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Geoid undulation modelling and interpretation at Ladak, NW Himalaya using GPS and levelling data

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Abstract. Fast and accurate relative positioning for baselines less than 20 km in length is possible using dual-frequency Global Positioning System (GPS) receivers. By measuring orthometric heights of a few GPS stations by differential levelling techniques, the geoid undulation can be modelled, which enables GPS to be used for orthometric height determination in a much faster and more economical way than terrestrial methods. The geoid undulation anomaly can be very useful for studying tectonic structure. GPS, levelling and gravity measurements were carried out along a 200-km-long highly undulating profile, at an average elevation of 4000 m, in the Ladak region of NW Himalaya, India. The geoid undulation and gravity anomaly were measured at 28 common GPS-levelling and 67 GPS-gravity stations. A regional geoid low of nearly −4 m coincident with a steep negative gravity gradient is compatible with very recent findings from other geophysical studies of a low-velocity layer 20–30 km thick to the north of the India–Tibet plate boundary, within the Tibetan plate. Topographic, gravity and geoid data possibly indicate that the actual plate boundary is situated further north of what is geologically known as the Indus Tsangpo Suture Zone, the traditionally supposed location of the plate boundary. Comparison of the measured geoid with that computed from OSU91 and EGM96 gravity models indicates that GPS alone can be used for orthometric height determination over the Higher Himalaya with 1–2 m accuracy.

Key words. Geoid · GPS · Himalaya · Tectonics · Suture zone

1 Introduction

Three-dimensional relative positioning with an accuracy of the order of 0.1 ppm is now possible using Global Positioning Systems (GPS) (Davis et al. 1989; Dong and Bock 1989; Dixon 1991; Springer et al. 1994). However, the elevation thus obtained is relative to the spheroid and has very little practical use. In most scientific, engineering and surveying work, heights relative to the geoid, a close approximation to mean sea level, are required – so called orthometric heights. The GPS-derived spheroidal heights (h) can be converted into orthometric heights (H) if the geoid–spheroid separation (N) of the area is known.

$$N = h - H + \varepsilon \tag{1}$$

where ε is a quantity, usually small, due to the deflection of the vertical and the curvature of the plumb line (Torge 1980). This can be eliminated if, instead of the absolute geoid separation, we consider the separation variation or undulation (ΔN) given by (Engelis et al. 1984, 1985)

$$\Delta N = (h_2 - h_1) - (H_2 - H_1) \tag{2}$$

Also, by using both GPS and levelling techniques it is possible to map the geoid separation over the continental surface of the Earth (Zilkoski 1990; Milbert 1991, 1992); this can provide useful tectonic information about lithospheric structure (Crough 1979; Turcotte and Harris 1984; Parmentier and Haxby 1986; Cazenave 1993). Geoid undulation mapping has been done extensively over the sea surfaces using satellite-based altimetry (Rapp 1983; Engelis and Knudsen 1989; Zlotnicki 1993). Limited work has been done over the continents, mainly because of the lack of spheroidal heights over the continental surfaces, although gravimetrically determined geoids have been studied exhaustively (Milbert 1992; Sjoberg and Nord 1992; Rapp 1993; Sideris 1993; Lemoine et al. 1997). The latest geopotential model, EGM96 (Lemoine et al. 1997), was developed jointly by the NASA Goddard Space Flight Center, the National Imagery and Mapping Agency and the Ohio State University, using: (a) satellite tracking data as well as altimeter data from TOPEX, GEOSAT and ERS-1; (b) $30' \times 30'$ terrestrial gravity data from

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National Imagery and Mapping Agency (NIMA) archives; and (c) 30′ × 30′ gravity anomalies derived from the GEOSAT Geodetic Mission altimeter data. When the EGM96 model was evaluated using GPS/levelling tests in the USA over more than 1000 stations, a mean difference of −1.12 m was obtained between the measured geoid undulations and EzGM96-derived geoid undulations (Lemoine et al. 1997). Understandably, the error in geopotential model is much larger in the Ladak Himalayan region, this being the most tectonically disturbed and topographically deformed and elevated part on the surface of the earth and with almost total absence of surface gravity data. In the region under study in Ladak in Western Himalaya, covering an area of approximately 20 000 km^2, not more than five surface gravity measurements along the Rumtse–Leh road (Das et al. 1979) existed prior to the present work.

We measured gravity and geoid undulations along a 200-km-long Rumtse–Leh–Panamik section in Ladak Himalaya (Fig. 1), by using 'GPS on levelling line' (Mainville and Veronneau 1990). Besides studying the deep crustal tectonic implications of the variation of geoid undulation while crossing the India–Asia plate boundary, an objective of the work was to study the viability of using GPS in 'fast-static' mode for determining the orthometric heights (of gravity stations) with the help of geoid undulation models derived from existing global geopotential models like OSU91 and EGM96. Both tectonically and topographically this profile presented the most severe test conditions as it passes over the highest motorable road in the world, at a height of 5400 m, and a sharply undulating topography. Whereas 28 GPS-levelling stations were established for the geoid undulation measurements, a total of 67 gravity stations were established, and GPS-derived heights for these gravity stations were converted into orthometric heights using the geoid undulation model.

The region under study has great tectonic significance as it covers both the Indus Tsangpo Suture and the Shyok Suture, which are believed to represent the collision boundary between the Indian and Eurasian-plates (Thakur 1981; Wang and Shi 1982; Lyon-Caen and Molnar 1985). This unique continental–continental plate boundary makes a very interesting case for studying the variation of the geoid while crossing from one continental plate to the other.

The geoid model along this transect, in combination with Bouguer gravity anomaly data, has revealed some very interesting crustal features regarding the India–Asia plate collision tectonics. Although the same information is present in the geoid and the gravity anomaly data, the geoid anomalies are more suitable for studying the deeper lithospheric structures, whereas shorter wavelength signals arising out of shallower mass inhomogeneities dominate in the gravity anomalies (Hayling 1994). Simple mathematical modelling of the step-func-

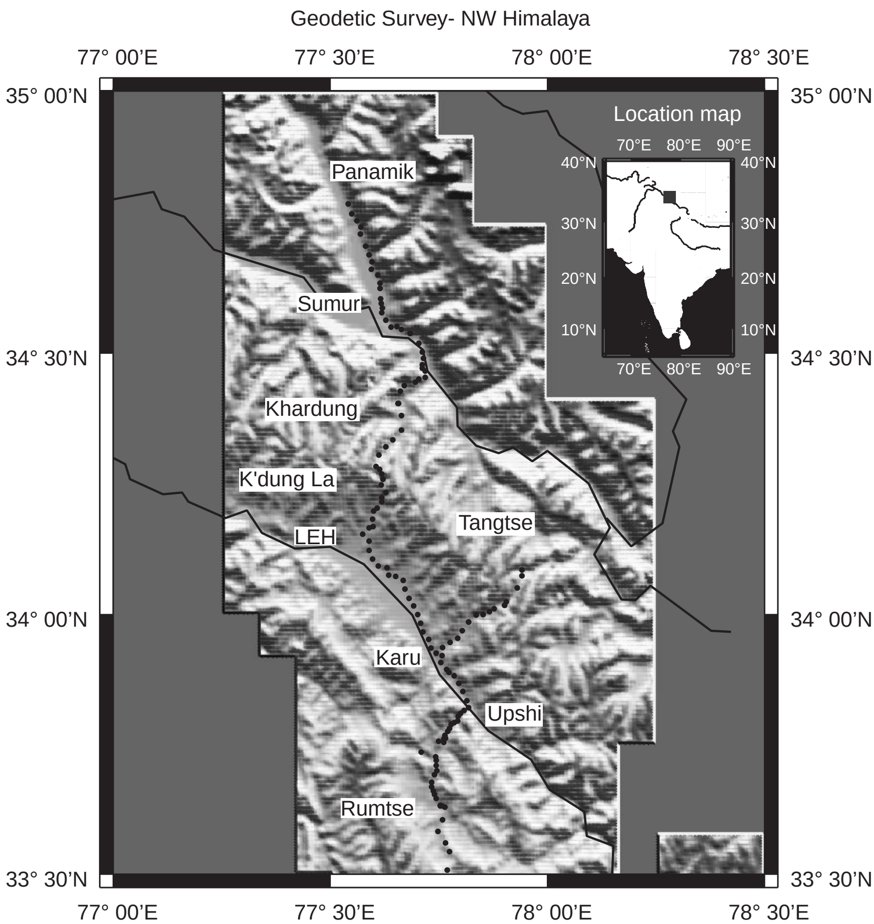

Fig. 1. Surface relief map of the Ladak region showing the GPS, levelling and gravity stations along the Rumtse–Panamik profile

tion type geoid undulation anomaly and the Bouguer gravity anomaly indicates the presence of a thick, low-velocity layer at mid-crustal depth in the southern part of the Tibetan tectonic plate, to the north of the suture zone. This finding is consistent with the results of IN-DEPTH-II studies in Nepal Himalaya, suggesting a similar low-velocity layer from seismic studies (Nelson et al. 1996). We used OSU91 and EGM96 to compute the geoid heights along the same transect. The OSU91-derived geoid matches more closely with our measured model with a mean difference of −0.382 m, compared to EGM96 where the mean difference is −2.96 m, and both the computed models lack the higher frequency signal, as is expected.

2 Field measurements

GPS, levelling and gravity measurements were carried out in the period 1994–96 along a 200-km-long profile passing through Rumtse, Upshi, Leh, Khardung La, Sumur and Panamik, in the Ladak region of NW Himalaya (Fig. 1). The profile cuts across the Indus Tsangpo Suture Zone, the Ladak granitic batholith, the Indus Shyok Suture zone and enters the Nubra valley. The topography along the profile varies from 3000 to 5400 m. Gravity measurements were carried out using Scintrex CG3M and Lacoste-Romberg gravimeters. As no previous gravity measurements exist from this region, a gravity base station was established at Leh, tying it to the Lamayuru and Kargil gravity base stations of the Survey of India, nearly 150 and 300 km further west respectively. The measurement errors of the gravity readings are less than 0.1 mGal. The positioning of the gravity stations was done using GPS. Orthometric heights for all the gravity stations were obtained by using linear interpolation over the geoid undulation measurements.

The GPS measurements using two dual frequency (Trimble 4000SSE) receivers were carried out in 'fast-static' mode, keeping one receiver fixed at a base station and running continuously in static mode while using the other receiver as a rover and occupying remote stations for a period of 8–20 min, depending on the availability of satellites. When operated in 'fast-static' mode, the receiver indicates automatically when sufficient data have been collected. The measured 'fast-static' baselines were all < 20 km in length. GPS base stations were established at Rumtse, Upsi, Leh, Khardung and Sumur (Fig. 2). Bench-marks were established for the base stations by drilling a 2-mm diameter hole on hard rocks, mostly on hill tops. Leh base station is on top of a small granitic hill at an isolated place nearly 4 km away from Leh city. All other base stations were tied to the Leh base station with a common occupation period of 5–6 hours. The Leh base station was occupied for 5 days continuously both in 1995 and 1996, and was tied to the Kitab (KIT3), Lhasa (LHAS) and Bangalore (IISC) IGS stations, while processing the data. In total 67 fast-static GPS stations were established over which gravity measurements were also done and, out of these, levelling measurements were made over 31 stations.

The levelling surveys were conducted using a Sokkia Set 2C Total Station, a microprocessor-based EDM–Theodolite combination. The levelling network was tied to Survey of India (SOI) bench-marks at Rumtse, Upsi, Igu and Leh. As no levelling line exists north of Leh, new bench-marks were established at South Pulu and Khardung, on the southern and northern flanks of Khardung peak. The Leh–Khardung section of the levelling line represents the steepest topography as elevation changes from 3400 m at Leh to 5400 m at Khardungla, within a horizontal distance of 15 km. Levelling work over this section was repeated twice, in 1995 and in 1996, with the usual procedure of loop closure, and incorporating refraction corrections for each EDM-Target setting. Following the restriction policy of SOI on the levelling data, the precesion of the orthometric heights of the existing bench-marks supplied by the SOI were limited to decimetre level.

3 Data analysis

The GPS data were processed using the Bernese Processing Engine of the Bernese 4.0 software. IGS precise orbits and pole files were used for the entire processing. The processing was split into different parts and very long (1200–2000 km), long (40–100 km) and short (2–20 km) baselines were processed using Melbourne–Wubbena, Quasi Ionosphere Free and SEARCH ambiguity resolution strategies (Rothacher and Mervart 1996). Leh (SABU, Fig. 2) station was tied with Kitab (KIT3) and Lhasa (LHAS) IGS GPS stations. The precise coordinates of the IGS stations were taken from the Institute Geographique National web site (http://lareg.ensg.ign.fr/ITRF). Other intermediate base stations were tied to Leh, and the fast-static stations forming short baselines with respective base stations were then processed.

The GPS observation times corresponding to very long baselines are 3–5 days, for intermediate baselines are 5–8 hours, and for short baselines are 8–20 min. The standard deviation of single-difference observations used for the final (floating-point) parameter estimations were of the order of 0.001 m, and the rms errors of the station coordinates were of the order of 0.001 arcsec for latitude and longitude and of the order of 0.01 m for the ellipsoidal height. These are unscaled formal errors, and the numbers of data are small as the observation windows were short. They give a substantial underestimate of the true errors, which are probably a factor of several larger. The repeatability cannot be assessed as each point was measured only once, except the stations along the Leh–Khardung La section where repeat GPS measurements over the stations were carried out in 1995 and 1996. The repeatability of these stations is 1–5 cm. The scatter of the profile points about a smooth geoid undulation curve (Fig. 4c) is up to a few tens of cm, and this is likely to be the approximate level of the combined true error of GPS and levelling measurements. The long baselines of the order of 30–50 km had rms errors in length of the

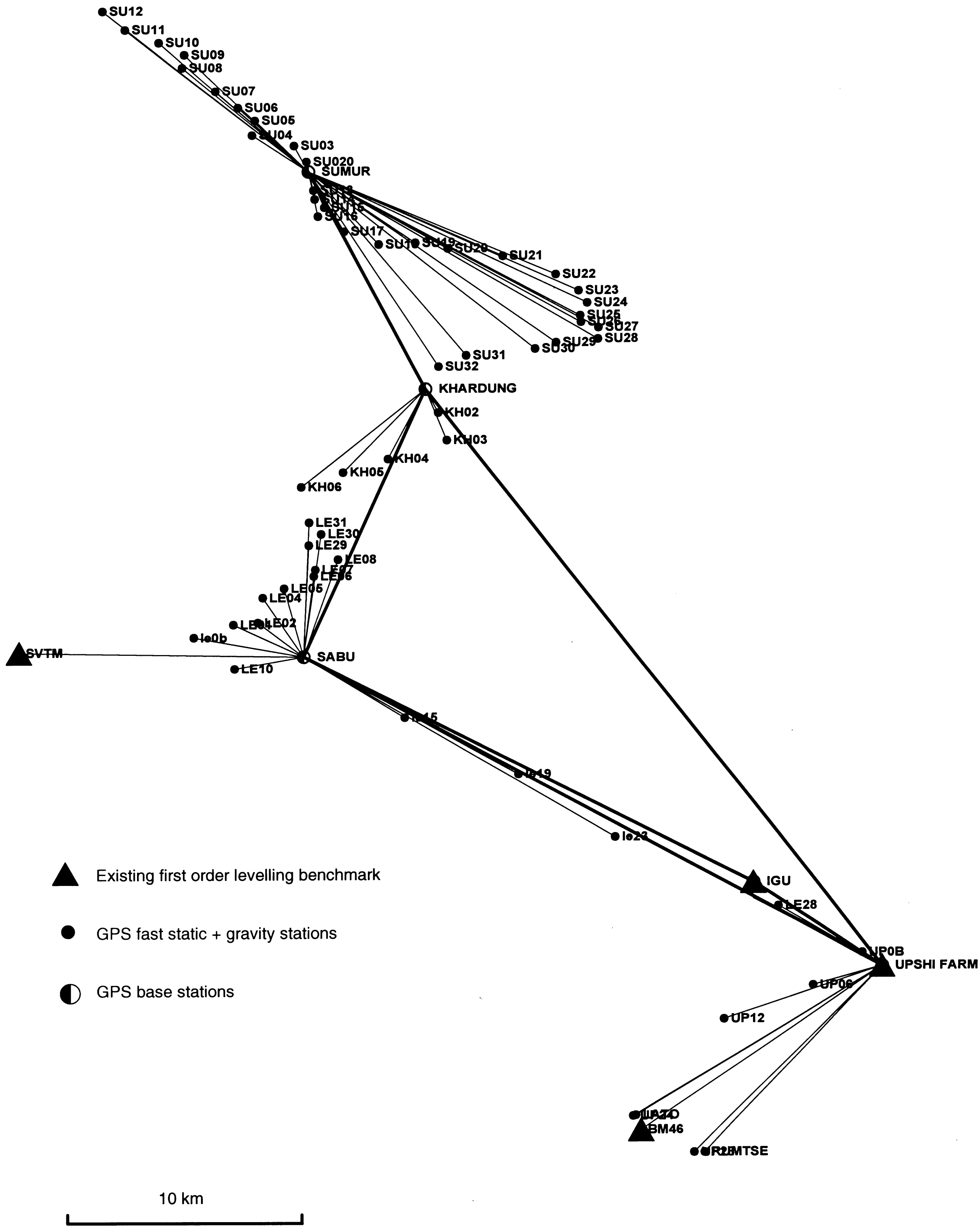

Fig. 2. Fast-static GPS network in Ladak. New levelling bench-marks are tied with existing first-order bench-marks. Leh (SABU) GPS base station is tied to Lhasa and Kitab IGS stations

order of 0.001 m. The coordinates of Leh base station and consequently of all the GPS stations are in ITRF94 system.

Levelling data were processed using an F77 program written by the first author. Because of the steepness of the survey line, several couples of vertical angle measurements were taken. EDM distance measurements were restricted to 3–5 times for each sighting because of good repeatability, and also to save power in extremely cold conditions. Loops were closed between each two consecutive GPS stations (1–3 km), which generally required 2–4 instrument settings. Closure errors (1–6 cm) were distributed uniformly. The relative accuracy of the station heights is within 2–5 cm. The whole network was tied to SOI bench-marks which are accurate to within few mm; however, the precision of the elevation data supplied to us was limited to decimetre level. Therefore the absolute accuracy of orthometric heights measured can be stated to be within 10 cm, although the measurement accuracy is much better. Geoid heights were obtained (Fig. 3) on each GPS-levelling station by simply taking the difference between the WGS84 spheroidal heights and orthometric heights, and neglecting the ε term which is usually small and below the combined GPS and levelling measurement error level.

We also computed the theoretical geoid values over the same points using two different global gravity models, i.e. OSU91 and EGM96. For OSU91 geoid computations, we used the OSU91A1F set of spherical harmonic coefficients, whereas for EGM96 (Lemoine et al. 1997) spherical harmonic coefficients as well as correction coefficient data set were downloaded from the NIMA web site (http://164.214.2.59). Slightly modified versions of a Fortran program developed by Rapp (1997) were used to compute geoid undulation values at our measurements points. The measured and computed geoid heights along the Rumtse–Panamik profile are shown in Fig. 3.

It is seen that the measured geoid matches closely with that derived from the OSU91 gravity model with a mean difference of −0.4 m, whereas the mean difference with the EGM96-derived geoid is nearly −3 m. This mean difference is the net effect of vertical datum offsets with the orthometric heights, non-geocentricity in the ellipsoidal heights, as well as errors in the geopotential model.

The geoid heights measured over 31 stations were interpolated by using simple linear interpolation technique over the rest of the gravity/GPS stations along the profile to obtain orthometric heights. These were used for reducing surface gravity measurements to Bouguer anomaly (Fig. 4b). Terrain corrections were applied using a new technique introducing a digital terrain model (Banerjee 1998). This model extends out to 170 km around each gravity station and uses 112 compartmentalized elevation data points covering the range 0 to 1.2 km, $30'' \times 30''$ digital terrain data for 1.2 to 20 km and $5' \times 5'$ global NGDC topographic data for the 20 to 170 km range. $30'' \times 30''$ data were manually digitized from 1:25 000-scale SOI topographic maps. The Bouguer anomalies are correct to within 1–3 mGal, and the errors are mainly caused by the uncertainty and variations in density within the thick (3–5 km) Bouguer slab.

The geoid anomaly is almost smooth along the Rumtse–Khardung La section (Fig. 4c), after which it drops down by nearly 4 m over a distance of 40–50 km and is again flat thereafter. This step-like change in the geoid anomaly is significant here as the profile crosses the Indian–Tibetan plate boundary. The surface geological signature of the plate boundary, i.e. the Indus Tsangpo Suture Zone (ITSZ), is located south of Leh. However, due to intense tectonic activity, it is possible that the shallow surfacial rocks were pushed further south. Gravity and other geophysical signals can significantly improve the knowledge about the deeper configuration of the tectonic set up. The southern part of the Tibetan plateau is known to have overridden a large part of the Indian continental crust (Ni and Barazangi 1983). A residual of the underthrusted India plate has been identified as a low-velocity/low-density slab at mid-lithospheric depth by other geophysical studies (Zhao and Morgan 1985; Nelson et al. 1996; Rodgers and Schwartz 1997). A simple mathematical model is used here to interpret the change in the geoidal anomaly in terms of tectonics, assuming that two different, semi-infinite blocks represent the India and Tibetan plates, that these are in isostatic equilibrium, and the observed geoidal anomaly change is produced by the lateral density contrast.

In the flat-earth approximation, the condition of isostatic equilibrium (Turcotte and McAdoo 1979) is

$$\int_0^h \Delta\rho(x,z)\,\mathrm{d}z = \text{constant} \tag{3}$$

where h is the depth of compensation.

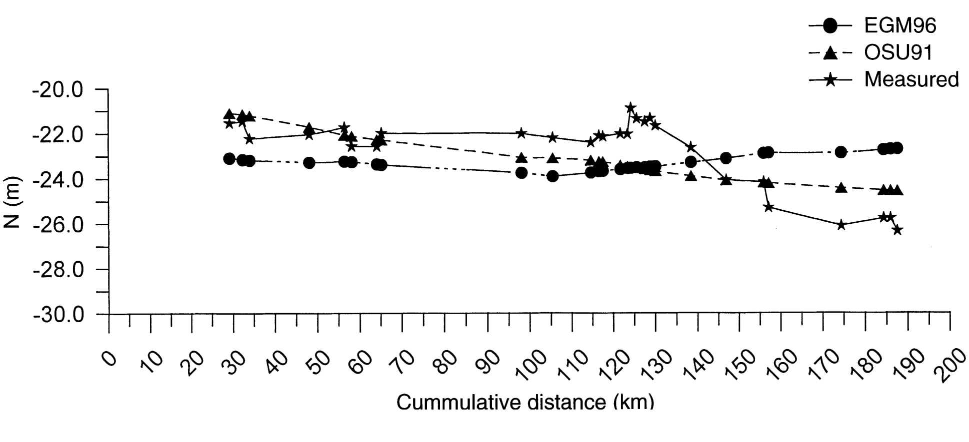

Fig. 3. A comparison of measured geoid undulation with EGM96- and OSU91-derived geoid models along Rumtse–Panamik section. OSU91 has a downward trend in the northern part in agreement with the measured geoid. In contrast, EGM96 model shows an upward trend in the northern part

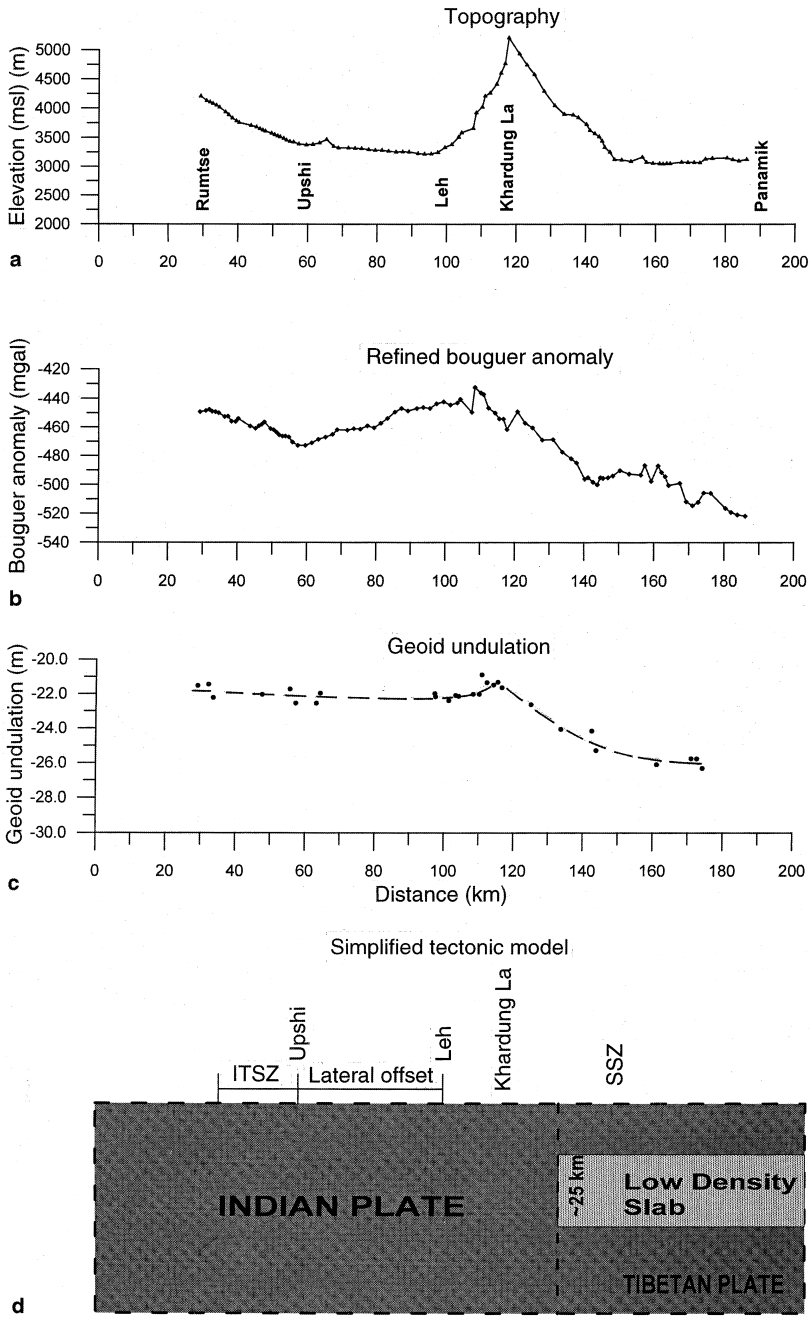

Fig. 4a–d. **a** Topography, **b** Bouguer gravity and **c** Geoid undulation anomalies along the Rumtse–Panamik profile. **d** Model of a 20–25-km-thick low-velocity layer within the Tibetan plateau. The actual plate boundary may be located north of Khardung La peak, beneath a granitic batholith, rather than at the Indus Tsangpo Suture Zone (ITSZ)

The gravity potential anomaly due to density inhomogeneity $\Delta\rho(x,z)$ is given (Angevine and Turcotte 1983) by

$$\Delta u = 2\pi G \int_0^h z\Delta\rho(x,z)\,\mathrm{d}z \qquad (4)$$

and the corresponding geoid undulation anomaly is given (Haxby and Turcotte 1978) by

$$\Delta N = -2\pi G/g \int_0^h z\Delta\rho(x,z)\,\mathrm{d}z \qquad (5)$$

Assuming an infinite horizontal slab of thickness h, Eq. (5) reduces to

$$\Delta N = 2\pi G/g\Delta\rho h^2 \qquad (6)$$

The assumption of an infinite horizontal slab is valid only when the computation point is sufficiently far away from the edge of the slab. Observing that the geoid undulation anomaly is about −4 m, a slab with a thickness of 20–25 km and a density contrast of $-0.2\ \mathrm{gm/cm^3}$ can be predicted to the north of the suture zone (Fig. 4d). A regional gravity low of about 60–80 mGal coincides with the geoid low (Fig. 4b and c). This is in approximate agreement with this model.

4 Discussion and conclusions

A nearly 70-km-thick crust was suggested for the Tibetan plateau by Jin et al. (1996). A low-velocity/

density layer inside the Tibetan plate to the north of the suture zone was suggested by Nelson et al. (1996) and Rodgers and Schwartz (1997). Thickening of the southern Tibetan plateau was suggested by McNamara (1995) and Dewey et al. (1988). It was suggested that a low-velocity layer may be associated with a laterally-pervasive weak zone that mechanically decouples the upper crust from the lower crust (Zhao and Morgan 1985; Jin et al. 1996), or simply an isolated zone of melted crust. Rodgers and Schwartz (1997) predicted a 20-km-thick mid-crustal low-velocity layer to the north of the suture zone. The presence of crustal melting is plausible given the existence of the huge Ladak granitic batholith north of the river Indus.

A second important conclusion that can be drawn from the geoid anomaly is that the actual boundary between the Indian and Tibetan plates is between Khardung La peak and the Shyok suture zone, rather than south of the river Indus, at what is known geologically as the ITSZ. The surface signature of the suture zone is perhaps masked by the Ladak batholith which lies right over the plate boundary.

GPS, when applied to geoid undulation modelling over continental regions, may lead to important tectonic findings. As geoid undulation is more affected by deep-seated bodies than the gravity anomalies, the geoid anomaly can be used very effectively to find major, deep tectonic discontinuities and also to contribute to regional residual separation for gravity anomalies prior to modelling.

GPS-derived ellipsoidal height can be used for computing 'gravity disturbances' for geophysical interpretation. However, to compute 'gravity anomalies' (Free Air, Bouguer, Isostatic etc.) ellipsoidal heights need to be converted into orthometric heights as has been discussed. Comparison between measured geoid and computed geoid suggests that for low-accuracy requirements such as gravity studies at a regional scale where a height accuracy of 1–2 m is acceptable for the Himalayan region, models like OSU91 can be used for directly converting spheroidal heights into orthometric heights. As this eliminates the necessity of levelling measurements in very hostile terrain conditions, gravity surveys become much faster and economical. In fact, this has remained one of the major reasons for the extremely sparse gravity coverage on Higher Himalaya. However, for more stringent accuracy requirements for deriving gravity anomalies, a few levelling stations have to be established as the global geoid models lack the higher-frequency components due to the almost total absence of surface gravity data for this region in the global gravity model.

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References

Angevine CL, Turcotte DL (1983) Correlation of geoid and depth anomalies over the Agulhas Plateau. Tectonophysics 100: 43–52

Banerjee P (1998) Gravity measurements and terrain corrections using digital terrain model in NW Himalaya. Comput Geosci 24, 10: 1009–1020

Cazenave A (1993) The geoid and oceanic lithosphere. In: Vanicek P, Christou NT (eds) Geoid and its geophysical interpretations. CRC Press, London, ch 13, pp 255–283

Crough ST (1979) Geoid anomalies across fracture zones and the thickness of the lithosphere. Earth Planet Sci Lett 44: 224

Das D, Mehra G, Rao KGC, Roy AL, Narayan MS (1979) Bouguer, free air, and magnetic anomalies over NW Himalayas. Him Geol Sem 1976, Sect. III, GSI Misc. Publ. No. 41, pp 141–147

Davis JI, Prescott WH, Svarc JL, Wendt KJ (1989) Assessment of global positioning system measurements for studies of crustal deformation. J Geophys Res 94: 635–650

Dewey J, Shackleton R, Chengfa C, Yiyin S (1988) Tectonic evolution of the Tibetan Plateau. Phil Trans R Soc Lond, Ser. A 327: 379–413

Dixon TH (1991) An introduction to the Global Positioning System and some geological applications. Rev Geophys 29: 249–276

Dong DN, Bock Y (1989) Global Positioning System network analysis with phase ambiguity resolution applied to crustal deformation studies in California. J Geophy Res 94: 3949–3966

Engelis T, Knudsen P (1989) Orbit improvement and determination of the ocean geoid and topography from 17 days of Seasat data. Manuscr Geod 14: 193

Engelis T, Rapp HR, Tscherning CC (1984) The precise computation of geoid undulation differences with comparison to results obtained from the Global Positioning System. Geophysical Research Letters, Vol. 1, No. 9, pp. 821–824

Engelis T, Rapp HR, Bock Y (1985) Measuring orthometric height differences with GPS and gravity data. Manuscr Geod 10: 187–194

Haxby WF, Turcotte DL (1978) On isostatic geoid anomaly. J Geophy Res 83: 5473–5478

Hayling KL (1993) The geoid and geophysical prospecting. In: Vanicek P, Christou NT (eds) Geoid and its geophysical interpretations. CRC Press, London ch 15

Jin Y, McNutt M, Zhu Y (1996) Mapping the descent of the Indian and Eurasian plates beneath the Tibetan plateau from gravity anomalies. J Geophys Res 101, B5: 11275–11290

Lemoine FG et al. (1997) The developement of the NASA GSFC and NIMA joint geopotential model. In: Segawa et al. (eds) Int Ass Geod Symp 117. Springer, Berlin, Heideberg, New York

Lyon-Caen H, Molnar P (1985) Gravity anomalies, flexure of the Indian plate, and the structure, support and evolution of the Himalaya and Ganga Basin. Tectonics 4: 513–538

Mainville A, Veronneau M (1990) Orthometric heights using GPS in Canada. In: Rapp RH, Sanso F (eds) Determination of the geoid present and future. Int Ass Geod Symp 106, 11–13 June 1990, pp 160–172, Springer-Verlag, New York

McNamara DE, Owens TJ, Walter WR (1995) Observations of regional phase propagation across the Tibetan plateau. J Geophys Res 100, B11: 22215–22229

McNamara DE, Walter WR, Owens TJ, Ammon CJ (1997) Upper mantle structure of the Tibetan Plateau from Pn travel time tomography. J Geophys Res 102, B1: 493–505

Milbert DG (1991) GEOID90; a high-resolution geoid for the United States. EOS, Trans Am Geophys Union 72, 49: 545–554 (vol. 72, no. 49)

Milbert DG (1992) Intercomparison of the Geoid90 model against ellipsoidal heights on benchmarks. EOS, Trans Am Geophys Union 73, 14 sp suppl

Nelson KD, Zhao Wenjin, Brown LD, Kuo J, Che Jinkai, Liu Xianwen, Klemperer SL, Makovsky Y, Meissner R, Mechie J, Kind R, Wenzel F, Ni J, Nabelek J, Chen Leshou, Tan Handong, Wei Wenbo, Jones AG, Booker J, Unsworth M, Kidd

WSF, Hauck M, Alsdorf D, Ross A, Cogan M, Wu Changde, Sandvol E, Edwards M (1996) Partially molten middle crust beneath southern Tibet; synthesis of Project INDEPTH results. Science 274: 1684–1688

Ni J, Barazangi M (1983) High-frequency seismic wave propagation beneath the Indian Shield, Himalayan Arc, Tibetan Plateau and surrounding regions; high uppermost mantle velocities and efficient Sn propagation beneath Tibet. Geophys J R Astronom Soc 72: 665–689

Nord T, Sjoberg LE (1992) A spectral analysis of geoid undulation and gravity anomaly data computed with Pratt's isostasy theory applied to Moho depth variations in Fennoscandia. Geophys J Intl 111, 1 179–184

Parmentier EM, Haxby WF (1986) Thermal stress in the oceanic lithosphere: evidence from geoid anomalies at fracture zones. J Geophys Res 91: 7193

Rapp RH (1983) The determination of geoid undulations and gravity anomalies from SEASAT altimeter data. J Geophys Res 88(C3): 1552

Rapp RH (1993) Global geoid determination. In: Vanicek P, Christou NT (eds) Geoid and its geophysical interpretations. CRC Press, London, ch 3, pp 57–76

Rapp RH (1997) Use of potential coeffiecient models for geoid undulation determination using a spherical harmonic representation of the height anomaly/geoid undulation difference. J Geod 71: 282–289

Rodgers JA, Schwartz YS (1997) Low crustal velocities and mantle lithospheric variations in southern Tibet from regional Pnl waveforms. Geophys Res Lett 24: 9–12

Rothacher M, Mervart L (1996) Documentation of the Bernese GPS software version 4.0. Astronomical Institute, Univ Berne

Sideris MG (1993) Regional geoid determination. In: Vanicek P, Christou NT (eds) Geoid and its geophysical interpretations. CRC Press, London, ch 4, pp 77–94

Springer TA, Ambrosius BAC, Noomen R (1994) Results of the WEGNER/GPS-92 Campaign, Geophys Res Lett 21(16): 1711–1714

Thakur VC (1981) Regional framework and geodynamic evolution of the Indus–Tsangpo suture zone in the Ladak Himalaya. Trans R Soc Edinburgh: Earth Sci 72: 89–97

Torge W (1980) Geodesy. Walter de Gruyter, Berlin

Turcotte DL, Harris RA (1984) Relationship between oceanic geoid and the structure of the lithosphere. Mar Geophys Res 7: 177

Turcotte DL, McAdoo DC (1979) Geoid anomalies and the thickness of the lithosphere. J Geophys Res 84: 2381–2387

Wang Chi-Yuen, Shi Yaolin (1982) On the tectonics of the Himalaya and the Tibet plateau. J Geophys Res 87: 2949–2957

Zhao W, Morgan W (1985) Uplift of the Tibetan Plateau. Tectonics 4: 359–369

Zilkoski DB (1990) Minimum steps required when estimating GPS-derived orthometric heights. Proc. GIS/LIS'90 Fall convention, Anaheim, CA

Zlotnicki V (1993) The geoid from satellite altimetry. In: Vanicek P, Christou NT (eds) Geoid and its geophysical interpretations. CRC Press, London, ch 5, pp 95–110