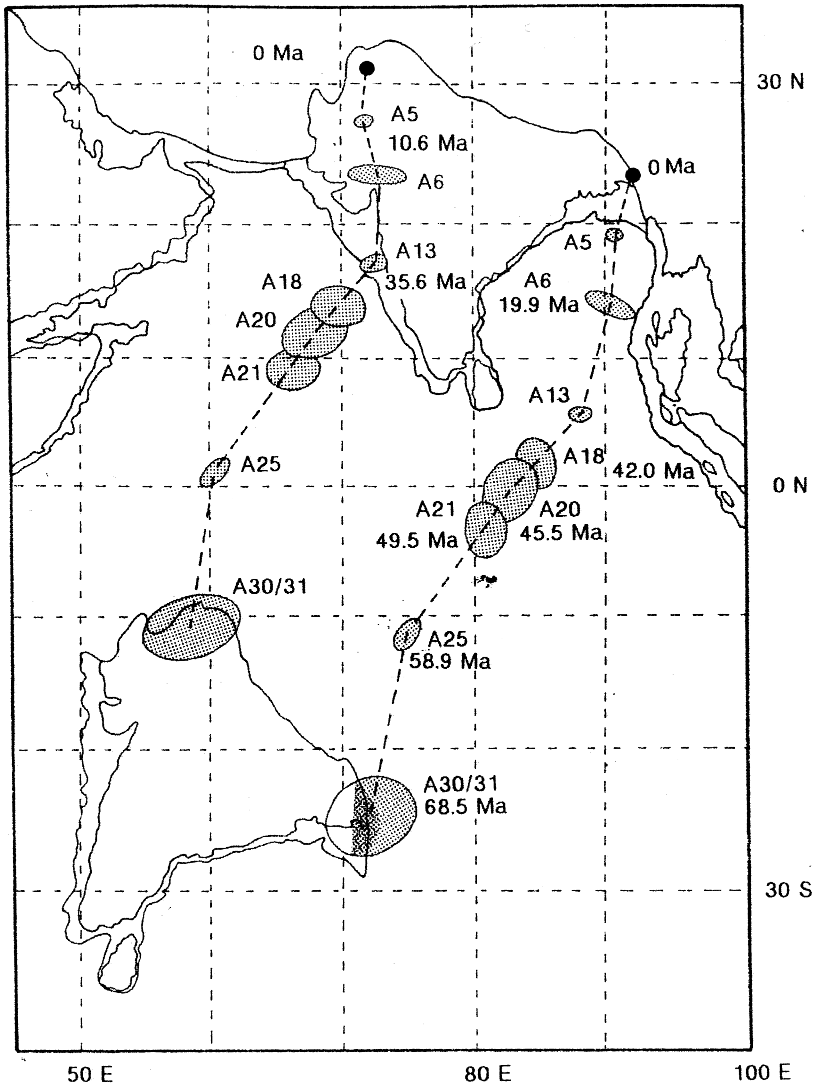

Fig. 1. Schematic paleomagnetic reconstruction of the Cenozoic northward collision of India and Central Asia. The path of two points, Nanga Parbat (left) and Namche Barwa (right) is shown. Points of corresponding anomaly are labeled by Ann, an epoch (in million years) and an ellipse of uncertainty (after Molnar *et al.*, 1993).

cency. A more recent analysis suggests that the upper mantle of Tibet is too strong to shorten by an amount needed to double the crust (Jin *et al.*, 1994). The Tibetan Plateau is folding at two wavelengths, 150 and 500 km in the east-west direction. Numerical models imply a «jelly sandwich» structure with two strong layers (upper crust and upper mantle) separated by a ductile layer (lower crust) (Zuber, 1987, 1994).

The large scale pattern of strike slip faults in Asia (fig. 2) suggested to Molnar and Tapponier (1975), Tapponier and Molnar (1977, 1979) and Armijo *et al.* (1989) that the Tibetan crust was extruded westward, bounded by the sinistral Altyn Tagh fault to the north and the dextral Karakoram fault to the south (Avouac and Tapponier, 1993). In a different approach (Dewey and Burke, 1973; England and Houseman, 1986), crustal thickening was proposed to occur as a consequence of distributed N-S shortening of the crust beneath Tibet. Abrupt elevation of the Tibetan Plateau some 8 Ma ago should have been produced by convective removal of the lower lithosphere and replacement by the hotter asthenosphere. In a third model (Powell and Conaghan, 1973; Willet and Beaumont, 1994), the Asian crust was thrust over India and the Asian lithospheric mantle was subducted beneath Tibet. This

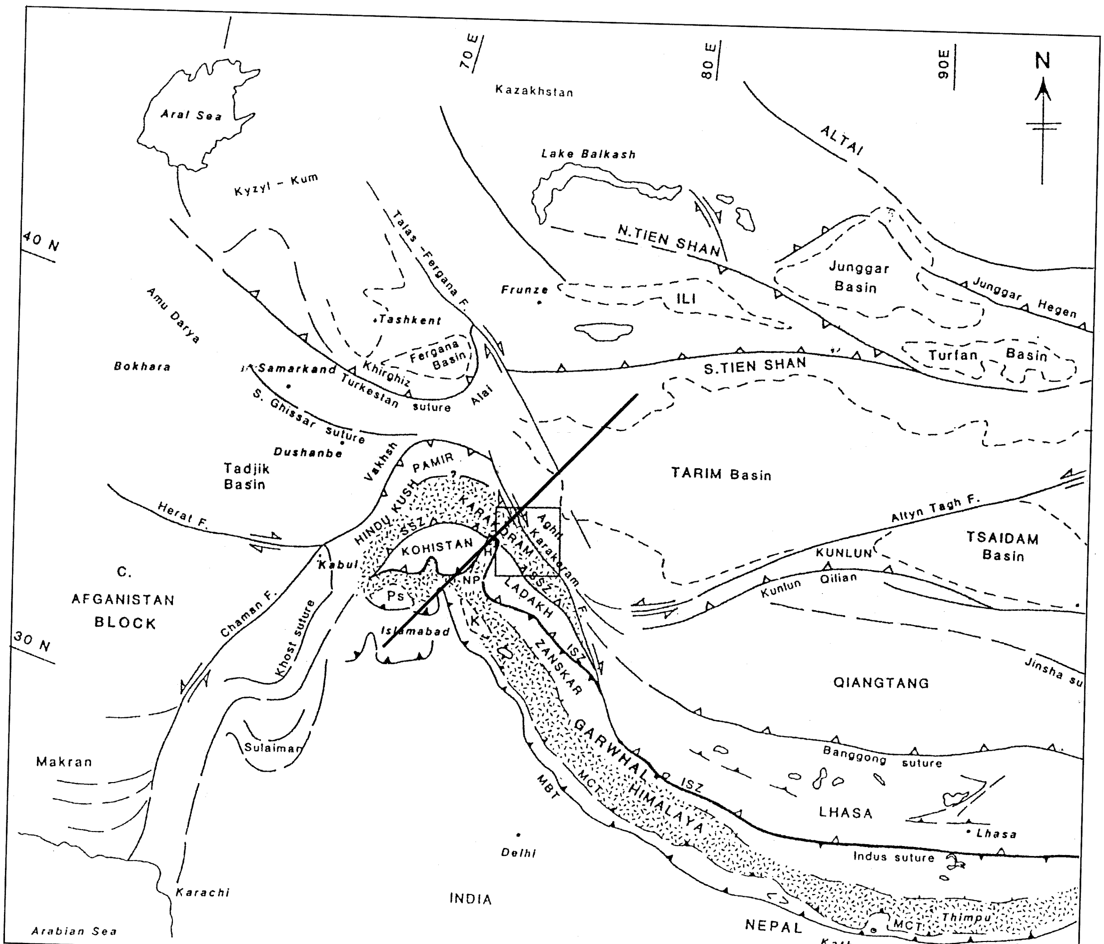

Fig. 2. Geological setting of Karakoram and SW China. The stippled areas indicate regions of post-collisional metamorphic rocks. ISZ = Indus Suture Zone; SSZ = Shyok Suture Zone; NP = Nanga Parbat; H = Haramosh; MBT = Main Boundary Trust; MCT = Main Central Thrust (after Searle, 1996). The straight line in SW-NE direction indicates the gravimetric profile studied in this paper.

model seems to be supported by evidence (Jin *et al.*, 1994) of a layered strength profile and of a decoupling zone in the lower crust.

In the foreland region of the Central Himalaya, the tectonic setting is similar to that of an active subduction zone: the Indian plate flexes down and forms the deep Ganga basin. Recorded seismic events are similar to those beneath the outer rise regions of oceanic trenches, suggesting flexure bending of the underthrust Indian plate.

Crustal thickening of the west side of the Tibetan Plateau is less extensive than on the east side (Le Pichon *et al.*, 1992). The dominant topographic features of this asymmetry are the Tarim basin and adjacent Tien Shan mountain range (fig. 3). The relatively low surface elevation of the Tarim basin suggests that there has been little crustal thickening, but – as we shall see later – gravity data are consistent with the crust being thick because the underlying lithosphere is rigid. The low rate of seismic activity supports this hypothesis by implying that deformation processes and faulting are negligible at present. A rigid Tarim implies, according to the thin-viscous-sheet model of Houseman and England (1996), that this block acts as a stress guide transmitting stress to the regions north of the basin, and as a secondary indenting block (Molnar and Tapponier, 1978; England and Houseman, 1985; Vilotte *et al.*, 1984, 1986). The predicted region of high strain immediately north of the indenter could be identified as the Tien Shan range which, according to Roecker *et al.* (1993), has a crustal thickness of 50 km and is one of the seismically most active areas of the collision zone (Molnar and Deng, 1984; Avouac *et al.*, 1993).

The Karakoram and Hindu Kush ranges have been under compression for 60 Ma with active continental crust subduction zones converging from both north (Tarim basin) and

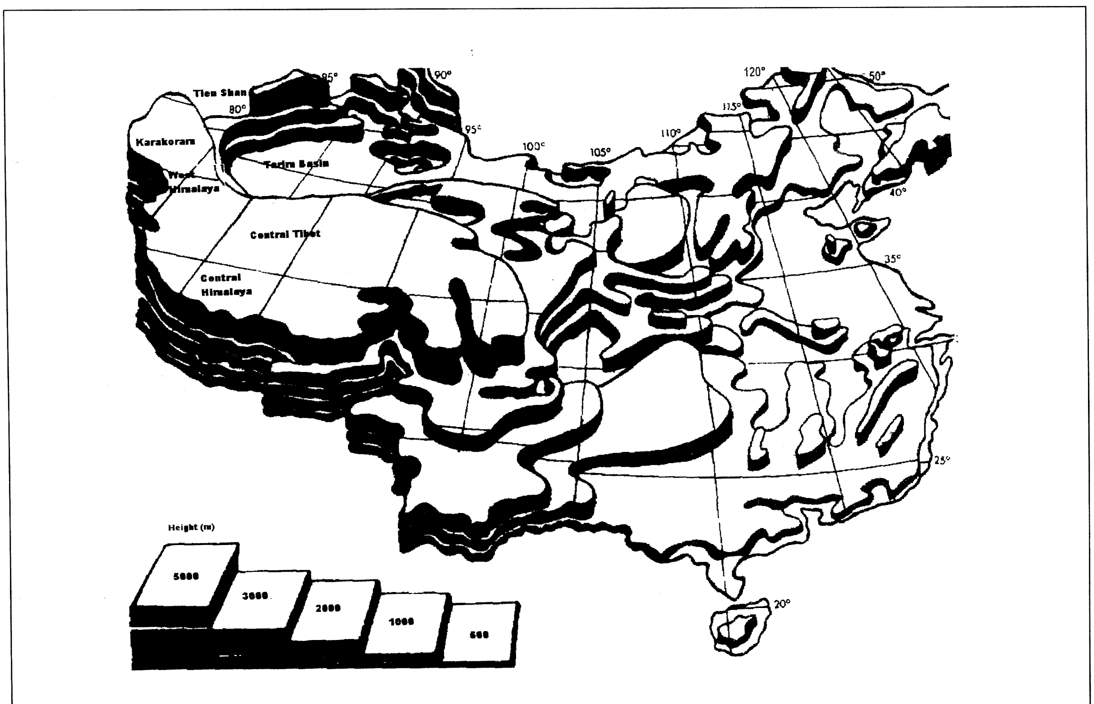

Fig. 3. Shematic elevation map of China (adapted from Hsieh, 1973).

south (Indian crust). Seismologic studies of the Hindu Kush have shown that intermediate depth earthquakes have been abundant at depths of 70-100 km and indicate subduction of crustal material from the south and the north. The upper mantle is relatively cold beneath the Karakoram, in contrast to the hot upper mantle in Central Tibet (Molnar, 1988).

2. Gravimetric constraints on the structure of the deep crust

The deep crustal structure can be constrained only by geophysical techniques. In particular, gravity anomalies show that the Moho steepens under the Karakoram southwards from the Tarim basin. Thus the Karakoram and Western Himalayas could be a unique example of continental crustal deformation with two actively subducting or underplating continental slabs converging from both north and south. Gravity data can be used to reconstruct the deformation geometry and constrain some of the rheological properties of the Indian and Tarim lithospheric units. In the Karakoram region Bouguer anomalies as low as -550 mgal (1 mgal $= 10^{-5}$ m/s^2) are recorded along a strip of approximately 100 km width aligned with the axis of the range. The flanking areas are associated with the Kun Lun and Aghil ranges to the north and the Kohistan Ladakh arc to the south. This arc has been interpreted as the remnant of an island arc sandwiched between the colliding Indian and Asian plates and then thrust onto the Indian plate (Tahirkeli *et al.*, 1979; Bard *et al.*, 1980, Malinconico, 1986). Bouguer anomalies along profiles across the Karakoram and Western Himalayas show large deviations form local isostatic equilibrium (Marussi, 1964; Molnar, 1988). Negative isostatic anomalies in the Pre-Cambrian Indian platform, Tethys Karakoram and Pre-Cambrian Tarim basin alternate with positive anomalies in Tethys Himalaya and Kun Lun. If the wavelength of this oscillatory pattern is considered as an indication of the elastic deformation of the lithosphere, and the lithosphere can be treated as a thin elastic plate floating on a denser, unviscid fluid (Lyon-Caen and Molnar, 1983, 1985; Karner and Watts, 1983), then gravity data can be used to constrain the elastic properties of the Indian and Tarim plates. Fortunately, gravimetric data exist in good amount in and around the Karakoram, in spite of its reputation as one of the most inaccessible regions of the world. Gravimetric surveys initiated with the pioneering work of Abetti and Alessio (1929) with the De Filippi expedition of 1913-1914, and of Amboldt (1948) with the Sino Swedish expedition of Sven Hedin, all with a pendulum apparatus. A major step forward came from 1954 to the 1970's, when Italian scientists led by Marussi and Caputo collected an impressive amount of data in Pakistan (Marussi, 1964; Ebblin *et al.*, 1983). Their work extends to the east the gravimetric surveys of McGinnis (1971) and was continued in the Karakoram and north of it by Caporali *et al.* (1991) Caporali (1993, 1995) with the expeditions led by prof. A. Desio in the framework of the Ev K2 CNR project. Although flexural models have been successfully developed for the Indo Gangetic foredeep up to the Kohistan Ladakh arc (Duroy *et al.*, 1989; Lillie, 1991), an understanding of the interaction between the Indian and Asian plate in the region of the Karakoram syntaxis, and of its differences from the collisional processes in the Central Himalaya, can be gained only by analysing together the flexure of the Indian and Tarim plates. To this purpose, following the approach of Lyon-Caen and Molnar (1983) and of Karner and Watts (1983), we introduce a flexural model in which, for each value of the horizontal abscissa x, the vertical profile $y(x)$ of the lithosphere is assumed to satisfy the differential equation

$$D\frac{d^4y}{dx^4} = P \tag{2.1}$$

where $D = \dfrac{ET^3}{12(1-\sigma^2)}$ is the flexural rigidity of the plate, defined in terms of the Young modulus E, the Poisson ratio σ and the thickness T of the elastic plate. The load $P(x)$ is the difference between the downward pressure of the topographic masses and of the sediments infilling the flexural basin, and the buoyancy force developed at a crust-mantle interface as a consequence of the density contrast:

$$P(x) = \rho_0 g\,[h(x) + y(x)] - \rho_m g y(x) \quad (2.2)$$

where g is the acceleration of gravity, ρ_0, ρ_m, are respectively the crust and mantle densities and $h(x)$ is the height of topography at the horizontal position x. Equation (2.1) shows that if the plate has zero flexural rigidity, the deformation caused by topographic load is entirely accomodated by vertical crustal thickening, according to the Airy model of isostasy. If the plate has non-zero flexural rigidity, then the deformation produced by a load applied vertically at a point on the plate will propagate in a wave-like manner along the elastic plate with a characteristic wavelength

$$\lambda = \left[\frac{ET^3}{3(1-\sigma^2)\Delta\rho g}\right]^{\frac{1}{4}} \quad (2.3)$$

where $\Delta\rho$ is the density contrast $\rho_m - \rho_0$. At each point of the profile the shearing force H and the bending moment M are given by

$$M = D\frac{d^2y}{dx^2}; \quad H = D\frac{d^3y}{dx^3}. \quad (2.4)$$

If, for the moment, the topographic load is ignored, the solution of eq. (2.1) is a linear combination of two damped waves traveling in opposite directions. Each wave depends on two arbitrary constants, amplitude and phase, which are defined by four boundary conditions:

$$y_e = Ae^{-\frac{x-x_0}{\lambda}} \cos\left(\frac{x-x_0}{\lambda} + \phi\right) + \\ + A'e^{\frac{x-x_0}{\lambda}} \cos\left(\frac{x-x_0}{\lambda} + \phi'\right) \quad (2.5)$$

where x_0 is an arbitrary origin. If the plate is broken at x_0, the bending moment must be zero there and the boundary condition

$$\left(\frac{d^2y}{dx^2}\right)_{x_0} = 0 \quad (2.6)$$

implies that $\phi = \phi' = 0$. If a shearing stress H_0 is acting on the plate at x_0, then the amplitudes of the wave are determined by the amplitude of the deformation at x_0

$$D\left(\frac{d^3y}{dx^3}\right)_{x_0} = H_0. \quad (2.7)$$

A specularly symmetric wave will result. More complicated waves can be built by linear superposition of these elementary solutions. For instance, if the plate is continuous at x_0 and a shearing stress H_0 is acting there, then the appropriate boundary conditions are (Heiskanen and Vening Meinesz, 1958):

$$\left(\frac{dy}{dx}\right)_{x_0} = 0; \quad D\left(\frac{d^3y}{dx^3}\right)_{x_0} = \frac{H_0}{2}. \quad (2.8)$$

The correct blend of elementary solutions can be constrained only by observational data, such as gravity anomalies. A solution to eq. (2.1) which includes the contribution of topography can be represented in the form

$$y(x) = \frac{\rho_0}{\Delta\rho} h(x) + y_e(x) \quad (2.9)$$

if the fourth derivative of h is negligibly small. This means that the dominant topographic features are at low frequency, compared to the characteristic wavelength (2.3), and that at this scale high frequency variations are small on average.

The deformation (2.9) produces a gravity anomaly which can be approximated by the Bouguer slab formula

$$\delta g(x) = 2\pi G\Delta\rho y(x) \quad (2.10)$$

where G is the gravity constant.

3. Numerical results

The data set used for this analysis consists of:

– No. 6 pendulum stations from the De Filippi expedition, as quoted by Marussi (1964);

– No. 4 pendulum stations from the Hedin expedition, as quoted by Marussi (1964);

– No. 197 stations measured in Pakistan by Marussi and co-workers (Marussi, 1964; Ebblin *et al.*, 1983) with the Worden gravimeter;

– No. 77 stations measured by Caporali and co-workers in 1988 (Aghil-Shaksgam region) with the Lacoste Romberg mod. G gravimeter (Caporali *et al.*, 1991);

– No. 68 stations measured by Caporali and co-workers in 1990 (Hispar-Biafo region) with the Lacoste Romberg mod. G gravimeter (Caporali, 1993);

– No. 12 stations measured by Caporali and co-workers in 1993 (Chogo Lungma-Haramosh region) with the Lacoste Romberg mod. G gravimeter (Caporali, 1995).

The data base additionally includes the astronomic and geodetic (by GPS satellites) coordinates of 16 stations measured by Caporali in 1988 and 1990. These are useful to compute geoid anomalies but will not be used in this context. For all available data terrain corrections have been computed using the ETOPO5 Digital Terrain Model published by the National Oceanic and Atmospheric Administration (NOAA). The resulting Bouguer anomalies (plate and topography) have been gridded with a minimum variance, «least squares collocation» method, under the assumption of an inverse square distance covariance function. Eventually, the grid was sliced along a profile extending from (33N, 72E) to (40N, 78E), that

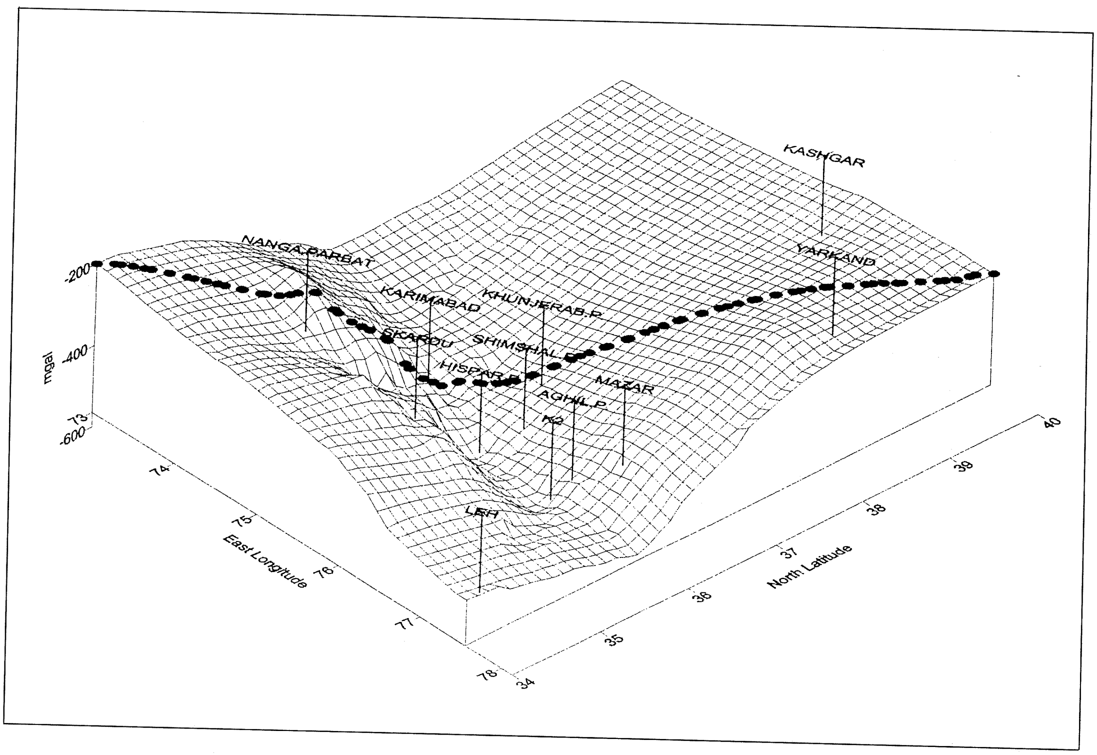

Fig. 4. 3-dimensional map of the Bouguer anomalies. The studied profile in fig. 3 is superimposed on the gravity anomaly map.

is from the Indo-Gangetic foredeep in the Islamabad area (Pakistan), in the NE direction to the Tarim basin, east of Kashgar in the autonomous province of Sin-Kiang, for a total of 955 km (fig. 4). The Bouguer anomalies and the ETOPO5 topography along the profile are shown in fig. 5a-c. For the first 316 km there is a descent of 9 mgal/km up to (74.0E, 35.4N) where the descent temporarily stops, to resume at the steep rate of -23 mgal/km. The minimum value of -532 mgal is reached at (74.9E, 36.3N) (433 km). The anomalies increase to less negative values at a rate of 8 mgal/km up to (76.7E, 38.4N) (km 716), and from here to the end of the profile the increase is of 0.3 mgal/km. The least squares fit of the model of eqs. (2.5) and (2.10) to the Bouguer anomalies, with the assumption that the lithospere is broken at an unknown point x_0 into an Indian and Tarim plate is described by a deformation

$$y(x) = \frac{\rho_0}{\Delta\rho} h(x) + \frac{H_0}{g\lambda} e^{\frac{x-x_0}{\lambda}} \cos\left(\frac{x-x_0}{\lambda}\right) \quad (3.1)$$

with values of the model parameters given in table I.

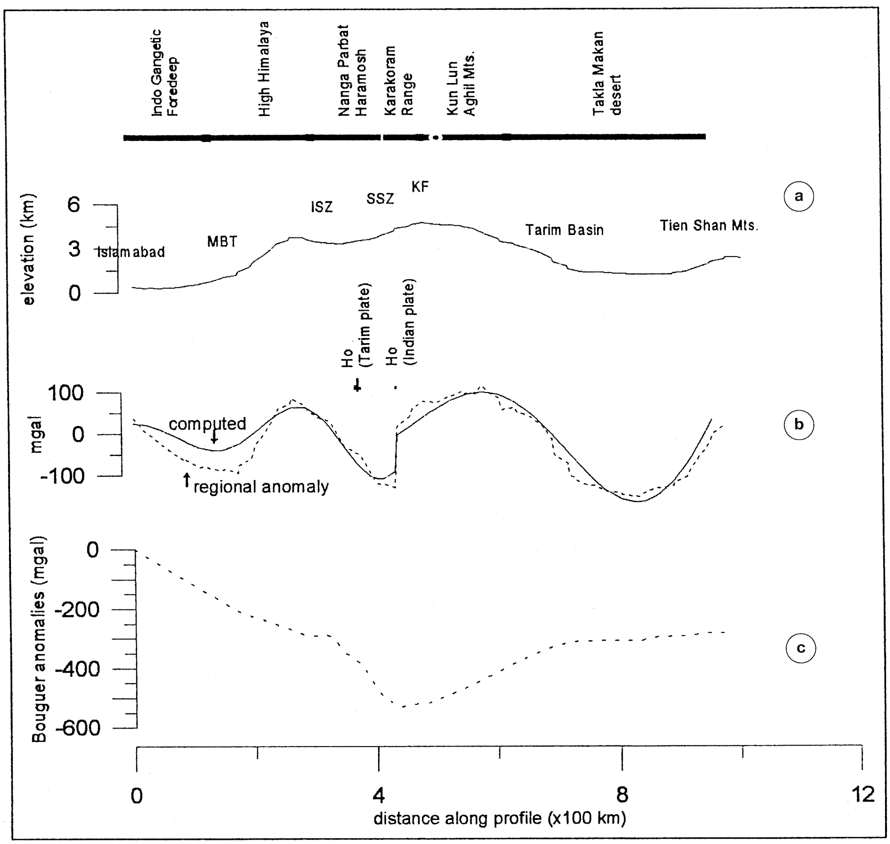

Fig. 5a-c. a) Height of topography along the chosen profile; b) regional isostatic anomaly and computed anomaly based on an elastic flexural model; c) observed Bouguer anomalies and shape of the total lithospheric deformation. In (b) regional anomalies have been computed using different values of the crust density ρ_0. MBT = Main Boundary Thrust; ISZ = Indus Suture Zone; SSZ = Shyok Suture Zone; KF = Karakoram Fault (see fig. 2).

Table I. Numerical values of the model parameters in eq. (3.1).

	ρ_0 (kg m^{-3})	H_0 (N m^{-1})	D (N m)	x_0 (km)	$\Delta\rho$ (kg m^{-3})	Mean and rms (mgal)
Indian plate	2.300	7 10^{12}	5 10^{24}	400 (74.6E, 36.0N)	405	-19 ± 28
Tarim plate	2.950	7 10^{12}	7 10^{25}	310 (74.0E, 35.3N)	400	-19 ± 28

4. Discussion

The continental crust is in general too buoyant to be subducted, yet in some regions of continental collision the continental crust has been subducted to considerable depth. The Karakoram is an active convergent zone on both the northern side, where the Tarim basin continental plate is underthrusting beneath the Kun Lun and northern margin of Karakoram (Lyon-Caen and Molnar, 1984), and southern side, where a crustal thickness of 30-38 km beneath the Indo-Pakistan foreland (Kaila, 1982) increases to 65 km beneath High Himalaya and Karakoram (Molnar and Chen, 1983; Molnar, 1984, 1988). The modeling of gravity anomalies in terms of a thin elastic plate helps in constraining the geometry of the post-collisional deformation, and hence the rheological and load parameters on which the model depends. In particular, the thickness of the plate, the density of the topographic load and of the material infilling the flexural basin, the crust-mantle density contrast, the value of the load at the edge of the plate and the location of this edge are reasonably well constrained by the gravimetric data, in the sense that changes from the values indicated in table I result in values of the root mean square dispersion of the post fit residuals higher than those indicated in the last column of table I. Both plates have a very high rigidity. To account for the negative values of the regional isostatic anomaly in the Kohistan Ladakh area it is necessary to introduce a shear stress which bends both plates down. Lyon-Caen and Molnar (1985) proposed that this shear stress could be explained by modeling the cold part of the Indian (or Tarim) lithosphere as a mass anomaly with density contrast $\delta\rho$ relative to the surrounding mantle. If h is the thickness and L is the down-dip extent, then $H_0 = L\delta\rho g h$. Assuming $L = 200$ km and $\delta\rho = 50$ kg m^{-3}, the thickness h of the anomaly would result 70 km, nearly the elastic thickness of the Indian plate.

4.1. *Indian plate*

Equation (3.1) combined with the Indian plate parameters in table I represents a plate loaded by light material. The oscillatory pattern of the regional isostatic anomalies (fig. 5b) is reproduced by the damped wave (3.1). The oscillation tends to damp out moving away south of the plate. The location x_0 of the shear stress H_0 coincides geographically with the northern termination of the Nanga Parbat-Haramosh syntaxis, which separates Kohistan from Ladakh (fig. 2). Possible explanations of the load can be the presence of an infracrustal obducted block (*e.g.*, ophiolites), or an increase in density related to granulite-to-eclogite phase transition (Searle, 1996). The equivalent elastic thickness is 87 km and there is no apparent need to introduce a lower rigidity – or thickness – at the end of the plate, as postulated by Lyon-Caen and Molnar (1985). From fig. 5b an explanation is obtained of the negative isostatic anomaly beneath Karakoram, first noticed by Marussi and tentatively explained by him in terms of a low density granitic intrusion: the negative anomaly is simply a consequence of an overthickened crust, relative to isostatic equilibrium, due to flexure.

4.2. *Tarim plate*

The regional anomalies shown in fig. 5b have been obtained assuming that the topographic load, the first term in eq. (3.1), has a density of 2950 kg m^{-3} which is considerably higher than in the Indian plate. The presence of granulite facies rock has been suggested in the Tarim crust (Searle, 1996). England and Richardson (1977) have argued that low density amphibolite-granulite facies rocks (2850 kg m^{-3}) can transform into higher density eclogite facies rocks (3150 kg m^{-3}). If the subduction is fast enough, the subducting crust maintains a low temperature and the increased density resulting from granulite-to-eclogite phase transition would decrease the buoyancy and facilitate the subduction of the Tarim continental crust. The longer wavelength of the elastic perturbation in the Tarim plate, and hence the higher rigidity relative to the Indian counterpart is evident in fig. 5. As shown by eq. (2.3) the wavelength is affected both by the thickness and by the density contrast. The density contrast is very nearly the same as for the Indian plate (table I). Thus the higher rigidity must result in a considerably higher thickness, of the order of 200 km. The wave-like response of the Tarim lithosphere to the applied loads implies a relatively thinner crust – with respect to isostasy – in the Kun Lun and Aghil range, and a likewise thicker crust in the Tarim. Figure 5b suggests that the amplitude of the oscillation increases proceeding north, toward the Tien Shan, implying that an important source of load controlling the amplitude and phase of the wave is located there, besides Karakoram. The assumption of a semi-infinite plate loaded at one free edge, applicable in the Indian shield, should be relaxed in the modeling of the Tarim plate, as this plate is most probably loaded at both ends.

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