Cenozoic deformation of the Tarim plate and the implications for mountain building in the Tibetan Plateau and the Tian Shan

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The Tarim basin in NW China developed as a complex foreland basin in the Cenozoic in association with mountain building in the Tibetan Plateau to the south and the Tian Shan orogen to the north. We reconstructed the Cenozoic deformation history of the Tarim basin by backstripping the sedimentary rocks. Two-dimensional and three-dimensional finite element models are then used to simulate the flexural deformation of the Tarim basin in response to the sedimentary loads and additional tectonic loads associated with overthrusting of the surrounding mountain belts. The results suggest that uplift of both the western Kunlun Shan and the Tian Shan orogens started during or before Oligocene. Since the Miocene, mountain building accelerated in the western Kunlun Shan but showed segmented development in the Tian Shan. The results also indicate reduced tectonic load along the Altyn Tagh fault since late Miocene, consistent with geological and geophysical evidence of the fault cutting the entire lithosphere and migrating southward during its cause of the Cenozoic evolution.


1. Introduction

The continued plate convergence between India and Asia since 50–70 million years ago resulted in the uplift of the Himalayan-Tibetan plateau, reactivated the Tian Shan orogen, and affected regions perhaps as far north as the Baikal [Molnar and Tapponnier, 1975; Yin and Harrison, 2000]. Following the pioneering work of Argand [1924] and Gansser [1964], significant progresses have been made on the geology and tectonics of the Himalayan-Tibetan orogen [Dewey and Burke, 1973; Molnar and Tapponnier, 1975; Allegre et al., 1984; Armijo et al., 1986; Burchfiel et al., 1992; Harrison et al., 1992; Avouac and Tapponnier, 1993; Nelson et al., 1996; England and Molnar, 1997; Larson et al., 1999; Yin and Harrison, 2000]. However, some fundamental questions, such as how stress has propagated across the collisional zone and how strain was partitioned between crustal thickening and lateral extruding of the lithosphere, remain controversial. One end-member model, approximating the Eurasia continent as a viscous thin sheet indented by the rigid Indian plate [e.g., England and McKenzie, 1982; England and Houseman, 1986, 1988], predicts progressively northward propagation of crustal thickening and plateau uplift, with limited lateral lithosphere extrusion. Another end-member model, based on indentation of a plastic Asian continent, suggests that a major part of the convergence was taken up by eastward extrusion of the lithosphere along major strike-slip faults activated by the collision [Tapponnier et al., 1982]. Constraints on the timing and spatial distribution of crustal deformation in the Tibetan plateau and surrounding regions, which are essential for illumination of the dynamics of collisional crustal deformation, remain sparse and often controversial [Coleman and Hodges, 1995; Harrison et al., 1995; Chung et al., 1998; Yin and Harrison, 2000].

The Cenozoic deformation history of the Tarim basin in northwestern China (Figure 1) may provide useful insights into the problems of stress propagation and strain partitioning following the Indo-Asian collision. Located between the Tibetan plateau and the Tian Shan orogen, the relatively rigid Tarim block behaved as a secondary “indenter,” transmitting collisional stresses to the Tian Shan [Molnar and Tapponnier, 1975; Neil and Houseman, 1997]. The Cenozoic deformation of the Tarim basin was closely coupled with mountain building in the northern Tibetan plateau and the Tian Shan, which caused major subsidence of the Tarim basin near its northern and southwestern margins. The nearly 10-km-thick Cenozoic sedimentary rocks derived from the surrounding mountain belts [Li et al., 1996; Jia, 1997] provide useful constraints on the deformation history of the Tarim basin and mountain building in the surrounding orogens.

In this study we investigate the Cenozoic deformation history of the Tarim basin and its implications for mountain building in the northern Tibetan plateau and the Tian Shan orogen. The basement deformation during three periods (Paleogene, early-middle Miocene, late Miocene-Pliocene) is reconstructed by backstripping the sedimen-
2. Geological Background

[5] The rhomb-shaped Tarim basin is surrounded by the Tian Shan ("Shan" means mountain ranges in Chinese) to the north, the western Kunlun Shan and the Altyn Tagh Shan of the northern Tibetan plateau to the south and southeast, and the Pamir to the west (Figure 1a). These mountain ranges originated in the late Paleozoic [Allen et al., 1993] or even earlier [Nakajima et al., 1990] and rejuvenated following the India-Eurasia collision 50–70 million years ago [Molnar and Tapponnier, 1978; Windley et al., 1990; Yin and Harrison, 2000]. Cenozoic deformation of the Tarim basin was coupled with mountain building in the surrounding orogenic belts through the bounding fault systems: the western Kunlun thrust belt and the Altyn Tagh fault to the south, and the southern Tian Shan thrust belts to the north (Figure 1b).

2.1. Western Kunlun Shan and Western Kunlun Thrust Belt

[6] The western Kunlun Shan is composed mostly of Precambrian schist and gneiss overlain by Paleozoic and Mesozoic carbonate and clastic sequences [Yin and Harrison, 2000]. The present western Kunlun Shan is defined as a Cenozoic fold and thrust belt in the north and the Karakaxi fault in the south (Figure 1b). Southward subduction of the Tarim plate beneath the northern Tibetan plateau is suggested as the cause of the western Kunlun thrust belt [Lyon-Caen and Molnar, 1984]. Geologic mapping in the eastern part of the western Kunlun thrust belt [Cowgill et al., 2000] suggest that the magnitude of north-south shortening across the eastern part of the thrust belt is between 50 and 100 km. This magnitude appears to increase westward based on paleomagnetic studies [Yin and Harrison, 2000].

2.2. Altyn Tagh Fault

[7] The Altyn Tagh fault connects the western Kunlun Shan in the west and the Nan Shan in the east. It extends over 1200 km and bounds the northern boundary of the Tibetan Plateau (Figure 1b).

[8] The Altyn Tagh fault may have played a key role in eastward extrusion of Tibet during the Indo-Asian collision [Tapponnier and Molnar, 1977]; however, many aspects of the fault remain controversial. The displacement history of the Altyn Tagh fault has been derived mainly from off-setting of Quaternary features. The estimated Quaternary left-slip rates range from 4 to 30 mm/yr in various segments of the fault [Molnar et al., 1987; Peltzer et al., 1989; Meryer et al., 1996]. Estimates of the total magnitude of left slip on the Altyn Tagh fault varies from ~1200 km to ~200 km [Peltzer and Tapponnier, 1988; Burchfiel et al., 1991; Zheng, 1994]. The time of initiation of the Altyn Tagh fault is suggested to be Oligocene [Wang, 1990], middle Miocene [Peltzer and Tapponnier, 1988; Yin and Nie, 1996] or Pliocene [Bally et al., 1986; Burchfiel et al., 1991] based on various geological arguments.

2.3. Tian Shan and Southern Tian Shan Thrust Belt

[9] Bordering the Tarim basin to the north is the Tian Shan mountain belt, which extends broadly east-west for over 2500 km across central Asia, with peaks exceeding 7000 m and basins lower than 100 m below sea level (Figure 1). A middle Paleozoic Tian Shan orogen may have developed from one or more arc-continental collisions [Coleman et al., 1992; Allen et al., 1993; Carroll et al., 1995], followed by reactivation of the older faults caused by successive collision of island-arc systems onto the southern margin of Asia in the late Paleozoic and the Mesozoic [Hendrix et al., 1992; Sobel, 1995].

[10] During the Cenozoic the Tian Shan was rejuvenated by the Indo-Asian collision [Tapponnier and Molnar, 1979; Burchfiel et al., 1991; Lu et al., 1994]. The present Tian Shan is flanked by east-west trending, active thrust systems on its north and south sides [Tapponnier and Molnar, 1979; Burchfiel et al., 1991; Avouac and Tapponnier, 1993; Molnar et al., 1994; Abdarakhmatov et al., 1996]. Between the Tian Shan and the Tarim basin is the southern Tian Shan thrust belt, which may be divided into four segments based on their deformation styles and structural trend [Yin et al., 1998]. From west to east (Figure 1b), they are (1) the Kashi-Aksu thrust system, (2) the Baicheng-Kuche thrust system, (3) the Korla transfer system, and (4) the Lop Nor thrust system [Yin et al., 1998; Allen et al., 1999]. The westernmost Kashi-Aksu system is characterized by the occurrence of evenly spaced (12–15 km) imbricate thrusts. The Baicheng-Kuche and Korla systems are expressed by a major north dipping thrust (the Kuche thrust) that changes its strike eastward to become a NW striking oblique thrust ramp (the Korla transfer zone). The Lop Nor system consists of widely spaced thrusts, all involved with basement rocks. The age of initial thrusting is estimated to be 20–25 Ma [Hendrix et al., 1994; Yin et al., 1998].

2.4. Tarim Basin

[11] The Tarim basin has a complex and protracted history of deformation since the late Precambrian [Hendrix et al., 1992; Wang et al., 1992; Carroll et al., 1995; Li et al., 1996]. The basement consists of Archean and Proterozoic metamorphic rocks [Tian et al., 1989; Chai et al., 1992]. The sedimentary cover, including Proterozoic and Phanerozoic sequences, ranges from 17 km within major depression centers to 5 km in the central uplift [Li et al., 1996]. The Paleozoic sequence is dominated by platform carbonates and fine-grained clastics [Carroll et al., 1995]. Mesozoic strata of north central Tarim are interpreted to record deposition as the result of multiple episodes of contraction and uplift of the margins of the basin [Hendrix et al., 1992]. These periods of conglomerate deposition may be related to the accretion of the Qiangtang block in the Permian-Jurassic, the Lhasa block in the latest Jurassic, and the Kohistan-
Dras arc complex in the Late Cretaceous [Hendrix et al., 1992]. Since then it has been largely associated with foreland depressions caused by basinward thrusting and overloading of the Tian Shan in the north and the western Kunlun Shan in the south [Jia, 1997; Metivier and Gaudemer, 1997].

During the Cenozoic the Tarim basin developed into a complex system of foreland basins largely in response to the Indo-Asian collision. Basinward thrusting of the Tian Shan and western Kunlun Shan caused further subsidence of the foreland depressions where Cenozoic sedimentary rocks are up to 10 km thick (Figure 2). During the Eocene, sedimentation was mainly centered in the Hotan (or the southwestern) depression adjacent to the western Kunlun thrust belt. By the end of Pliocene, both the Hotan and the Kuche depressions adjacent to the Tian Shan orogen had

Figure 1. (a) Topographic relief map of the Tarim basin, NW China, and surrounding regions. The dots mark the approximate locations of some of the industrial wells drilled before 1990 [Zhao et al., 1997]. (b) Simplified geological map of the Tarim basin and surrounding regions [after Yin et al., 1998].
received >5000 m sediments (Figure 2). Cenozoic sediments in the basin were mainly terrestrial facies. The sedimentary strata in the Tarim basin have been well documented from dense oil wells drilled over the past decades [Jia, 1997].

3. Basement Deformation History

[13] To understand the causes of Cenozoic deformation of the Tarim basin we need to reconstruct the basement deformation history. The basic data are the isopach of the Cenozoic sedimentary rocks. Petroleum and gas exploration in the Tarim basin for the past three decades has produced detailed information about the sedimentary cover of the basin. Based on industrial drilling, shallow seismic reflection, and geological studies, the Chinese Petroleum and Gas Corporation produced the isopach maps [Jia, 1997] (Figure 2) that are used in this study. Because of industrial secrecy, no details of the location of the oil wells or seismic lines used to construct the isopach maps were given by Jia [1997]. We obtained the locations of 55 industrial oil wells drilled before 1990 from Zhao et al. [1997] and show them in Figure 1a. This information is incomplete, and the isopach maps (Figure 2) were constructed in 1995, and presumably were based on data from more drill wells than those shown in Figure 1a. Both Jia [1997] and Figure 1a indicate that there are no drill well data in the Kepingtage area. The isopachs in the Kepingtage area are generated by interpolation and therefore are not used in our calculations of basement deformation or sedimentary loads. The general isopach patterns in Figure 2, including the separated depression centers at Yecheng and Hotan in the Paleogene, are consistent with other published data [Metivier and Gaudermer, 1997; Allen et al., 1999; Metivier et al., 1999]. Whereas the lack of detailed information of the source data makes it difficult to assess the errors in these isopachs, the isopach maps in Figure 2 are probably sufficient for a first-order modeling.

[14] The depth of the basement (relative to present basin surface) during the youngest period (late Miocene-Pliocene) may be estimated from the thickness of sedimentary rocks formed during this time (Figure 2c). Basement deflection during the earlier periods can be reconstructed using backstripping [Lerche, 1990; Makhouse et al., 1997]. As illustrated in Figure 3, the original depth of layer A can be restored by stripping off the overlying layer B and correcting for compaction. Assuming the porosity $\phi$ decreases with depth following the empirical relation [Falvey and Middleton, 1981]:

$$\phi = \phi_0 \exp(-cz),$$

where $z$ is depth and $c$ is a constant, mass conservation of the solid material in layer A can be written as

$$z_2 - z_1 + \frac{\phi_0}{c} [\exp(-cz_2) - \exp(-cz_1)] = d + \frac{\phi_1}{c} [\exp(-cd) - 1],$$

where $z_1$ and $z_2$ are the depth to the bottom of layer B and layer A, respectively, and $d$ is the original thickness of layer A before layer B was deposited (Figure 3). In the calculations we used the values of $\phi_0 = 0.429$ and $c = 0.675$ $\text{km}^{-3}$ determined by Liu et al. [1996] based on drill hole data from the Tarim basin.

[15] Figure 4 shows the reconstructed history of basement deformation for the Paleogene and the early-middle Miocene. During Paleogene the Hotan depression was separated from the broad southwest depression centered near Yecheng, where the basement depression was up to 2200 m (Figure 4a). Minor depression centers also developed along the southeastern margin of the Tarim basin, near Minfeng and Qiemo. By early Miocene both the southwestern and the Kuche depression centers deepened and propagated basinward (Figure 4b). Note that the result in Figure 4b is the incremental subsidence during the period of early-middle Miocene. We consider the Paleogene sedimentary rocks as part of the basement by this time. The Hotan depression became part of the southwestern depression where the maximum subsidence reached 3500 m near Yecheng. The incremental basement subsidence since late Miocene can be estimated directly from the isopach of this period (see Figure 2c). It is clear that basement subsidence accelerated since Miocene. Near Yecheng the incremental basement depression was up to 5000 m during late Miocene-Pliocene. Similar basement depression occurred in the Kuche depression, which branched southwestward to form another depression center near Aksu (Awati) where the subsidence reached 5000 m. Not much subsidence occurred along the southeastern margin of the Tarim basin during late Miocene-Pliocene.


[16] It is clear from Figures 2 and 4 that basement deformation of the Tarim basin through the Cenozoic was dominated by the development of a two major foreland basins adjacent to the western Kunlun Shan and the Tian Shan mountain ranges. The theory of foreland basin formation resulting from flexural deflection of the lithosphere has been well developed [Turcotte, 1979; Beaumont, 1981]. Flexure of a thin elastic plate can be described by [Turcotte, 1979]

$$D\nabla^4 w + p\nabla^2 w = q,$$

where $w$ is deflection, $\nabla^2$ and $\nabla^4$ are the Laplace and the biharmonic operators, respectively, $p$ is horizontal load and $q$ is vertical load per unit area on the lithosphere. For the flexure of continental lithosphere the vertical load may be written as $q = q_t + p_g g w_0 - \rho_m g w$, where $q_t$ is the tectonic load, and $(\rho_g g w_0 - \rho_m g w)$ is the net force of sediment loads and the buoyancy force arising from the displaced mantle. Here $\rho_g$ is the density of the overlying sedimentary material, $g$ is gravitational acceleration, $w_0$ is the thickness of the sediment layer, and $\rho_m$ is the density of the underlying mantle. Thus equation (3) could be written as

$$D\nabla^4 w + p\nabla^2 w + \rho_m g w = q_t + \rho_g g w_0.$$
identical and equation (4) can be written in the more familiar form [Turcotte, 1979]:

\[ D \nabla^4 w + p \nabla^2 w + (\rho_w - \rho_r)gw = q_t. \]

However, equation (4) is more useful for illustrating our approach: we used \( \rho_s gw_0 \) as the sedimentary load because \( w_0 \) is derived directly from the sedimentary records, whereas \( w \) is the theoretical basement deflection. The parameter \( D \) is the flexural rigidity:

\[ D = \frac{Eh^3}{12(1 - v^2)}, \]
where $E$ is the Young’s modulus, $\nu$ is the Poisson’s ratio, and $h$ is the thickness of the elastic plate. For lithospheric flexure an effective elastic thickness, $T_e$, may be used to replace $h$ [Watts, 1976; McNutt, 1984]. Equation (3) can be modified for other rheologies [Turcotte, 1979; McMullen et al., 1981; Burov and Diament, 1992]. Various rheologies have been applied to foreland basins [Beaumont, 1981; Jordan, 1981; Royden and Hodges, 1984; Patton and O’Connor, 1988] and other tectonic settings [Walcott, 1970; Melosh, 1978; Turcotte, 1979; Karner and Watts, 1983; Watts and ten Brink, 1989], with varying degrees of success. In many cases an elastic rheology has been shown to be a good first-order approximation [Turcotte, 1979; Burov and Diament, 1992], and the effects of various factors on the lithospheric rheology, such as thermal structures and mechanical weakening, may be lumped into $T_e$ [McNutt, 1984; Burov and Diament, 1995]. Note that equations (3) and (4) are for thin elastic plate only, and we used these

Figure 3. Sketch showing reconstruction of basement subsidence using backstripping. The original depth of layer A is restored through peeling off layer B and correcting for compaction.

Figure 4. Basement deformation in (a) Paleogene and (b) early-middle Miocene derived from backstripping. Basement deformation for the youngest period (late Miocene-Pliocene) can be estimated directly from the isopach in Figure 2c. The two straight lines in (a) show the location of the 2-D models in Figures 6 and 7.
equations here to illustrate the roles of sedimentary and tectonic loads in basement deformation. Although the thin elastic plate may be a viable approximation of the Tarim plate, we have developed a fully three-dimensional model based on the elastic theory \cite{Timoshenko1970}. The models solve for the full sets of 15 linear equations (six stress and six strain components plus three displacement components) without requiring the thin-plate approximation. We also used a viscoelastic model and will compare the results below.

The major loads causing the flexural foreland basin development in the Tarim basin include sedimentary loads distributed across the basin and tectonic loads near the margins of the basin associated with overthrusting of the surrounding mountain belts (Figure 5). Horizontal load is usually neglected in elastic flexure models because of the unreasonably high values of the load (6.4 GPa for a 50 km thick elastic plate) required to cause buckling \cite{Turcotte1982}. However, Karner \cite{Karner1986} pointed out that horizontal load could have significant effects for plates with preexisting deformation, such as sedimentary basins. Compressional horizontal load tends to amplify the plate deformation; the amount of amplification is a function of the elastic thickness and the wavelength. We do not include horizontal load in the model because its value is difficult to constrain and doing so would increase the number of free parameters, thus introduce large uncertainties to the tectonic loads, the major target of this study. Applying Karner’s \cite{Karner1986} results to the Tarim basin, we estimated that <5% of error may be introduced by neglecting the horizontal load. Although bending moments may also contribute to flexure, they are difficult to constrain from geological data and therefore are not included in our calculations.

Because the spatial distribution of the sedimentary rock (sedimentary loads) for each of the three periods are known (see Figure 2), the optimal tectonic loads for each period may be derived from matching the observed basement deformation for a given rheological structure. For this study we developed both two-dimensional (2-D) and three-dimensional (3-D) finite element models using FEPG (available at http://www.fegensoft.com/english/index.htm), an automatic finite element codes generating system \cite{Yang1998}. Both the 2-D and 3-D models solve for the full set of linear elastic equations without requiring the thin-plate approximation. We verified the numerical results by comparison with the analytic solutions of numerous simple 2-D problems.

4.1. Two-Dimensional Models

Most previous studies of foreland basin formation were based on 2-D models \cite{Beaumont1981, Burov1992}, because they are easy to compute and useful for illustrating the effects of major model parameters. We conducted a series of 2-D finite element modeling to explore the model parameters and the relative roles of sedimentary and tectonic loads in the basement deformation of the Tarim basin. Figure 6 compares the predicted and observed Paleogene basement subsidence along a profile from Yecheng to Kuche. Using the sedimentary loads along the profile (see Figures 4a and 2a), we could determine the required tectonic loads by fitting the observed basement deformation. The problem here is $T_e$, the effective elastic thickness of the Tarim plate, that is also unknown. There are
certain trade-offs between the $T_e$ value and the tectonic loads (see below), and the method of trial and error was used to find the best combination of these two variables. The density of the sedimentary material and the mantle are assumed to be 2600 kg/m³ and 3300 kg/m³, respectively, and the other elastic constants are represented implicitly by the $T_e$ value. The Winkler spring mattress is used at the bottom to simulate restoring body forces induced by displacement of density boundaries [Williams and Richardson, 1991]. For this profile the best fit is found when $T_e = 60$ km (Figure 6), and nearly 70% of the observed basement subsidence can be explained by the sedimentary loads. The largest misfits are near the margins of the basin where additional loads may result from overthrusting of the adjacent fold and thrust belts, i.e., tectonic load. Applying the optimal tectonic loads significantly improved the fit (Figure 6).

We should note that the boundary between the Tarim basin and the surrounding mountain ranges are not clear-cut. Thus, it is difficult to completely separate the sedimentary load from tectonic load. Some of the sedimentary rocks near the margins of the basin, while being counted as sedimentary load in our model, are part of the fold and thrust belts associated with mountain building and hence perhaps should be regarded as tectonic load.

For elastic flexures one controlling parameter is the effective elastic thickness. The physical effects of various $T_e$ can be seen from Figure 7, which shows the predicted and observed Paleogene basement deformation along a profile from Qiemo to Kuche (see location in Figure 4a). A thicker plate (larger $T_e$) allows the effects of tectonic load to propagate further into the basin, resulting in smoother and broader flexure. A thinner plate, on the other hand, tends to localize subsidence to where the loads are applied.

The good fits between the predicted and observed basement deformation in Figures 6 and 7 also illustrate the limitations of 2-D models for the Tarim basin. In 2-D models we are free to choose the tectonic loads and $T_e$, consequently we can always find a combination of these parameters that produce close fit to the observed basement deformation. However, different results may be obtained for different profiles. For example, the optimal tectonic load at Kuche is 3 MPa over an 80-km width for the Yecheng-Kuche profile (Figure 6), but 45 MPa over a 20-km zone for the Qiemo-Kuche profile (Figure 7). Furthermore, the optimal $T_e$ is $\sim 60$ km for the Yecheng-Kuche profile but 15 $\sim 30$ km for the Qiemo-Kuche profile, and there is no geological evidence supporting such a sharp lateral change of the lithospheric rheology over the Tarim basin. Part of the problem may be the location of the 2-D profiles. The Yecheng-Kuche profile is perpendicular to the Southwestern depression but not to the Kuche depression, thus is not an ideal representation for deformation in the Kuche region. The particular geometry of the Tarim basin, with limited spatial extension in all directions, and the complex internal deformation patterns, are difficult for 2-D representations. In the following we shall focus on three-dimensional models.

### 4.2. Three-Dimensional Models

Figure 8 shows the numerical mesh of the 3-D model. The rheology of the Tarim plate is assumed to be either elastic or viscoelastic. We focus on elastic flexure here and discuss the results of viscoelastic models later. The elastic model of the Tarim plate has two layers (Figure 8). The Young’s modulus is taken to be $3 \times 10^{10}$ Pa for the top layer and $5 \times 10^{10}$ Pa for the lower layer. The somewhat lower...
value for the top layer is to reflect a relatively weaker upper crust because of fractures and the sedimentary cover. We used multiple layers in the model to explore the effects of rheologic variations with depth, although for elastic models the results are dependent mainly on the total effective elastic thickness. The Poisson’s ratio is taken as 0.25 for all elements. These elastic constants are within the typical values for crustal rocks [Turcotte and Schubert, 1982]. The model includes 3102 nodal points and 1932 isoparametric solid elements, providing a horizontal element size of 25–30 km. We divide the margin zone of the Tarim basin into 14 domains. Domains KL1 – KL4 are along the western Kunlun Shan thrust belt, AT1–AT3 are along the Altyn Tagh Shan and fault system, and TS1 – TS7 line from west to east along the Tian Shan thrust belt. Within each domain the tectonic load is assumed to be uniform for a given period, and the optimal values are obtained by regression stated below. The domain division roughly reflects the general geological settings, but the number of boundary domains and their boundaries are somewhat arbitrary.

[24] We constrain the horizontal displacement along the eastern side, which is allowed to move only vertically. This boundary condition prevents artificial rigid-body translation and rotation. Other sides of the model plate are free, and the Winkler spring mattress is used at the bottom to simulate restoring body forces induced by displacement of density boundaries [Williams and Richardson, 1991].

[25] Similar to the approach used in 2-D models, we determine the optimal tectonic loads by fitting the predicted flexure to the observed basement deformation. For a given rheology structure, the tectonic load on each of the boundary domains during each geological period may be determined. A major problem we face here, as in 2-D models, is that we have to determine the trade-off between the tectonic loads and the effective elastic thickness. In 2-D models we used the method of trial and error to look for the combination of tectonic loads and \( T_c \) that produces the best fit to the observed basement deformation. This caused the problems of having large variations of \( T_c \) along different profiles, as discussed above. The situation is greatly improved here by having an additional constraint, the requirement of linearity. Because elastic flexure is a linear process, the basement deformation should be the sum of deformation caused by the sedimentary loads over the whole basin and the tectonic load on each of the boundary domains. Thus we can choose the \( T_c \) value that gives the highest linearity, determine by the F-test (see below), and obtaining the optimal tectonic load on each of the margin domains by linear regression.

[26] Assuming \( W(x, y) \) is the response to tectonic load of unit magnitude on the \( i \)th boundary domain, \( W_0(x, y) \) is the response to sedimentary load, and \( W^*(x, y) \) is the observed basement deflection (see Figure 4), the linear assumption can be expressed as

\[
W^*(x, y) = W_0(x, y) + \sum_{i=1}^{14} a_i W_i(x, y),
\]

where \( a_i \) is the magnitude of tectonic load on the \( i \)th boundary domain. We use the method of multivariate linear regression to determine the values of \( a_i \), and the F-test to test the linear correlation between \( W^*(x, y) \) and the tectonic loads on the 14 boundary domains. A high confidence level from the F-test will indicate the assumption of linearity is acceptable.

[27] We started with a uniform \( T_c \) for the entire Tarim plate. For a given \( T_c \) value, the optimal tectonic load on each of the boundary domain can be derived by linear regression, and the F-test is conducted to score the linear correlation between observed basement deflection and those caused by the tectonic and sedimentary loads. The process is repeated systematically with different \( T_c \) values. The best \( T_c \) value is the one with which the highest confidence level of the F-test is reached.

[28] The results for the three Cenozoic periods are shown in Table 1. For the Paleogene the best fit is reached when \( T_c = 60 \) km. The corresponding confidence level of the F-test (\( \alpha \)) is above 97%, and the correlation coefficient (\( R \)) is 0.80, so the linearity assumption is acceptable. In other words, the Paleogene basement deformation can be satisfactorily explained by the summation of flexural response to the sedimentary loads and tectonic loads on the margins of a uniform elastic plate. The optimal tectonic load on each of the boundary domain for this period is shown in Table 1 and discussed later. Similar quality of the linear regression is obtained for the early-middle Miocene period. However, the assumption of linearity is unacceptable for the Late Miocene-Pliocene period: the best confidence level from the F-test is only 35% with the optimal \( T_c = 156 \) km, and unreasonably high tectonic loads (up to 1110 MPa or a 41 km high mountain on some boundary domains) are

<table>
<thead>
<tr>
<th>Boundary Domain</th>
<th>Paleogene</th>
<th>Early-Middle Miocene</th>
<th>Late Miocene-Pliocene</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>MPa km$^a$</td>
<td>MPa km$^a$</td>
<td>Mpa km$^a$</td>
</tr>
<tr>
<td>KL1</td>
<td>26 0.96</td>
<td>46 1.7</td>
<td>−57 −2.1</td>
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<tr>
<td>KL2</td>
<td>43 1.6</td>
<td>101 3.7</td>
<td>286 10.6</td>
</tr>
<tr>
<td>KL3</td>
<td>21 0.78</td>
<td>−27 −1.0</td>
<td>62 2.3</td>
</tr>
<tr>
<td>KL4</td>
<td>31 1.1</td>
<td>59 2.2</td>
<td>46 1.7</td>
</tr>
<tr>
<td>AT1</td>
<td>4 0.15</td>
<td>5 0.18</td>
<td>86 3.2</td>
</tr>
<tr>
<td>AT2</td>
<td>10 0.37</td>
<td>2 0.07</td>
<td>−40 −1.5</td>
</tr>
<tr>
<td>AT3</td>
<td>18 0.67</td>
<td>36 1.3</td>
<td>66 2.4</td>
</tr>
<tr>
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</tr>
<tr>
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<td>−101 −3.7</td>
<td>−152 −5.6</td>
</tr>
<tr>
<td>TS3</td>
<td>8 0.3</td>
<td>37 1.3</td>
<td>61 2.3</td>
</tr>
<tr>
<td>TS4</td>
<td>32 1.2</td>
<td>40 1.5</td>
<td>−54 −2.0</td>
</tr>
<tr>
<td>TS5</td>
<td>5 0.18</td>
<td>18 0.67</td>
<td>94 3.5</td>
</tr>
<tr>
<td>TS6</td>
<td>−93 −3.4</td>
<td>−38 −1.4</td>
<td>−180 −6.7</td>
</tr>
<tr>
<td>TS7</td>
<td>8 0.3</td>
<td>5 0.18</td>
<td>49 1.8</td>
</tr>
</tbody>
</table>

$^a$Tectonic loads converted to the weight of a rock column with a density of 2700 kg/m$^3$.

$^b$\( T_c \) is 24 km near Aksu and Minfeng.
These results indicate that the basement deformation during this period cannot be satisfactorily explained by flexure of a uniform elastic plate. Numerous factors not included in these calculations, such as internal faulting and the basinward growth of the thrusting belts, may contribute to the poor fit in this case. We modified the model to (1) include five major faults in the areas where the fit is particularly poor (see Figure 8), (2) widen the area of tectonic load in the Kepingtage region (boundary domains TS1–TS3) where geological evidence indicating a significant basinward growth of the thrust belt [Lu et al., 1998], and (3) lower the $T_c$ value (to 24 km) near the Aksu (Awati) and Minfeng depression where the Tarim plate may have been weakened by intensive late Cenozoic faulting [Jia, 1997]. With all these factors included in the model, the confidence level of linearity for late Miocene-Pliocene improved to 85%, and the misfit is significantly reduced. Inclusion of these factors for earlier periods, however, resulted in poorer fit, suggesting that these features probably developed after late Miocene.

Figure 9 shows the misfit between the predicted flexural deformation and the observed basement deflection for the Paleogene. Note that $\sim70\%$ of the basement deformation can be explained by sedimentary loading, and the largest misfits are near the margins of the basin (Figure 9a). The fits are systematically improved by applying the optimal tectonic loads on the margin domains (Figure 9b). The relative roles of sedimentary and tectonic loads are further illustrated in Figure 10. Figures 10a–10c shows the response to sedimentary loads alone for the three Cenozoic periods. Although sedimentary loads can explain much of the observed basement deformation, there are systematic misfits for deep depressions that are concentrated near the margins of the basin (Figures 2 and 4). The predicted
subsidence by sedimentary loads is always lower than the observed values for deep depressions, indicating the need for additional tectonic loads. The misfit is greatly reduced when the optimal tectonic loads (Table 1) are applied (Figures 10d–10f).

The temporal and spatial evolution of the optimal tectonic loads is plotted in Figure 11. During the Paleogene the tectonic loads on the Tarim basin were relatively minor. The maximum load is 43 MPa, roughly the weight of a ~1.6-km mountain range, in western Kunlun Shan and 32 MPa in the central segment of the Tian Shan thrust belt. On the boundary domains TS2 and TS6 the predicted tectonic loads are negative (upward force or unloading). Tectonic loads accelerated since the early Miocene, especially in the western segment of western Kunlun Shan thrust belt and the central segment of the Tian Shan thrust belt (note the different vertical scales in Figure 11). During the late Miocene, tectonic loads along the Tian Shan thrust belt split from the central Tian Shan. In the Kepingtage region tectonic loads broadened significantly, consistent with geological data [Lu et al., 1998]. Tectonic loads along the western Kunlun Shan thrust belt continued to accelerate, while moderate unloading (negative tectonic loads) is predicted for the central section of the Altyn Tagh fault system. These results may also be compared with the general topography of these mountain ranges. For the period of late Miocene-Pliocene, the average tectonic load over western Tian Shan thrust belt (boundary domains TS1–TS4) is larger than that over the eastern segments (TS5–TS7), consistent to higher elevations to the western part of south Tian Shan. The predicted large tectonic load over the western Kunlun Shan and relative small tectonic loads over the Altyn Tagh Shan are also comparable to the relative topography of these mountain ranges. However, caution is needed when Figure 11 is compared with topography because (1) the results in Figure 11 are the incremental tectonic loads for each period, (2) the division of the boundary domains is somewhat arbitrary, and (3) other factors may influence the tectonic loads. Some of these factors are discussed below.

5. Discussion

5.1. Effects of Model Parameters

One of the most important parameters for elastic flexure is the effective elastic thickness. Many factors that affect lithospheric rheology (e.g., geothermal condition, lithospheric composition, and mechanical weakening due to densely developed basement faulting), can be implicitly represented by $T_e$ [Burov and Diament, 1995]. In our models the optimal $T_e$ values were determined by minimizing the misfit between the observed and predicted basement deformation in 2-D models and by producing the highest linearity in 3-D models. Our 2-D models predicted the optimal $T_e$ varying from 60 to 15 km for different profiles, similar to the 2-D flexural results of Burov and Diament [1992]. However, the $T_e$ values obtained from the 3-D models are generally greater than 2-D values: they range from 60 km in the Paleogene to 108 km in the Late Miocene-Pliocene for most parts of the Tarim basin. The large discrepancies of the $T_e$ values among different profiles indicate the limitations of 2-D models for the Tarim basin. The high $T_e$ values from the 3-D models are similar to that of the India plate (90 km) [Jin et al., 1996] and consistent with the notion of the Tarim plate behaving as a secondary indenter during the Indo-Asian collision [Molnar and Tappin, 1975]. A stiff Tarim plate is also consistent with the large contrast of deformation styles between the Tarim basin and the surrounding mountain belts, and is supported by seismic data that indicate abnormally high velocity structures extending >220 km depth beneath the Tarim plate.
Figure 11. The predicted incremental tectonic loads (see Table 1) for the periods of (a) Paleogene, (b) early-middle Miocene, and (c) late Miocene-Pliocene. Note the different scales for tectonic loads in these plots.
Late Tertiary to different Tertiary strata (uplift as evidenced by the angular unconformity between Miocene this area experienced multiple phases of block)

The Bachu Uplift is bounded by a system of faults. Since torily fit by an intact elastic plate. One example is the Bachu

tion (Figure 8), the fit over this region is significantly improved. The poor fit to the late Miocene-Pliocene bas-
to the observed basement deformation. This may suggest that internal faulting of the basement had not caused

fit to the model (Figure 8), the fit over this region is significantly improved. The poor fit to the late Miocene-Pliocene bas-
ment deformation using an intact elastic plate indicates stronger fault activity since late Miocene.

We have also tested the effects of different model rheology. Over the long history of Cenozoic deformation of the Tarim basin viscous flow and stress relaxation may be expected, which cannot be simulated in elastic models. We thus designed a linear (Maxwell) viscoelastic flexure model with four layers of different material simulating the mechanical properties of the upper and lower crust, the strong layer in the uppermost mantle and the underlying mantle. The boundary conditions around the Tarim plate are free for vertical displacement but restricted for horizontal displacement. Lacking detailed age control on the sedimentary strata, we assumed the sedimentary and tectonic loads were applied linearly during each of the three Cenozoic periods, and a full range of rheological parameters are explored to seek for the best fit between the predicted and observed basement deformation. In general the results are worse than those of the elastic models. The results are generally better when the effective viscosity of the crust and the uppermost mantle is higher. These results are consistent with the notion of the Tarim plate being a relatively cold and stiff block through the Cenozoic [Lyon-Caen, 1986; Liu and Jin, 1993], and suggest that elastic flexure is a reasonably good first-order approximation of the Cenozoic deformation of the Tarim basin.

5.2. Implications for Mountain Building in the Tibetan Plateau and the Tian Shan

We have shown that the major part of the Cenozoic deformation of the Tarim basin can be explained by flexure under the sedimentary loads. Because these sediments were mainly derived from the surrounding orogenic belts [Hen- drix et al., 1992; Metivier and Gaudemer, 1997], the basin deformation can be directly related to the collisional mountain building in the surrounding regions. The additional tectonic loads may provide further insights into the geo-
dynamic coupling between the Tarim basin and the sur-
rounding Tibetan Plateau and the Tian Shan.

5.2.1. Uplift of the Western Kunlun Shan and the Pamir Plateau

The timing of tectonic uplift of the western Kunlun Shan is important for understanding the fundamental mechanics of the growth of the Tibetan Plateau [England and Houseman, 1988; Yin and Harrison, 2000]. 40Ar/39Ar fission track dating [Arnaud et al., 1993; Matte et al., 1996; Sobel and Dumitrut, 1997] suggest multiple events of cooling of the western Kunlun Shan in the Cenozoic, but the initial age of the western Kunlun Shan uplift in response to the Indo-Asian collision remains unclear. The ~1800-m Paleogene sedimentary rocks in the southwestern depres-
sion (Figure 2a) and the average 30 MPa tectonic load along the southwestern margin of the Tarim basin (Table 1) suggest that mountain building in western Kunlun Shan started at least in the Oligocene, consistent with recent stratigraphic studies in the southwestern depression that place the initiation age of the western Kunlun thrust belt in ~30 Ma [Yin and Harrison, 2000]. Since the Miocene mountain building in the western Kunlun Shan may have accelerated. The thickness of the sedimentary rocks in the southwestern depression reached 2800 m during the early-middle Miocene and more than 5000 m since the late Miocene, and the incremental tectonic loads over the entire western Kunlun belt, averaged over margin domains KL1–KL4, are 44 MPa and 84 MPa respectively for these two periods. Through the Cenozoic the maximum tectonic load has been on the boundary domain KL2 (Figure 8), in front of the depression center near Yecheng. The accelerating tectonic loads are consistent with the change in depositional facies from distal alluvial plains to proximal alluvial fans near Yecheng in the western Kunlun Mountains during the past 4.5 Myr [Zheng et al., 2000].

The accelerated tectonic load in this region has important implications for tectonic evolution and climate change. The increasing sedimentation rate and the associated change of facies of sediments in this region may indicate accelerated tectonic uplift or climate change, and a number of recent studies have interpreted these evidence as indicators of climate change in the past few million years [Molnar and England, 1990; Zhang et al., 2001]. In this study the sedimentary loads have been considered explicitly.
The increasing tectonic load derived here is independent of sedimentation rates and thus provides an unambiguous evidence for accelerated tectonic uplift of the western Kunlun Shan since late Miocene.

[39] Whether the Tarim plate subducted beneath the western Kunlun Shan has been an issue of debate. The south dipping western Kunlun Shan thrust belt is believed to have developed in response to the Indo-Asian collision [Lyon-Caen and Molnar, 1984]. Recent seismic survey shows no clear evidence of a subducting Tarim plate near 80.5°E [Kao et al., 2001]. This seems consistent with the predicted relatively small tectonic load (46 MPa) since late Miocene on boundary domain KL4, where the seismic profile was located. However, the predicted tectonic load increases westward, jumps to 286 MPa on domain KL2 (near Yecheng). This would be equivalent to have a 10-km-high rock column on the domain. Because the division of the boundary domains in the model is arbitrary, this result may be interpreted as requiring a lower tectonic load spread over a broader region. The high tectonic load may have resulted partly from overestimation of the basement subsidence near Yecheng, where the present land surface is about 500 m higher than the average elevation of the basin. By assuming a completely compensating basin and determining the basement depth from the isopachs, we may have overestimated the basement deformation near Yecheng by ~10%. Nonetheless, the high tectonic load over the western segments of the boundary (KL2 and KL3) since late Miocene seems real and would be difficult to explain by overthrusting of the western Kunlun thrust belt alone. It seems necessary to have at least part of the Tarim plate extended southward to beneath the western Kunlun Shan, especially near the western end of the Tarim basin where the predicted tectonic load is the high through the Cenozoic.

[40] The negative tectonic load on the boundary domain KL1 since late Miocene (Table 1) seems inconsistent with the indentation and uplift of the Pamir plateau since early Oligocene, which has caused up to 180 km crustal shortening in central western Kunlun Shan [Rumelhart et al., 1999]. One possible explanation is that the Karakaxi fault (Figure 1b) may have deepened in the past 10 Myr or so, weakening mechanical coupling between the Tarim basin and the Pamir plateau in this region. This would cause shift of tectonic loads to the two sides and hence help to explain the large tectonic loads on the adjacent boundary domains KL2 and TS1. The origin of the negative tectonic load on the boundary domain KL3 during early-middle Miocene is not clear.

5.2.2. Uplift of the Tian Shan Mountains

[41] The timing of the rejuvenation of the Tian Shan orogen by the Indo-Asian collision is critical for understanding the propagation of collisional stresses [Tapponnier and Molnar, 1979; Burchfiel et al., 1991; Lu et al., 1994]. Estimates based on fission tracks and stratigraphic analysis place the minimum age around ~25 Ma [Hendrix et al., 1994; Yin et al., 1998]. Our results do not show significant delay of mountain building in the Tian Shan relative to the western Kunlun Shan. The >900-m Paleogene sediments in the Kuche depocenter (Figure 2a) and ~30 MPa tectonic load from the central Tian Shan thrust belt (domain TS4 in Table 1) indicate that rejuvenation of the Tian Shan started before the end of Oligocene. Since the Miocene the tectonic loads along the Tian Shan-Tarim boundary accelerated and migrated westward, probably related to the indentation of the Pamir plateau. The major phase of uplift in the Tian Shan probably occurred after late Miocene, indicated by the >5000-m sedimentary rocks in the Kuche and Aksu (Awati) depressions and the tectonic loads in this period (Figure 11c and Table 1). There are also significant variations within the Tian Shan belt, such as tectonic unloading in domain TS2 since late Miocene that may be correlated to the Bachu block uplift [Jia, 1997]. The predicted basinward broadening of the Kashi-Aksu thrust belt (Figure 11c) is consistent with geological studies [Lu et al., 1994, 1998]. These internal variations of the optimal tectonic loads may reflect the tectonic segmentation along the Tian Shan and the Tarim boundary [Yin et al., 1998]. The accelerated mountain building in the Tian Shan since late Miocene is also consistent with the deformation rates extrapolated from geodetic measurements [Abdrakhmatov et al., 1996].

[42] Whereas positive tectonic loads may be related to overthrusting of mountain block onto the margins of the basin, the meaning and origins of negative tectonic loads are not always clear. For the boundary domains TS2 and TS6, negative tectonic loads are predicted for the entire Cenozoic. Although some of these results may be model-dependent artifacts, an inevitable consequence of arbitrarily dividing the boundary domains and making a first-order approximation, we cannot dismiss all the negative tectonic loads as artifacts, because many of the predictions are consistent with geological features in these regions. As mentioned above, the negative tectonic load on TS2 is correlated with the Bachu Uplift, an uplifted-region since late Paleozoic. However, the cause of the uplift is unclear. The Upper Cambrian to Permian strata was exposed in TS2, the Kepingtage area [Jia, 1997], suggesting strong erosion that could explain the negative tectonic loading. The negative load on domain TS2 may also be related to an abnormally hot mantle beneath the central Tian Shan where mantle upwelling is inferred from seismic evidence [Makeyeva et al., 1992]. The boundary domain TS6 is geographically correlated with the Yanji Basin (Figure 1). However, the origin of the Yanji basin is not clear.

5.2.3. Tectonics of the AltyTagh Fault System

[43] The AltyTagh fault has been the focus of intensive studies because of its hypothesized role in facilitating large-scale eastward extrusion of Asian continent during the Indo-Asian collision [Tapponnier and Molnar, 1977]. Closely related to the current controversy of the displacement history of the AltyTagh fault is the nature of the fault activity and the timing of its initiation, which ranges from the Oligocene to Pliocene in various studies [Bally et al., 1986; Peltzer and Tapponnier, 1988; Wang, 1990; Burchfiel et al., 1991]. Some workers regard the AltnTagh fault as a fundamental lithologic boundary [Tapponnier et al., 1982; Deng, 1989; Arnaud et al., 1993; Wittlinger et al., 1998]; others consider it to be a crustal structure [Burchfiel et al., 1989]. It remains controversial as whether or not the Tarim
Miocene, which may have caused dynamic decoupling of the isopach maps (Figure 2). The tectonic unloading may be northwestward after late Miocene, as one may infer from the Paleogene to middle Miocene, but was flipped to part of the basin boundary, which was southeastward from change of the dipping direction of the basement along this central segment of the Altyn Tagh fault is reflected in the native (unloading) during late Miocene. This unloading in the AT2, Table 1) decreased since Miocene and become negative for late Miocene-Pliocene. Nonetheless, some useful insights of the Cenozoic sedimentation and tectonic loads along the southern margin of the Tarim basin. The model predicts the optimal thickness of the effective elastic plate is 60 km in the Paleogene, 84 km in the early-middle Miocene, and >100 km in most part of the Tarim plate since Miocene.Plate subduction under the western Kunlun Shan and Altyn Tagh Shan started in the Paleogene and accelerated since Miocene, whereas the Altyn Tagh fault is characterized by tectonic unloading since mid-Miocene.

[44] Because horizontal loading and shearing are not the dominant causes of the flexure of the lithosphere and are not included in the model, the results cannot provide much constraint on the timing and magnitude of shearing along the Altyn Tagh fault. Nonetheless, some useful insights of tectonic activity along the fault can be derived from the Cenozoic sedimentation and tectonic loads along the southern margin of the Tarim basin. The model predicts increasing tectonic loads in the boundary domains AT1 near Minfeng and AT3 near Ruoqiang (Table 1 and Figures 8 and 1) throughout the Cenozoic, probably related to continuous mountain building in the northern part of the Tarim basin. The model predicts the tectonic uplift along the central segment of the Altyn Tagh fault (domain AT2, Table 1) decreased since Miocene and become negative for late Miocene. This unloading in the central segment of the Altyn Tagh fault is reflected in the change of the dipping direction of the basement along this part of the basin boundary, which was southeastward from the Paleogene to middle Miocene, but was flipped to northwestward after late Miocene, as one may infer from the isopach maps (Figure 2). The tectonic unloading may be explained by deep cutting of the Altyn Tagh fault since Miocene, which may have caused dynamic decoupling between the Tarim basin and northern Tibetan Plateau in this region. A young deep-cutting Altyn Tagh fault is consistent with a low-velocity anomaly extending to 140-km depth based on teleseismic imaging across the central Altyn Tagh fault [Wittlinger et al., 1998], and Quaternary basaltic eruptions along the trace of western Altyn Tagh fault [Deng, 1989].

[45] The evolution of tectonic loads in the western Kunlun Shan, the central part of the Tian Shan (TS3 ~ TS5) and the central segment of the Altyn Tagh fault are compared in Figure 12. Uplift in both western Kunlun Shan and Tian Shan may have started in Paleogene and accelerated since Miocene, where the Altyn Tagh fault is characterized by tectonic unloading since late Miocene.

6. Conclusions

[46] The deformation history of the Tarim basin during the Cenozoic provides useful insights into mountain building in the Tibetan Plateau and the Tian Shan mountains in response to the Indo-Asian collision. Major conclusions we may draw from this study include the following:

1. The Cenozoic history of basement deformation in the Tarim basin can be largely (up to 70%) accounted for by foreland flexure under the sedimentary loads. Near major compression centers additional tectonic loads are needed, suggesting that these depressions were associated with tectonic uplift and erosion of the surrounding mountain ranges. The fit between the observed and the predicted basement deformation using a uniform elastic plate was unsatisfactory for late Miocene-Pliocene and required involvement of additional factors, such as internal faulting. The rhomb-shape geometry of the Altyn Tagh basin and large internal heterogeneity make 3-D models particularly useful.

2. The Tarim plate is indeed cold and rigid throughout the Cenozoic. The optimal thickness of the effective elastic plate is 60 km in the Paleogene, 84 km in the early-middle Miocene, and >100 km in most part of the Tarim plate since late Miocene. Abnormally low Tc (~24 km) near the Aksu (Awati) and Minfeng depression since late Miocene is probably caused by intensive and localized faulting. The rigid Tarim plate is consistent with seismic tomography indicating a high velocity anomaly extending to >220 km depth under the Tarim basin. The increasing Tc throughout the Cenozoic is consistent with the cooling history of the Tarim basin.

3. Tectonic uplift of the western Kunlun Shan mountain ranges occurred during Oligocene or earlier. Rapid uplift started around early-middle Miocene and has been accelerating ever since. The large tectonic loads required for this region cannot be entirely explained by the loads of the western Kunlun Shan thrust belt, and suggest that part of the Tarim plate has subducted underneath the western Kunlun Shan mountains. The strongest underthrusting of the Tarim plate may have occurred under the western end of the western Kunlun thrust belt.

4. Tectonic rejuvenation of the Tian Shan mountains in response to the Indo-Asian collision is not significantly later than mountain building in the western Kunlun Shan. More
tectonic loads were predicted in the western part of the southern Tian Shan thrust belts, probably related to the indentation of the Pamir plateau. The large along-strike variation of tectonic loads since early Miocene is consistent with geological evidence of fault-controlled segmentation of the southern Tian Shan thrust belt. The basinward broadening of the Kepingtage thrust belt initiated in Late Miocene.

5. Throughout the Cenozoic tectonic loads near Minfeng and Ruqiang increased moderately, indicating continued uplift in the eastern end of the western Kunlun Shan and the Altyn Tagh Shan. However, along the central segment of the Altyn Tagh fault tectonic loads started to decrease since Miocene and become negative (unloading) since late Miocene. These results may be explained by dynamic decoupling between the Tarim basin and the northern Tibetan Plateau, probably due to the Altyn Tagh fault cutting the entire lithosphere and the southward migration of the active fault zone. The same results do not favor subduction of the Tarim plate under the Altyn Tagh fault.

[47] Most studies of collisional tectonics in central Asia have been focused on the Tibetan Plateau and other mountain belts in this region. Our results suggest that much about the collisional orogenesis can be learned from the Tarim basin and probably the numerous other intermountain basins in central Asia. At present one major hurdle is the lack of age constraints on the Cenozoic sedimentary strata in the Tarim basin and the surrounding regions. However, promising progress has been made in recent years by combining the lithostratigraphy, biostratigraphy, magnetostratigraphy, and 39Ar-40Ar thermochronology [Yin et al., 1998; Yin and Harrison, 2000]. Future studies, with improved age constraints and 3-D models that explicitly include the fault systems bounding the Tarim basin, should be able to significantly refine the picture of temporal-spatial development of crustal deformation in central Asia in response to the Indo-Asian collision and help to address many remaining questions, such as the control of strain partition between crustal thickening and tectonic extrusion following the collision.

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