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Gravity anomalies over the Ferghana Valley (central Asia) and intracontinental deformation

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Abstract. Gravity anomalies over the Ferghana Valley, a deep sedimentary basin in the western Tien Shan, differ from those expected for local Airy isostatic equilibrium by 120 mGal and imply a deficit of mass beneath the basin equivalent to a Moho deeper by 10 km than prevails where isostasy holds. Although the Ferghana Valley is underlain by thick Mesozoic and Cenozoic sediment, measurements of density show that this material cannot account for the deficit of mass. Gravity anomalies over both the surrounding mountain belts and the Kazakh Platform farther north imply that local Airy isostatic equilibrium is approached; flexure of a relatively thin, effectively elastic, plate (thickness ~ 10-15 km) loaded by the present topography can account for the range of Bouguer anomalies of -50 to -450 mGal over these regions. We infer that approximately north-south shortening of the relatively thin lithosphere has created mountains north and south of the basin, has warped the basement of the immediate surroundings of the basin up by folding the mantle lithosphere, and has forced the basin floor down beneath the Ferghana Valley. It appears that the preexisting thermal structure and variations in crustal thickness have dictated the styles and distribution of deformation in this region.

1. Introduction

Although both the highest and most active mountain range of the world, the Himalaya, and the best studied, the Alps, occur where one plate of continental lithosphere has underthrust the margin of another, a comparably important class of mountain belts develops in intracontinental settings. In Asia, intracontinental mountain building accounts for a large fraction of India's penetration into Eurasia. Convergence occurs across the Tien Shan, the highest range outside the Himalaya-Karakorum chain, at a rate (~20 mm/yr [*Abdrakhmatov et al.*, 1996]) that is comparable to that deduced for the Himalaya (20-23 mm/yr [e.g., *Bilham et al.*, 1997; *E. Lavé and J.-P. Avouac*, Abandoned fluvial terraces across the Siwalik hills (Nepal): Part I, Fluvial response to climatic and tectonic forcing, submitted to *Journal of Geophysical Research*, 1998]). Similarly, much of the other high terrain north of the Himalaya seems to have been created by intracontinental mountain building, such as in the Nan Shan [e.g., *Meyer*, 1991], Altay of Mongolia [e.g., *Burov et al.*, 1993], and Gobi-Altay [e.g., *Tapponnier and Molnar*, 1979]. Moreover, the Rocky Mountains of the western United States formed in an intracontinental setting, much like that of the eastern cordillera of the Andes [e.g., *Jordan et al.*, 1983; *Molnar and Lyon-Caen*, 1988], and the eroded ranges surrounding the late Paleozoic Amadeus Basin in central Australia [e.g., *Shaw*, 1991; *Shaw et al.*, 1991a, b]. A logical question posed by presence of these intracontinental regions is what processes dictate where such intracontinental mountain building occurs.

Two classes of answer present themselves naturally. In one, processes occurring beneath the regions, such as some form of small-scale convection, exert forces on the base of the lithosphere that cause, or at least contribute to, the convergence. In the other, forces applied from afar are responsible for the convergence, and weaknesses in the lithosphere dictate where deformation occurs. Clearly, one class of explanation might work in one area, whereas the other works in another. Yet, given our difficulties in determining the deep structure, let alone the processes occurring beneath some regions, the task for the moment seems to be to test one or the other class of mechanisms in a few regions and let such tests eliminate the other.

If intracontinental mountain building occurs because regions are stressed from afar, and weak areas deform, then the next question must be, What aspect of a region makes it weak? Again two possibilities present themselves. Hot areas are likely to be weaker, but chemical, or more precisely mineralogical, differences can also cause heterogeneity in strength. The great strength of oceanic lithosphere seems to derive from its mineralogy being dominated by olivine, which has been measured to be stronger than the dominant crustal minerals in the continents (calcite, quartz, feldspars, micas, and maybe pyroxene) [e.g., *Brace and Kohlstedt*, 1980], and from oceanic crust being so thin as to play no role in the strength profile, compared with that of thicker and presumably

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weaker continental crust [*Burov and Diament*, 1996]. Thus we might expect a cold region to be stronger and more plate-like than a hot region, and, similarly, a region of thin crust to be stronger than one of thicker crust, for a similar thermal profile.

With the goal to test the idea that lateral variations in strength can dictate where intracontinental deformation occurs, we studied the westernmost Tien Shan, which includes the Ferghana Valley (Tadjik and Uzbek "Farghona," hereinafter also referred to as Ferghana Basin, since both names, Ferghana Valley and Ferghana Basin, are equally used in the Russian/Soviet geological literature) and the Chatkal Ranges to its north and the Alai (South Tien Shan) Range to its south (Figure 1).

2. Tectonic Setting and Geological Evolution of the Ferghana Valley, Chatkal Ranges, and South Tien Shan

The Tien Shan terminates west of the right-lateral Talas-Ferghana fault by splaying into two narrow mountain chains that surround the Ferghana Valley. The Central Tien Shan farther east consists of a more uniform structure that includes an alternation of ranges and basins separated by reverse faults [e.g., *Makarov*, 1977; *Sadybakasov*, 1990; *Yudakhin*, 1983]. The dominantly east-northeast topographic trend of ranges and basins terminates at this fault. The northwest end of the Talas-Ferghana fault also bounds the northeast end of the Chatkal Ranges, a zone of northeast trending narrow ranges and basins similar to those east of the fault, but oriented somewhat differently [e.g., *Sadybakasov*, 1990].

Slip on the Talas-Ferghana fault of 180 km since Paleozoic time [*Burtman*, 1963, 1964, 1975] appears to include only a relatively small fraction (~60 km) in Cenozoic time [e.g., *Burtman et al.*, 1996]. Yet because this fault separates the Ferghana Valley from a deforming region to its east, neither the amount nor the current rate of slip on the fault can be constant along it. For our purposes here, it suffices to note that this fault seems to separate two areas with very different styles of deformation, different topography, and perhaps different crustal and upper mantle structures [e.g., *Kosarev et al.*, 1993; *Roecker et al.*, 1993]. A narrow belt of mountains parallel to the central section of the fault slopes downward to the Ferghana Valley, underlain by a deep basin with as much as 8 km of Cretaceous-Cenozoic sediment [*Cobbold et al.*, 1993, 1996; *Yudakhin*, 1986; *Yudakhin et al.*, 1991].

The geologic structure of the Chatkal Ranges requires north-northwest-south-southeast crustal shortening, by reverse faulting and basement folding [e.g., *Sadybakasov*, 1990]. Apparently relatively narrow, east-northeast trending valleys mark loci of reverse faults and tightly closed synclines. Inferences of crustal thickness from seismic refraction lines and gravity anomalies imply that the crust thickens from roughly 40 km north of the Chatkal Ranges to as much as 60 km beneath them [e.g., *Ulomov*, 1974]. Yet no major thrust fault seems to bound the ranges on the north side. Thus if one major thrust fault bounds the Chatkal Ranges, it must lie along their southern margin, a possibility that gravity anomalies, discussed below, seem to rule out. Thus crustal shortening seems to have occurred by distributed deformation across the ranges, as it does within the Tien Shan farther east.

Paleomagnetic declinations of Cretaceous [*Bazhenov*, 1993] and Cenozoic [*Thomas et al.*, 1993] sedimentary rock demonstrate approximately 20°-30° of counter-clockwise rotation of the Ferghana Valley with respect to Eurasia. This rotation seems to have occurred about an axis at the southwest end of the Chatkal Ranges. Thus it calls for convergence between the Ferghana Valley and Eurasia that increases from southwest to northeast and could exceed 100 km at the northeast end of the Ranges.

South of the Ferghana Valley, the east-west trending South Tien Shan separates the apparently rigid block beneath the Ferghana Valley from the Pamir to the south. Fault plane solutions of earthquakes show active underthrusting of the Ferghana Valley crust beneath the South Tien Shan [*Nelson et al.*, 1987]. The rate of underthrusting is difficult to estimate, but intense Quaternary deformation [e.g., *Nikonov*, 1977; *Nikonov et al.*, 1983; *Trifonov*, 1983], plus geodetic measurements farther west [e.g., *Guseva*, 1986], imply that most of the convergence between the Pamir and the Ferghana Valley occurs within the Alai Valley, which separates the Pamir and South Tien Shan [e.g., *Burtman and Molnar*, 1993].

Although deformation clearly occurs within the Chatkal Ranges and just south of the South Tien Shan, the Ferghana Basin can be considered as a composite ramp basin. Overthrusting on two principal reverse faults of opposite vergence, on the northern and southern sides of the Ferghana Valley, may have depressed the basin. What is most puzzling about the basin, as revealed by gravity data, is its great depth, much deeper than would be expected simply from the loading by ranges on its north and south sides. Thus the basin might be the result of flexure of the footwalls of the reverse faults or perhaps of compressional buckling instability associated with the roughly north-south compression of the region. The valley surface is less than 500 m high, with the surrounding Chatkal, Ferghana, and Alai ranges each reaching 4000 - 5000 m height. On its western side, according to seismic and borehole data, the basin is filled with more than 8 km of, mainly Neogene, sediment [*Cobbold et al.*, 1993, 1996; *Yudakhin*, 1986].

3. Gravity Anomalies

A complete Bouguer gravity map was constructed at a scale of 1:2,000,000 (Figure 2) using ~ 15′ by 15′ average values from data made available through the Center of Geophysical Computer Data Studies of the Joint Institute of Physics of the Earth Russian Academy of Sciences and from the gravity maps used by *Burov and Kogan* [1990], *Burov et al.* [1990], *Burov et al.* [1993], *Beekman* [1994], *Artemjev et al.* [1994]. Topography is from the GETECH (Geophysical Technology Ltd.) Global DTM5 (5′ × 5′) database. Separate one-dimensional gravity profiles (Figure 3) and the isostatic residuals (Figure 2c), prepared in 1991 from 5′ × 7.5′ data, have 10 km interval, which provides sufficient resolution for the purposes of this study. Because original gravity measurements were made with a closer spacing (about 1-2 points per 1 km), there should be no artifacts due to interpolation. High-quality data sets (gridded at spacings of 5′ × 5′, 1′ × 1′, and smaller) for the former Soviet Union (FSU) do also exist at International Scientific Environmental Center of Russian Academy of Sciences, the Russian Military Geodetic Service, or GETECH.

We analyze gravity anomalies in terms of deviations from Airy isostasy using flexed and buckled (or folded) elastic and inelastic plates. Such plates may be loaded both by vertical

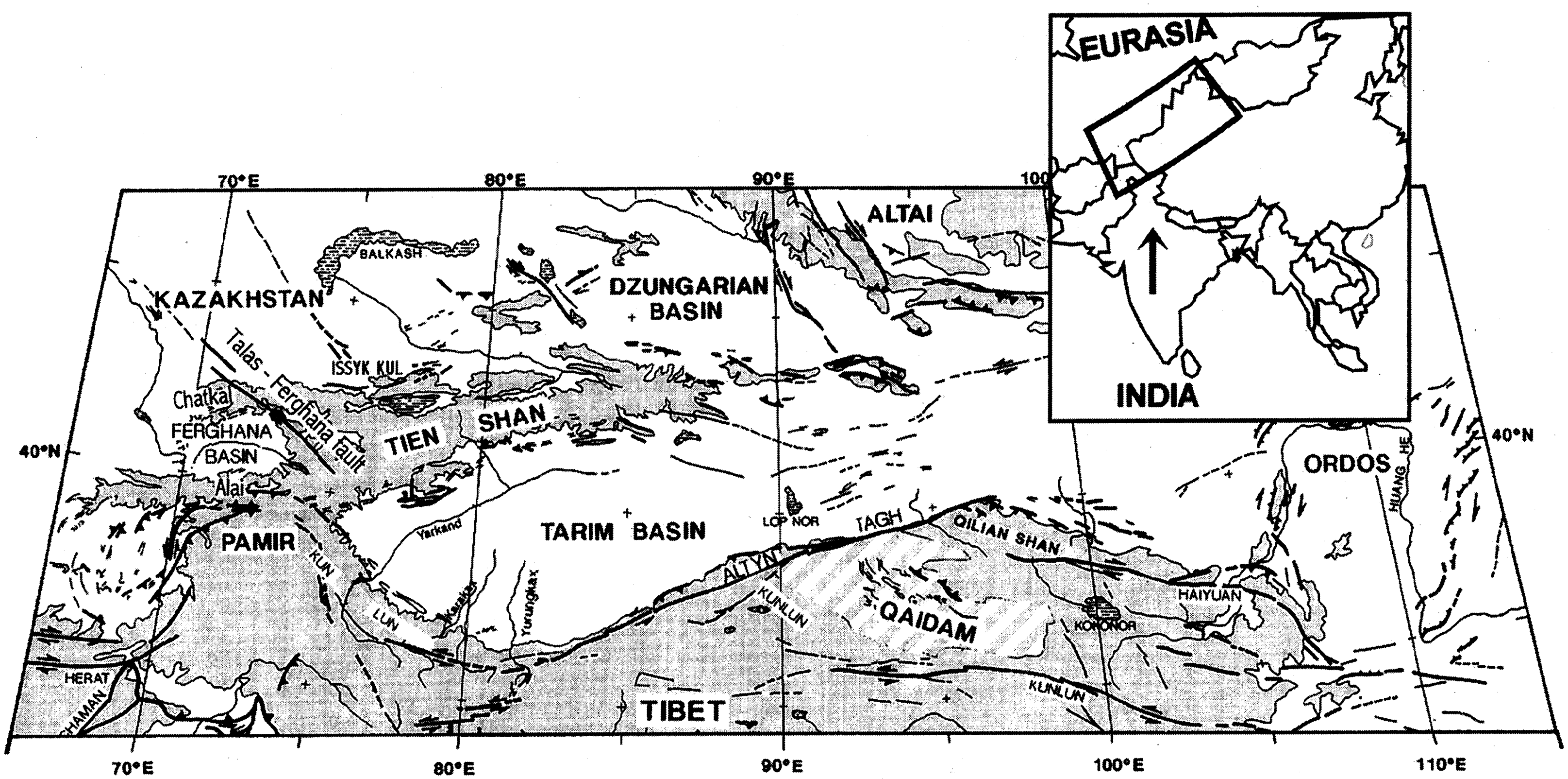

Figure 1. Map of region showing topography, with an insert showing where in Asia the area studied lies (modified from *Avouac and Tapponnier* [1992] and *Avouac and Burov* [1996]. Grey areas correspond to mountainous regions (> 2000 m).

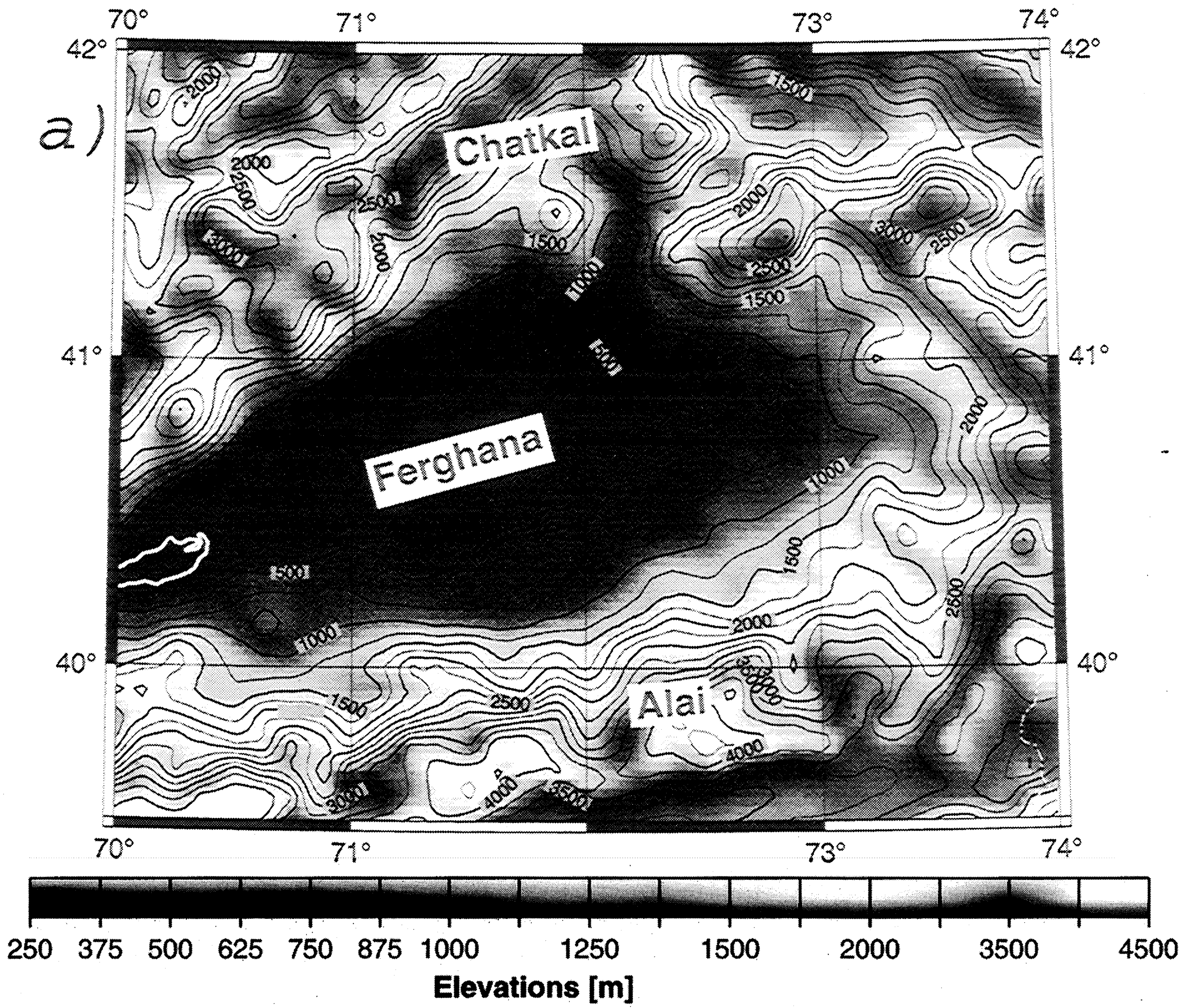

Figure 2. Maps from (a) 5′ × 5′ DTM5 topography catalogue, (b) complete Bouguer gravity anomalies (smoothed 15′ × 15′) and (c) residual (isostatic) gravity anomalies over the Ferghana Valley and the surrounding mountains, with location of the data profiles. The residual (isostatic) gravity anomaly is equal to the difference between Bouguer anomalies computed from measured gravity (Figure 2b) and those calculated from the mean elevations in (Figure 2a) assuming local Airy isostasy (using Talwani's method, uniform density of the crust 2670 kg/m^3, density of the mantle lithosphere 3330 kg/m^3, normal crustal thickness 45 km). Note that a 120-mGal isostatic anomaly overlies the valley, whereas the surrounding area seems to be very close to local isostatic equilibrium.

and by horizontal forces per unit horizontal length of plate. For pure flexure due to vertical forces, we use the finite difference method of *Burov and Diament* [1992, 1995] to calculate the finite deformation of the plate, using either simple elastic plates and more "realistic" structures with brittle, elastic, and ductile layers. For the latter, we integrate plate strength over depth and scale it to agree with that of an elastic plate of a given thickness (10 km in most calculations) [*Burov and Diament*, 1995, 1996; *Burov et al.*, 1993]. For flexure with folding and buckling, we used the finite element code TECTON [e.g., *Melosh*, 1990].

3.1. The Density Structure and Sedimentation History of the Ferghana Valley

Sedimentation in the Ferghana Valley and surrounding regions record a history of submergence and emergence of the valley beneath a shallow sea. *Sinitsyn* [1960, p. 101] reported Triassic sediment only along the southern side of the valley. He thought that the modest amount of sediment reflected an absence of relief necessary to provide a source of sediment [*Sinitsyn*, 1960, p. 102], but such relief seems to have become widespread in Jurassic time. Jurassic sediment varies in thickness and type throughout the region, with basal sections characterized by conglomerate and with a fining upward [*Kuzichkina*, 1972]. *Sinitsyn* [1960, p. 103] reported widely deposited Cretaceous marine sediment; although nowhere do depths appear to have been great. This sediment thins to the north and south of the Ferghana Valley. A rim around the basin had formed by this time [*Kreydenkov and Raspopin*, 1972, p. 249; *Sinitsyn*, 1960, p. 109], with an easily traced shoreline (*A. Bakyrov and V. P. Sankova, personal communication*, 1997). Marine sedimentation continued into Paleogene time, but thicknesses of Paleogene marine sediment are commonly only 100-200 m [*Sinitsyn*, 1960, p. 105; *Sinitsyn*, 1962].

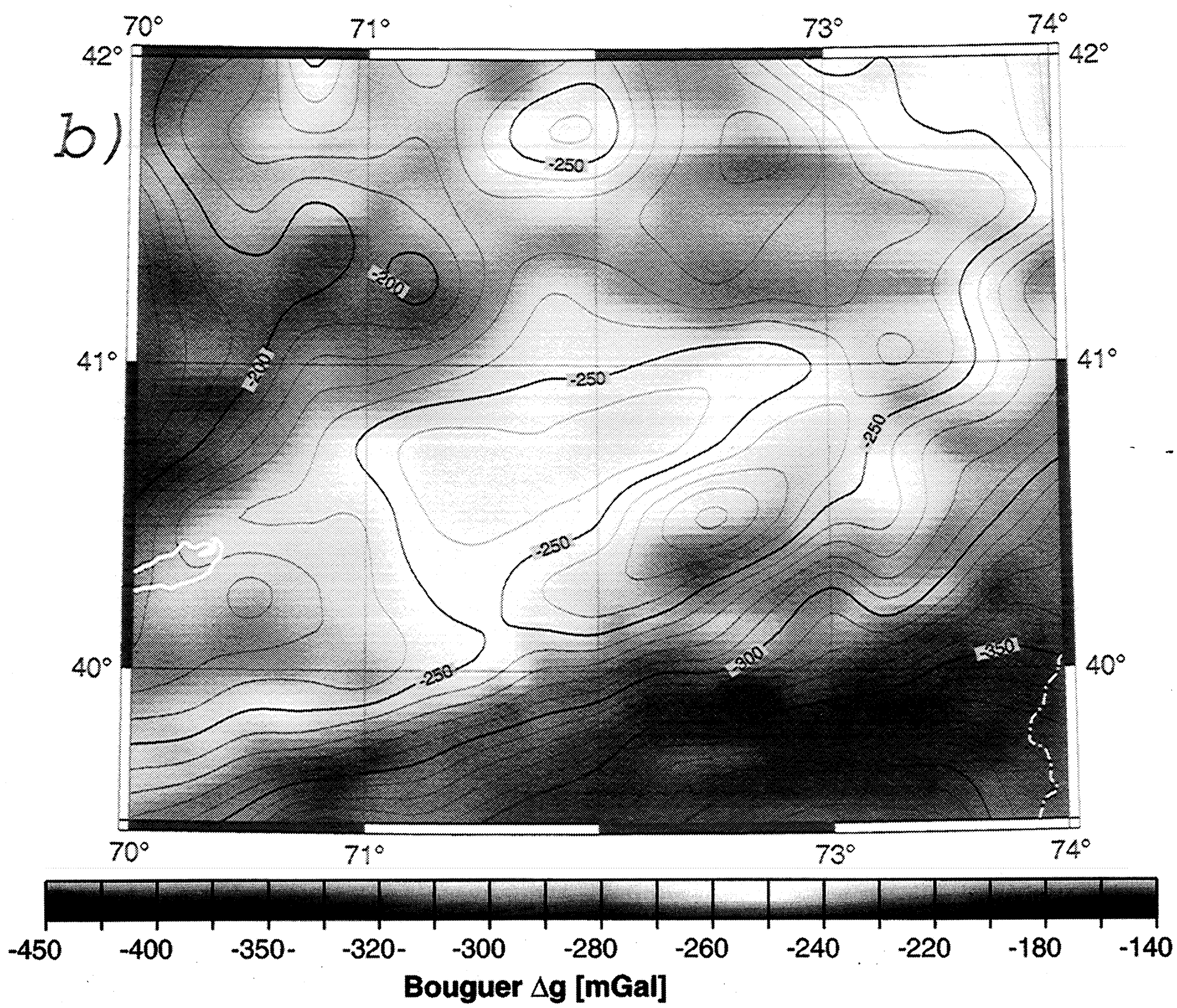

Figure 2. (continued)

Kreydenkov and Raspopin [1972] reported that the shorelines of the Ferghana Valley in Cretaceous time remained approximately the same as in Paleocene time, until the Eocene retreat of the sea. Thus the general form of the Ferghana Valley, with a low region surrounded by higher terrain, was built nearly 100 m.y. before India collided with the rest of Eurasia. The bulk of the basin fill beneath the Ferghana Valley, however, was deposited in Neogene time, when several thousands of kilometers of terrigenous sediment eroded from the neighboring hills accumulated in the basin.

Accounting for the near-surface density structure is crucial for interpreting the gravity data. The density of Mesozoic and Cenozoic sedimentary rock in the basins of the Tien Shan area varies over a large range from 1500-2000 kg/m^3 to 2600-2750 kg/m^3 [*Yudakhin*, 1986], depending on lithology, porosity, depth, age, and location with respect to the active orogenic environment. These density measurements are based on numerous borehole studies made at different times in the Ferghana, Tadjik, and Issyk-Kul basins by various exploration and research institutions of the FSU. For all basins the density of the sediment rapidly increases with the depth from the surface, with an average gradient of about 0.07-0.1 kg/m^{-4} that drops to 0.03-0.04 kg/m^{-4} at the depth of 3000 m. In general, the density reaches 2500 kg/m^3 at the depth of ~3000 m (the maximum depth of boreholes in the area). In the Ferghana Basin the average density of Neogene sedimentary rock is 2180 kg/m^3 at a depth of 1000 m, 2440 kg/m^3 at 2000 m, and 2510 kg/m^3 at 3000 m [*Yudakhin*, 1986]. *Yudakhin* [1986] approximated the density versus depth for the Ferghana Basin with a second-order polynomial (Table 1), which reveals some variability across the basin but also shows that densities approach 2500 kg/m^3 rapidly with depths exceeding 2-3 km. As we show below, this prohibits the sediment in the basin from accounting for much of the large negative Bouguer anomaly over the basin.

3.2. General Features of the Gravity Field

The Ferghana Valley is bordered on the northwest by the Chatkal (Figure 1) and Kurama mountains, on the northeast by the Ferghana Range (the last two small ranges are not shown in Figure 1), and on the south by the Alai (Figure 1) and Turkistan ranges (also not shown), which rise over 5000 m. As is common to regions of high terrain, Bouguer gravity over the Chatkal Ranges, the Ferghana Range, the South Tien Shan, and the Pamir are negative; those over the Pamir with mean altitudes in excess of 4000 m are especially negative (Figures 2 and 3). The negative anomalies over the Chatkal

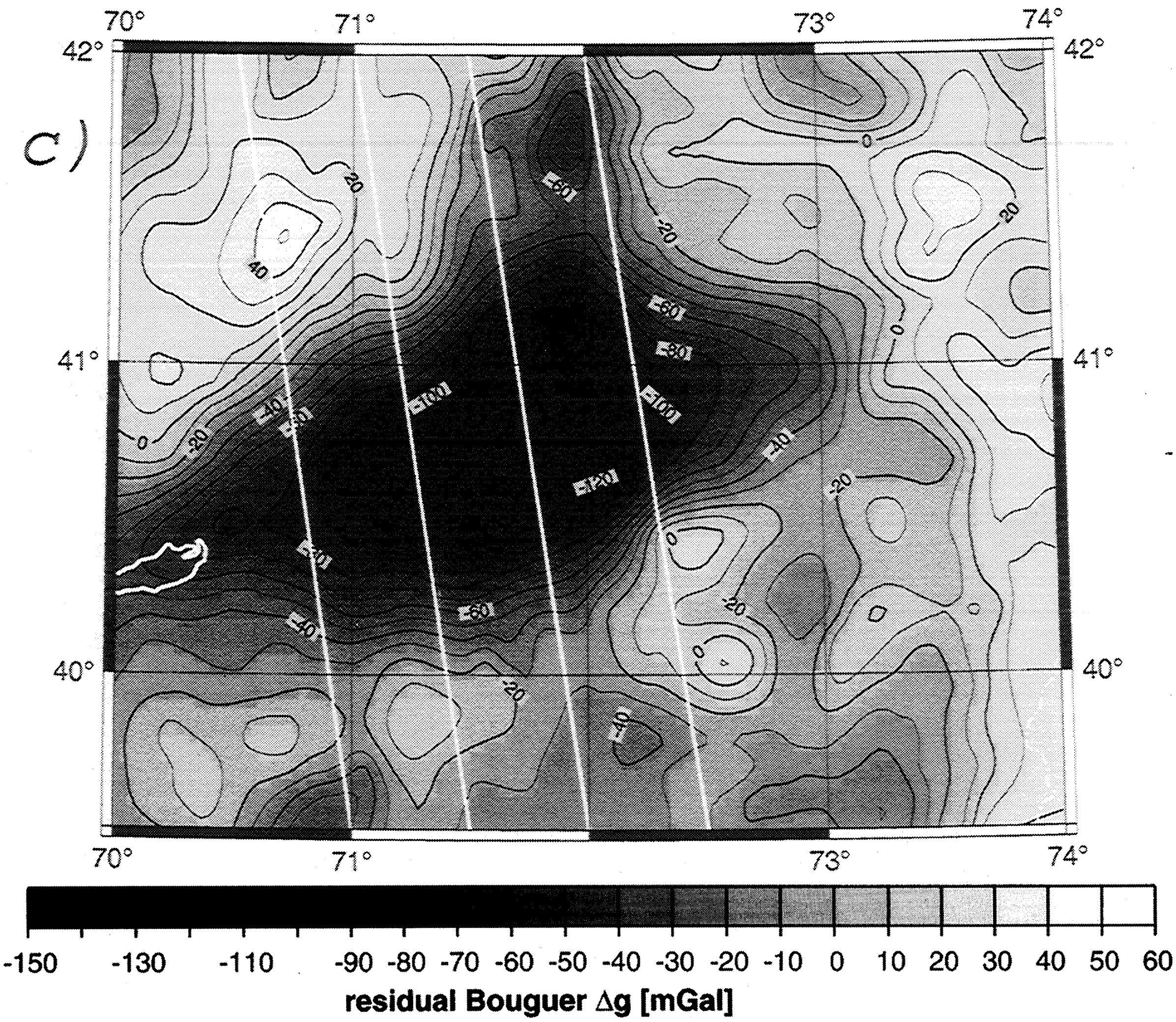

Figure 2. (continued)

Ranges and the South Tien Shan mimic the topography and require a state of nearly local Airy isostatic equilibrium (assuming a uniform density of 2670 kg/m^3 for all mass above sea level). A continuous, effectively elastic plate as thick as 50 km cannot underlie the region, for it fails to reproduce the ~70-80 mGal minimum over the Chatkal Ranges (Figure 3). Even a thin plate, with an equivalent elastic thickness of only 10 km, fits slightly less well than local Airy isostasy (no elastic plate at all). This agrees with previous estimates of the equivalent elastic thickness of the Kazakh Platform of no more than 15 km [*Burov et al.*, 1990]. If effectively elastic plates underlie the basins adjacent to the ranges, they are masked by short-wavelength features associated with the laterally varying structure. The sparse data available to us prohibit resolving variations in structure beneath the Pamir, but clearly a thick plate seems unacceptable for it too.

Although a thin plate seems surprising for a stable region, other observations concur with it. First, evidence from different sources suggests that the Tien Shan region at least underwent Jurassic rejuvenation [*Burtman*, 1975; *Hendrix et al.*, 1992], which might have resulted in significant re-heating of the lithosphere. Leith [1985] suggested such re-heating from the subsidence history of the Tadjik Depression, and a similar distribution of Jurassic to Paleogene sedimentation in the Ferghana Valley suggests that it might have undergone a similar Mesozoic and early Cenozoic evolution. From heat flow measurements, *Schwartzman* [1991] inferred a shallow depth for the 400°C isotherm not only in the Ferghana Basin (< 15 km) but also almost everywhere beneath the Kazakh shield (10-20 km). This contrasts with a depth of 30-40 km expected for crust whose thermal structure was last reset in Paleozoic time [e.g., *Burov and Diament*, 1995] and supports the inference that the effective elastic thickness of the Kazakh lithosphere is small. Correspondingly, a depth of 15 km for the 400°C isotherm corresponds to effective thermal age less than 150 m.y. (Jurassic). Moreover, calculations of equivalent elastic thicknesses of lithosphere with a thermal structure appropriate for 150-Ma-old lithosphere with 40-50 km of overlying crust should be around 10-15 km [e.g., *Burov and Diament*, 1995]. Finally, *Djanuzaikov et al.* [1977] reported that the average thickness of the seismogenic layer, associated with the brittle upper crust, is 15 km in most of the Tien Shan, reaches 30 km at the junction of the Chatkal, Ferghana, and Talas Ranges, but decreases to 8-10 km (base of the sedimentary cover) in the Ferghana Valley. Thus several independent arguments concur with a relatively thin effective elastic thickness for the Tien Shan region.

Although local Airy isostasy accounts for the Bouguer gravity anomalies over much of the region, those from the Ferghana Basin reveal significant negative deviations with re-

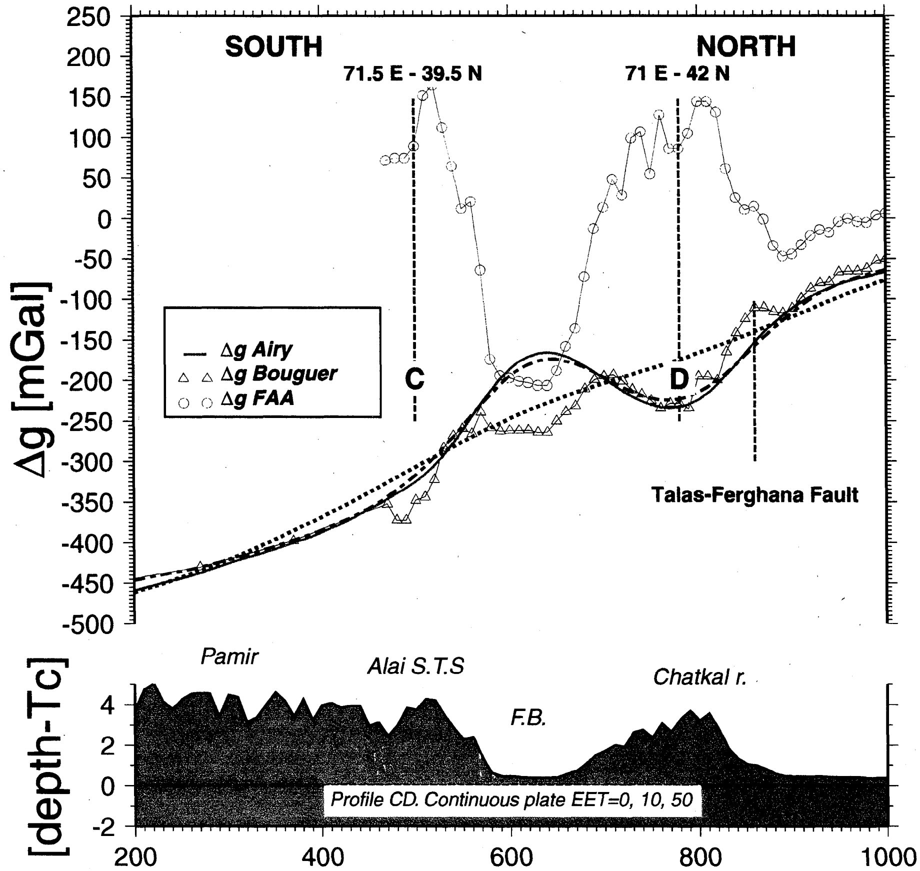

Figure 3. A characteristic profile of (bottom) topography and plate deflections and (top) gravity anomalies across the Southern Tien Shan (S.T.S.), Ferghana Basin (F.B.) and the Alai Range. For gravity anomalies: circles, free-air anomalies (FAA); triangles, Bouguer gravity anomalies; solid line, calculated Bouguer anomalies assuming local Airy isostasy; dot-dashed line, calculated Bouguer anomalies for a continuous elastic plate with an equivalent elastic thicknesses of 10 km; dashed line, the same for such a plate 50 km thick. Short-wavelength gravity lows to the south of the Alai Range and highs to the north of the Chatkal Range correspond to the narrow Alai Valley and Talas-Ferghana fault, respectively. The finite difference method of *Burov and Diament* [1995] was used to calculate the finite deformation of the plate and the resulting gravity anomalies. The Moho depression (bottom) is shown in terms of plate deflections from undeformed crustal thickness (T_c) of 45 km.

spect to local Airy isostasy, suggesting overcompensation of the surface topography (Figure 2). Although the basin occupies the lowest area, Bouguer anomalies over it are yet more negative than those over the ranges immediately to the north and south. A large regional trend in Bouguer anomalies (Figure 3), with values becoming increasing negative to the south over the high plateau of the Pamir, obscures this minimum somewhat, but from both the map (Figure 2) and the profiles (Figure 3), the difference between Bouguer anomalies and those for an Airy model can be seen to reach 50 - 100 mGal over the central parts of the basin. Such a minimum requires there be a mass deficit beneath the basin. In fact, a clearer demonstration of the mass deficit is provided by free-air anomalies of -200 mGal over the basin. The mass deficit beneath the Ferghana Valley might be associated with either low-density sediment within the basin or a deeper Moho than isostatic equilibrium requires.

The Bouguer anomaly minimum over the basin is also sharply bounded by relatively short-wavelength highs, especially along the southeastern and northwestern sides of the basin, which must be caused by relatively shallow sources (Figures 2 and 3). With a width of the gravity low over the basin of less than 100 km, and surrounding highs only 50 km wide, the gravity field cannot be accounted for by any structure deeper than the Moho.

3.3 Role of Sediment in the Ferghana Valley

The correlation of a gravity low over a basin with thick sediment obviously suggests that the low density of the sediment should account for the low. With this possibility in mind, we calculated Bouguer anomalies for a state of Airy

Table 1. Polynomial Approximations [*Yudakhin*, 1986] for the Depth-Density Dependence in the Ferghana Basin

Region of the Ferghana Basin	Density, kg/m^3, as a Function of Depth z, km
Southwest	$1000 \times (2.320 + 0.07z + 0.04z^2)$
Southeast	$1000 \times (2.340 + 0.065z + 0.005\ z^2)$
Northern	$1000 \times (2.489 + 0.0026z + 0.0006z^2)$
Central	$1000 \times (2.120 + 0.325z - 0.073\ z^2)$ for $z \leq 1$ km
Central	$1000 \times (1.940 + 0.46z^{1/4})$ for $z \geq 1$ km

isostatic equilibrium but with sediment assigned both a low density and a large thickness. Figure 4 shows gravity anomalies calculated for Airy isostasy, but assuming a basin with 0, 2, and 8 km of sediment with an average density of 2300 kg/m³ (and therefore 350 kg/m³ less dense than the surrounding crustal rock). Only the limiting assumption of 8 km of sediment, the maximum thickness of the sedimentary cover in the Ferghana basin, assigned a very low density can account for the observed anomaly. Yet *Yudakhin's* [1986] profiles of density (Table 1) prohibit such a low value at a depth below ~1.5-2 km. His measured profiles of density versus depth imply compaction of the sediment so that its density exceeds 2400 kg/m³ at a depth of 2000 m. As shown in Figure 4, calculated gravity anomalies assuming Airy isostasy and only a thin layer with such low density fit the observed Bouguer anomalies over the Ferghana basin poorly. A reasonable, if simplified, density profile for the Ferghana basin would include about 2 km of sediment with an average density of 2300 kg/m³, below which the density can be assumed to be equal to the average crustal density of 2650 kg/m³.

A relatively good fit between calculated and observed Bouguer anomalies for 8 km of low-density sediment is achieved because of a trade-off of two aspects. For Airy isostasy, 8 km of low density sediment implies a thinner crust than normal, in this case by ~10 km (compared with a Moho at ~35 km for normal crust, Figure 4). Although the structure beneath the Ferghana Valley is difficult to study because of seismic noise over the thick sediment, some seismological data suggest that the Moho is at least 50 km deep beneath the Ferghana Basin [*Sabitova*, 1986, 1991] and perhaps at 56-62 km. According to the same data, the Moho reaches a depth of 60 km beneath the Chatkal Ranges [see also *Ulomov*, 1974]. Similarly, receiver functions of teleseismically recorded *P* waves at stations in and surrounding the Ferghana Valley suggest an average crustal thickness of 50 km [*Kosarev et al.*, 1993]. The smallest crustal thickness values beneath the Ferghana Valley are proposed by *Zunnunov* [1985, pp. 60-64] who reports a depth to the Moho of 46-48 km from *PmP* reflections but with a deepening to the northern and south. Thus seismological data, as well as the combination of relatively high-density sediment in the basin [*Yudakhin*, 1986] and the mass deficit implied by the Bouguer anomalies, suggest that crust cannot be unusually thin beneath the basin.

We conclude that the relatively large variations in Bouguer anomalies over short distances require, first, that the effective elastic thickness be small. The relatively narrow Chatkal Ranges do not appear to be supported by strength in the lithosphere, but seem to be compensated locally. Yet the most marked deviation from local isostatic equilibrium, that is found over the Ferghana Valley and its surroundings, cannot be explained by a low average density of sediment in the valley; measured densities rule that out. Thus the deficit of mass in this area must include a crustal thickness that is greater than that expected from its low altitude.

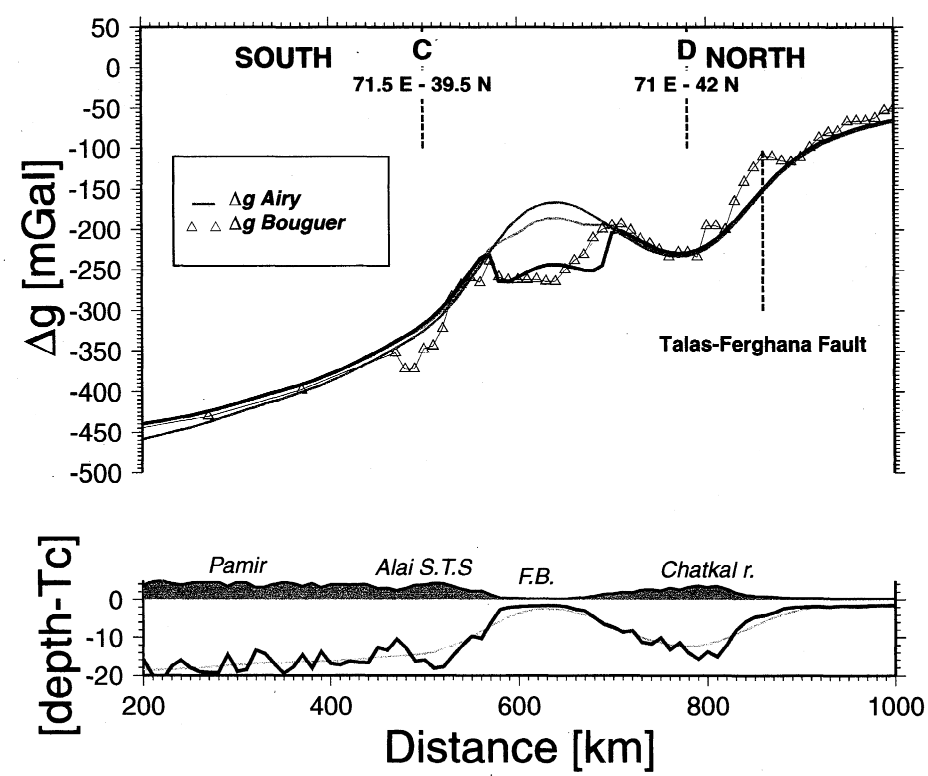

Figure 4. (top) Profile of observed and calculated Bouguer gravity anomalies assuming a continuous elastic plate 10 km thick or local Airy isostasy, but for different thicknesses of sediment in the Ferghana Basin (modeled as a rectangular prism 100 km wide) with a density of 2300 kg/m³ (i.e., 350 kg/m³ less dense than the surrounding crust). Solid gray line, calculations for thickness of 0 km (no sediment); light gray line, 2 km; dark solid line, 8 km. (bottom) Light gray and dark lines show corresponding variations in estimated crustal thickness along the profile (three-dimensional plate deflections). The finite difference method of *Burov and Diament* [1995] was used to calculate the finite deformation of the plate and the resulting gravity anomalies.

4. Mechanical Processes for Creating the Ferghana Valley

The question posed by the data is thus, What mechanisms could account for localized deformation beneath the neighboring ranges and for the deepening of the valley between them (Figure 5a)? Various mechanisms may be responsible for the uncompensated depression of the Moho beneath the Ferghana Valley: lithospheric subduction of the basin beneath the adjacent ranges, folding, or buckling, of the lithosphere, or perhaps mantle dynamics at depth. In particular, *Burg et al.* [1994] suggested that the Ferghana Basin could result from downward flexure due to a folding or buckling instability of the lithosphere under compression. A key observation supporting this idea may be the short-wavelength gravity high surrounding the basin (Figures 2 and 3) which can be interpreted as a result of localized uplift of edges of the basin due to the development of a compressional instability. Such uplift is not implied by the other models suggested above.

4.1. Lithospheric Subduction

A logical possibility to account for the mass deficit beneath the Ferghana Valley is that overthrusting of the adjacent ranges has loaded the crust of the Ferghana Valley and depressed it. In this case, the depression of crust of normal thickness, plus the burial by 8 km of sediment, would cause a negative Bouguer (and free-air) gravity anomaly. As noted above, such a southward underthrusting of the basin beneath the South Tien Shan is implied by fault plane solutions of earthquakes [e.g., *Nelson et al.*, 1987], and a similar northward underthrusting beneath the Chatkal Ranges seems likely from the geologic structure [e.g., *Sadybakasov*, 1990]. Such a double ramp structure has been inferred for negative gravity anomalies over the Adriatic Sea [*Moretti and Royden*, 1988].

As attractive as this simple structure is, it fails to account the gravity high that surrounds the basin. As is clear from studies elsewhere, the underthrusting of lithosphere beneath a mountain chain requires that there be a negative anomaly over the basin, with the minimum at the edge of the chain and even over the lowest part of the chain, rather than over the basin. This is especially clear for the Himalaya [*Lyon-Caen and Molnar*, 1983, 1985]. The flexure of the lithosphere requires a thickening of crust beneath the chain, but the underthrusting of sediment beneath the chain can add additional low-density material beneath the chain. In the case of the Himalaya, a positive isostatic anomaly overlies the chain, because the strength of the underthrust lithosphere supports part of the mass of the mountain chain, but that positive is approximately 100 km from the edge of the basin. The gravity anomalies associated with such a structure do not resemble those of Ferghana Basin and its surroundings. Instead of a gravity low, or minimum, in the Airy isostatic anomaly field over the edge of the ranges surrounding the Ferghana Valley, we observe a gravity high, implying a mass excess compared to the surroundings (Figures 2 and 3).

We tested subduction scenarios beneath and south of the Ferghana Basin by trying to reproduce the gravity anomaly over the basin with a discontinuous plate extending southward beneath the Alai and Pamir. Merely extending a plate a finite distance under these ranges fails to reproduce the gravity anomaly over the Ferghana Basin. This of course does not deny the existence of such subduction but does show that flexure associated with that subduction is not sufficient to explain the gravity anomaly over the basin.

4.2. Deep Mantle Dynamics

Studies of seismic wave velocities suggest that the deep structure beneath the Ferghana Valley is quite different from that beneath the Tien Shan east of the Talas-Ferghana fault. Whereas relatively low-speed material characterizes the uppermost mantle east of the fault, that west of it beneath the Ferghana Valley and the Chatkal Ranges seems to be normal, if not high [*Roecker et al.*, 1993]. This suggests that the upper mantle is hotter beneath the Tien Shan east of the Talas-Ferghana fault than west of it. The transition at the Moho also seems different east and west of the fault, with a sharper transition beneath the Ferghana Valley and Chatkal Ranges than beneath the Tien Shan farther east [*Kosarev et al.*, 1993]. *Makeyeva et al.* [1992] interpreted variations in seismic anisotropy as evidence of small-scale convection, with upwelling east of the fault and downwelling west of it. Thus downwelling beneath the Ferghana Basin might draw the crust down, creating a deep Moho, much as we inferred from gravity anomalies over the eastern Tien Shan [*Burov et al.*, 1990]. The deepened Moho would cause negative gravity anomalies, which would be only partially counterbalanced by the long-wavelength positive anomaly caused by the cold, dense downwelling material at greater depth.

We cannot eliminate the possibility that such a process contributes a long-wavelength anomaly, but again, such a process alone cannot account for the short-wavelength anomalies. In particular, the gravity highs adjacent to the basin cannot be explained by structure at a depth where mantle flow occurs. Moreover, such flow is not likely to wrinkle the Moho with a wavelength of only 50-100 km, as we observe over the margins of the Ferghana Valley. Thus, although mantle dynamics might contribute to the regional field, it cannot account for both the negative gravity anomalies over the basin and the adjacent gravity highs.

4.3. Folding, or Buckling, of the Lithosphere

The short wavelengths that characterize the gravity field over the entire region, including the Chatkal Ranges, the Ferghana Valley, and the transition from the valley to both the Chatkal Ranges and the South Tien Shan, require that the variations in density responsible for the field be quite shallow, within the crust or at the Moho. Of course, it is a simple matter to insert variations in density and calculate anomalies that fit those observed, but we prefer to address calculations based on likely processes. Because the largest density contrast is associated with the Moho, the smallest amount of deformation at this boundary than can account for the observed anomalies will deflect it. Consequently, we examined mechanisms that create both a relatively short-wavelength depression of the Moho beneath the basin and very short-wavelength uplifts on its sides. Folding of a viscous layer or buckling of an elastic layer [e.g., *Biot*, 1961; *Burov et al.*, 1993; *Martinod and Davy*, 1992, 1994] could produce such geometry of the Moho.

To calculate the characteristic dimensions of a folded viscous layer we used a modified version of the non-Newtonian, viscoelastic finite element code "TECTON" developed by Melosh and Raefsky [e.g., *Melosh*, 1990], which is particularly useful for calculating finite amplitude growth of folding

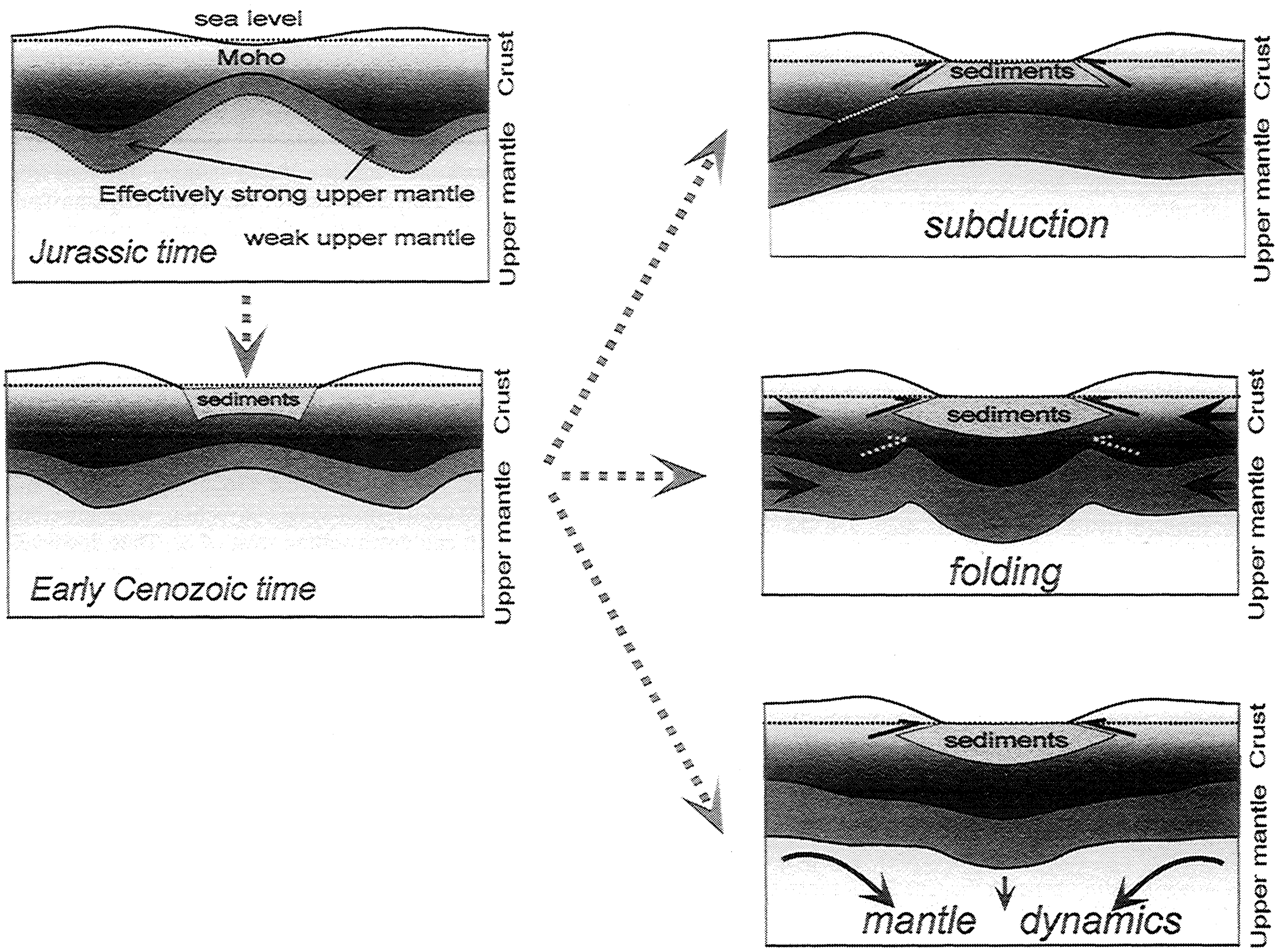

Figure 5a. Various plausible mechanical processes for creating the Ferghana Valley and the deep sedimentary basin beneath it: "classical" (Early Cenozoic time), subduction of lithosphere to the south of the basin, folding of the lithosphere, deep mantle dynamics.

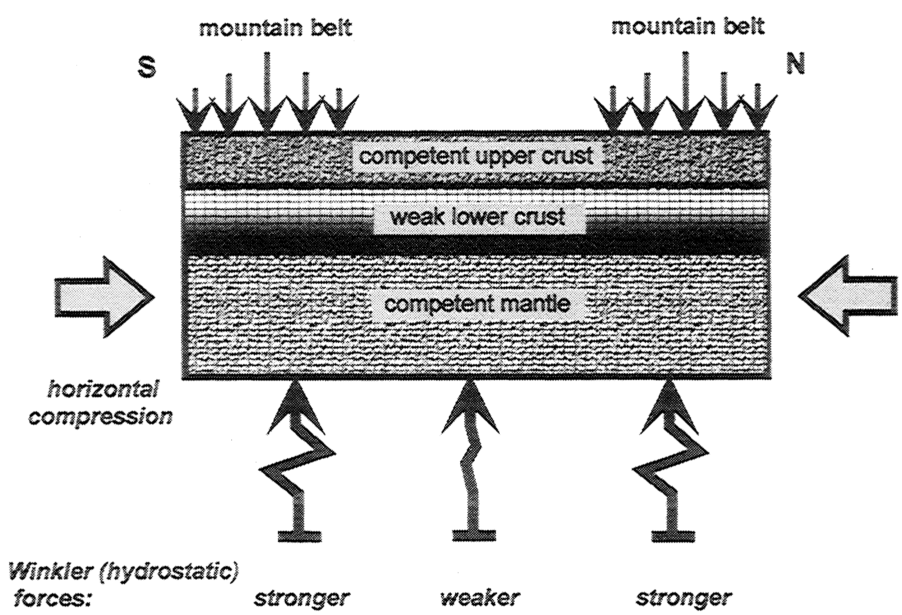

Figure 5b. Sketch of geometry and boundary conditions of layered elastoviscous plate model. Finite element model used here is based on standard four-node linear elements.

and buckling instabilities [*Beekman*, 1994]. The numerical finite elements used in this code are standard four-node linear elements equipped with incompatible quadratic shape functions. The size of the elements was between 2.5 × 5 km and 5 × 7.5 km. We used both elastic and viscous structures. However, in most calculations, we treated the competent parts of the plate as elastic, because this calculation is much simpler and faster than that for a more realistic layered viscous medium. As is well known, an equivalent elastic plate can reproduce the static deformation of such a layered structure. Thus, to expedite calculations, we assumed a thin (Figures 3 and 5) two-dimensional plate not only with a horizontal force per unit length but also with upward vertical forces imposed beneath the borders of the Ferghana Basin (commonly 1×10^{12} to 2×10^{12} N/m), acting downward at the center of the basin and upward on its northern and southern edges (Figure 5b). These vertical forces make no direct physical sense. Needed only in the elastic models, they model downward buckling of the basin and upward deflection of the bordering lithosphere on the south and north of the basin, as deduced from the calculations using Tecton to model the folding of the layered structure. As is well known [*Smith*, 1975], finite amplitude of deformation in an unstable compressive regime can be influenced by many, usually negligible, factors. Thus our primary goal was only to obtain a reasonable amplitude of deformation applying reasonable forces (i.e., not exceeding mechanical strength of the rocks). The wavelength of the deformation is a stable parameter and is unaffected by variations in the loads.

The analysis described above suggests an effective elastic thickness of ~10 km (Figure 3), and therefore we used it to calculate the shape of an elastic plate loaded horizontally. Such an effective thickness guarantees the wavelengths that we observe, but the magnitudes of deflections and corresponding gravity anomalies depend also on density contrasts and magnitudes of the horizontal force. Using a variety of horizontal loads and density contrasts, with the topography along the profile specified, we calculated gravity anomalies for comparison with those observed. As is common, no unique set of assumed parameters that yield a fit between measured and calculated gravity anomalies can be found. We show two examples here to illustrate the effects of changing densities.

In one case (Figure 6), we ignored the presence of light material in the Ferghana Valley, and in the other (Figure 7), we used an overestimate of it. In the latter case, we included a top layer of sediment in the basin with a relatively low density of 2300 kg/m^3. In both cases, the downward deflection of the Moho under the Chatkal Ranges and the South Tien Shan occurs because of the weight of the ranges, but that beneath the Ferghana Valley is due to the imposition of a horizontal force per unit length. Between the basin and the surrounding ranges, upward flexure elevates the Moho and provides the local mass excesses needed to create the gravity highs. For the case where low-density sediment in the basin is ignored, the Moho extends to the same depth as beneath the Chatkal Ranges, and therefore is roughly 10 km deeper than both the adjacent area where the Moho is not flexed down and north of the Chatkal Ranges. For undeformed crust 45 km thick, the calculated depth of the Moho reaches 56 km beneath the basin in the first case (Figure 6) and 50 km in the second (Figure 7), compared to 45 km for Airy isostasy. Thus, for the case with sediment occupying the basin, the Moho beneath the Ferghana Valley is only 4-5 km deeper than that adjacent to it and 6 km deeper than that north of the Chatkal Ranges.

The horizontal force per unit length F_h required to initiate buckling of an elastic plate of constant thickness T_e is $F_h = (4D\Delta\rho g)^{1/2}$ [*Timoshenko and Woinowsky-Krieger*, 1959], where $D = ET_e^3/12(1-\nu^2)$ is the flexural rigidity of the plate, E (= 80 GPa) is Young's modulus, ν (=0.25) is Poison's ratio and $\Delta\rho$ is the density difference between that of the surface load and mantle. For the viscous case, D can be written as $D=D_v= G_{eff}h^3/3 = 2\dot{\varepsilon}\,\mu_{eff}h^3/3n$, where $G_{eff} = 2\dot{\varepsilon}\,\mu_{eff}/n$ is the effective shear modulus, h is the thickness of the competent high-viscosity layer, usually equal to 1-1.5 T_e depending on the depth gradient of the viscosity, μ_{eff} is the averaged effective viscosity of this layer estimated at the background compressional strain rate $\dot{\varepsilon}$, and n is power law exponent (= 1 in the Newtonian case). For an elastic plate with T_e = 10 km, $F_h \sim 10^{13}$ N/m, a magnitude to which, as discussed below, we attach little significance, and the corresponding wavelength of buckling, λ_b ($= 2\pi(2D/F_h)^{1/2}$), is in the range of 200-250 km. The shape of the deflection of the Ferghana Basin fits half of such a wavelength (Figure 6b).

Sedimentary fill in the basin reduces the effect of gravity by reducing the difference in density between the overlying and underlying material. As a result, buckling of a layer overlain by sediment that accumulates as buckling occurs can be initiated with a lower horizontal force (2 to 5 × 10^{12} N/m), than a layer that buckles into air. Moreover, the horizontal force required to buckle an elastic plate is much higher than that which can be expected for folding of plates with a more realistic rheological structure [e.g., *Burov et al.*, 1993; *Zuber*, 1987]. Using TECTON, we confirmed that folding of a 10-km-thick ductile plate with effective average viscosity of 10^{23} Pa s to 10^{24} Pa s at characteristic strain rates of 10^{-15} s^{-1}, requires a horizontal force per unit length of the order of 10^{12} N/m. In any case, because of ignorance of the rheological structure, we assign no significance to these horizontal forces, but merely to the characteristic wavelengths and approximate range of amplitudes of deformation.

4.4 Folding and Faulting of the Lithosphere

A number of authors [e.g., *Burg et al.*, 1994; J.-P. Burg, personal communication, 1997] have suggested that lithospheric folding in Central Asia may result in formation of deep mantle thrust faults. Though there is no clear evidence for existence of such faults, it is believed that they can form (or be reactivated) as a consequence of stress localizations in the narrow upward inflexion zones on the sides of the basin.

As we have shown above, such zones of locally uplifted Moho can be predicted on the base of the gravity data and reproduced in the elastic or viscous mechanical model (Figures 3 and 5-7). However, the problem of cointeraction between folding and brittle faulting can be only assessed using more appropriate rheologies, for example, brittle-elastic-ductile rheology laws derived from rock mechanics data [e.g., *Kirby and Kronenberg*, 1987].

The parameters of these laws are quite uncertain, but in each particular case they can be constrained by various supplementary data such as estimates of the equivalent elastic thickness, strain rate, depths of seismicity, etc. [*Burov and Diament*, 1995]. In the case of the Ferghana Basin (40-50 km thick crust), the estimated value of the effective elastic thickness (EET, 10-15 km) is compatible with the assumption of a

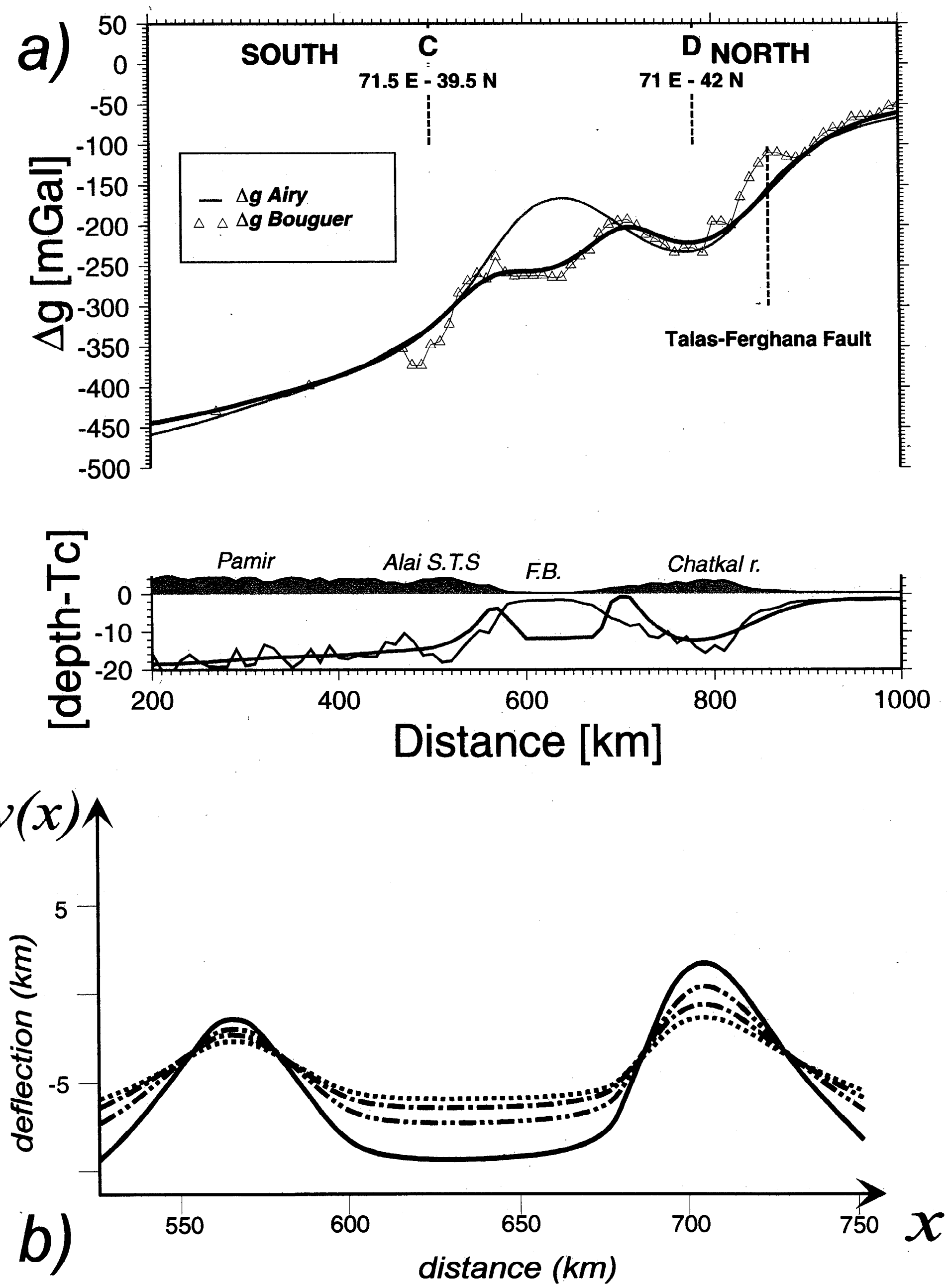

Figure 6. (a) (top) Measured and calculated Bouguer gravity anomalies for a competent layer loaded horizontally by a force F_h (per unit length) ($F_h = 2 \times 10^{13}$ N/m for an elastic plate or 10 times smaller for a viscous plate (μ=1 $\times$ 10^{23} Pa s) and with the topography appropriate for the region. For an elastic plate an additional downward vertical force (3×10^{12} N/m) is applied at the center of the basin, and two upward forces (1.5×10^{12} N/m) are applied at its southern and northern edges to fit the gravity. These additional forces are not needed for inelastic layers. This calculation ignores the contribution made by light material in the basin. (bottom) Moho profiles for Airy isostasy and for the buckled layers (thick dark line). The finite element code TECTON [e.g., Melosh, 1990] was used to calculate the deformation of the plate. (b) Progressive deflection of Moho during plate shortening (2, 4 , 8, and 10 Ma since start of compression). During compression the amplitude of plate deflection changes significantly whilst the wavelength of folding stays constant.

quartz-dominated crust and olivine dominated mantle with the temperature profile corresponding to the age of cooling of about 150 Ma [*Burov and Diament*, 1995]. The assumption of such thermal age for this basin agrees with the heat flow data and with the evidences for the thermal rejuvenation in the Jurassic time (see above) [*Burtman*, 1975; *Hendrix et al.*, 1992; *Schwartzman*, 1991].

To investigate the possible interplay between folding and faulting, we used the PARAVOZ code developed by Poliakov and Podladchikov [*Poliakov et al.*, 1996] from the FLAC algorithm originally proposed by *Cundal* [1989]. The essential advantage of this fully explicit time-marching scheme is that it allows us to model formation and development of brittle faults in various brittle-elastic-viscous materi-

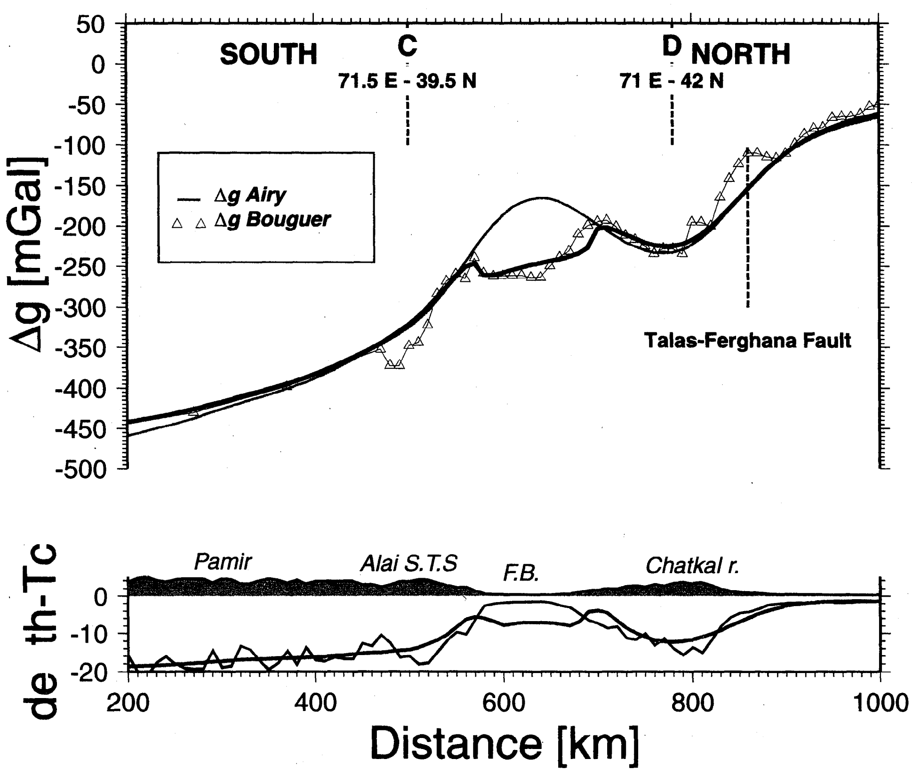

Figure 7. (top) Measured and calculated Bouguer gravity anomalies for a competent layer loaded horizontally by a force (F_h) per unit length of (F_h= 0.5 × 10^{13} N/m for an elastic plate or 10 times smaller for a viscous plate (μ= 1 × 10^{23} Pa s) and with the topography appropriate for the region. For an elastic plate an additional downward vertical force (2 × 10^{12} N/m) is applied at the center of the basin, and two upward forces (1.0 × 10^{12} N/m) are applied at its south and northern edges to fit the gravity. These forces are not needed for inelastic models. For this calculation we assumed that the Ferghana Valley is filled with 4 km of sediment with a density of 2300 kg/m^3. Thus this calculation overestimates the contribution made by light material in the basin. (bottom) Moho profiles for Airy isostasy and for a buckled plate (thick dark line). The finite element code TECTON [e.g., *Melosh*, 1990] was used to calculate the deformation of the plate.

als without needing to predefine them [*Poliakov et al.*, 1996]. The general setup of this experiment was essentially the same as for the previous experiments (no predefined faults, laterally homogeneous lithosphere, Figure 5a), but for the sake of simplicity we did not introduce any preexisting topography loads. The brittle-elastic-ductile rheology used in these experiments is based on the quartz/olivine power flow law (assuming a 150 Ma continental geotherm) and Byerlee's brittle failure law (see *Burov and Diament* [1995] for the inelastic parameters; the elastic parameters are the same as in the previous experiments). The primary results of our experiments (Figure 8) show the following: (1) Indeed, for the particular case of the Ferghana-like basin, crustal and mantle faults may develop as a result of folding. (2) Folding continues even after the development of these faults, so that folding and faulting can coexist together for a long time (many millions of years). (3) Presence of the low-viscosity lower crust may lead to biharmonic (decoupled) folding of the crust and mantle so that the shorter crustal wavelength (~50 - 70 km) will be superimposed on the longer (~200-250 km for the given case) mantle wavelength (quartz flow law for the Jurassic geotherm yields μ=10^{22} Pa s at the top of the crust and only 10^{19} Pa s at the Moho depth, whereas the olivine mantle maintains μ=10^{23} - 10^{24} Pa s). (4) In the case of crust-mantle decoupling, one can expect quite complicated distributions of the crustal faults and surface undulations, and the crustal faults may be not necessarily connected to the mantle faults.

It is unlikely that the mantle faults can be longer than a few kilometers in depth. It is possible as well that such faults may not develop at longer wavelengths of folding (stronger lithosphere).

5. Conclusions

Gravity anomalies from the Ferghana Valley deviate markedly, by as much as 120 mGal, from what they would be in local Airy isostatic equilibrium. Thus they imply a large deficit of mass beneath the basin and require that that mass deficit be supported either by strength in the lithosphere or by dynamic processes that deform the lithosphere. This aberration in the gravity field differs from the field over the neighboring mountain ranges north and south of the valley and over the Kazakh Platform farther north, where deviations from Airy isostasy are small and require a relatively thin effectively elastic plate (thickness ~ 10 to 15 km).

Although the thick sedimentary fill of the Ferghana Valley, with as much as 8 km of postmiddle Mesozoic sediment, might seem to offer a simple explanation for the mass deficit, measurements of density in the valley [*Yudakhin*, 1986] and seismologically determined Moho depths prohibit low-density

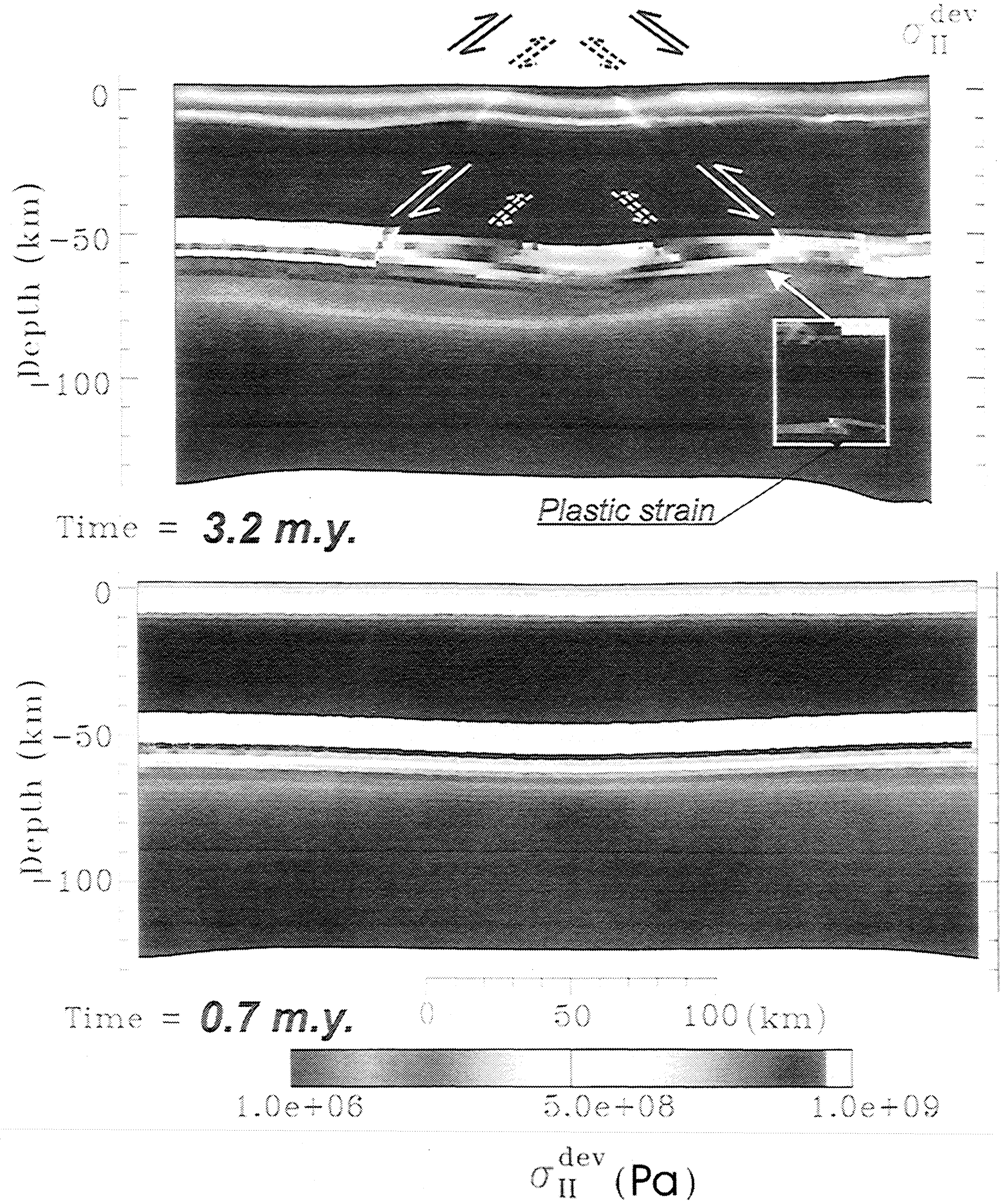

Figure 8. Folding and faulting of a thin brittle-elastic-ductile lithosphere (second invariant of the deviatoric shear stress, σ_{II}^{dev}) assuming quartz-dominated crust and olivine-dominated mantle and thermal age of 150 Ma. At the initial stage (t = 0.7 Ma, bottom) the lithosphere folds but remains laterally homogeneous. As the amplitude of folding grows with time (t=3.1 Ma, top), fold-associated stress concentrations result in formation of periodically spaced localized brittle shear zones (faults) in the upper crust and mantle, which can further coexist with folding. Note different wavelength of crustal (50-70 km) and mantle (200-250 km) folding. Velocity (10 mm/yr) lateral boundary conditions are used. To account for the sedimentary load, a diffusion erosion law at the surface is adjusted to keep the basin full [*Avouac and Burov*, 1996]. Insert in the white square (on top) shows accumulated plastic strain in the fault area (read 1.0e+06 as 1.0×10^{06}, etc.)

sediment from accounting for the mass deficit. Instead, it appears that the crust beneath the basin has been forced down. For a uniform difference in density between crystalline crust and mantle, the entire thickness of crust, including the thick sediment in the basin, is greater than that expected for isostatic equilibrium, if nevertheless apparently slightly thinner than that of the neighboring mountain ranges. In contrast to the neighboring mountain ranges, where horizontal shortening and thickening of crust have occurred, the absence of deformation of sediment within the valley implies that such shortening and thickening have not thickened the underlying crust.

Although small-scale mantle convection beneath the region could depress the crust above a narrow zone of downwelling, a second aspect of the gravity field seems to prohibit such a process from accounting for the isostatic gravity anomalies.

Adjacent to the large negative anomaly over the Ferghana Valley, a narrow gravity high surrounds the valley and requires that the relatively deep Moho beneath the valley be isolated from that beneath the neighboring mountains. Thus the Moho is forced down both beneath the surrounding mountains by isostatic compensation of thickened crust and by some other process beneath the Ferghana Valley, but between them, the crust is actually thinner than that expected for isostatic equilibrium. We attribute this pattern of deepened Moho to folding of the relatively thin uppermost mantle (mantle lithosphere), which is similar to the buckling of a thin elastic plate, loaded horizontally.

It appears that the Ferghana Valley and its surroundings were predisposed to such markedly inhomogeneous deformation because of two aspects of the preexisting structure. First, in part, because of an apparent resetting of the thermal structure in Jurassic time, and second, because of relatively thick crust in the region, the effective elastic thickness of the lithosphere in this region is small. This thin plate then allows deformation with a relatively short wavelength. Second, the distribution of Mesozoic sediment implies the existence of relatively deep basins adjacent to elevated terrain in Mesozoic time. Thick marine sediment was deposited both in the Ferghana Valley and in the Tadjik Depression south of the South Tien Shan, but not within the South Tien Shan or in the Chatkal Ranges north of the valley. This variation in elevation presumably was compensated, at least in part, by variations in crustal thickness, which in turn manifested itself as variations in temperature at the Moho. As lithospheric strength is parameterized well by the temperature at the Moho [e.g., *England*, 1986], we may presume that this preexisting topography implies a preexisting distribution of strength, which then resulted in high areas deforming into higher, mountainous regions with thick crust, and areas below sea level being too strong to deform significantly. The Ferghana Valley, being relatively strong, did not deform, but nevertheless was forced down by the horizontal loads applied to it. Comparison of the gravity anomalies over the Ferghana basin, Issyk-Kul (to the east of Ferghana) and Tadjik depression (to the west) suggests the possibility of a similar folding scenario for these basins.

Some of our experiments suggest that folding of the lithosphere in the Ferghana Basin could be followed by formation of zones of localized brittle failure such as deep crustal and mantle faults. This result is, however, very sensitive to the assumed parameters of the brittle-elastic-ductile rheology laws.

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