Hydro-Isostatic Deflection and Tectonic Tilting in the Central Andes: Initial Results of a GPS Survey of Lake Minchin Shorelines

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Abstract. Sufficiently large lake loads provide a means of probing rheological stratification of the crust and upper mantle. Lake Minchin was the largest of the late Pleistocene pluvial lakes in the central Andes. Prominent shorelines, which formed during temporary still-stands in the climatically driven lake level history, preserve records of lateral variations in subsequent net vertical motions. At its maximum extent the lake was 140 m deep and spanned 400 km N-S and 200 km E-W. The load of surficial water contained in Lake Minchin was sufficient to depress the crust and underlying mantle by 20-40 m, depending on the subjacent rheology. Any other differential vertical motions will also be recorded as departures from horizontality of the shorelines. We recently conducted a survey of shoreline elevations of Lake Minchin with the express intent of monitoring the hydro-isostatic deflection and tectonic tilting. Using real-time differential GPS, we measured topographic profiles across suites of shorelines at 15 widely separated locations throughout the basin. Horizontal and vertical accuracies attained are roughly 30 and 70 cm, respectively. Geomorphic evidence suggests that the highest shoreline was occupied only briefly (probably less than 200 years) and radiocarbon dates on gastropod shells found in association with the shore deposits constrain the age to roughly 17 kyr. The basin-wide pattern of elevations of the highest shoreline is composed of two distinct signals: (27 ± 1) m of hydro-isostatic deflection due to the lake load, and a planar tilt with east and north components of $(6.8 \pm 0.4)\ 10^{-5}$ and $(-5.3 \pm 0.3)\ 10^{-5}$. This rate of tilting is too high to be plausibly attributed to steady tectonism, and presumably reflects some unresolved combination of tectonism plus the effects of oceanic and lacustrine loads on a laterally heterogeneous substrate. The history of lake level fluctuations is still inadequately known to allow detailed inferences of crust and mantle rheology. However, it is already clear that the effective elastic plate thickness is closer to 40 km than the 60-70 km crustal thickness in the central Andes and the effective viscosity is less than $5\ 10^{20}$ Pa s .

Introduction

The altiplano of Bolivia is presently a cold high desert, but at elevations below 3800 m the landscape is dominated by coastal geomorphic features. During colder and/or wetter episodes in the past, the hydrologically closed basins of the altiplano have contained substantial lakes. The largest of these was lake Minchin, which extended 200 km E-W and 400 km N-S, and was 140 m deep. Despite having formed as a level surface, the highest shoreline of lake Minchin is presently found at orthometric (geoidal) elevations that range from 3750 to 3790 m. The spatial pattern of net vertical motions (accumulated since the shoreline was formed roughly 17,000 years ago), contains important new information about the response of the crust and upper mantle to normal loads from two sources (variations in the lake load itself, and fluctuations in sea level during the last deglaciation) and places constraints on the vertical tectonism associated with subducting the Nazca plate under the South America plate.

Depositional shorelines of large endorheic lakes have three essential properties that make them useful for geodynamic studies: they form quickly, they disappear slowly, and they are initially horizontal. As a result, the climatically forced oscillations in depth and areal extent of major lakes in hydrologically closed basins throughout the world provide a unique opportunity to investigate the Earth's response to normal loads on length scales of 10's to 100's of kilometers and time scales of decades to millennia. Gilbert's (1890) pioneering work on Lake Bonneville (in western Utah and parts of Nevada and Idaho) established the basic framework for limnologic neotectonics and has exerted a major influence on subsequent studies (Crittenden, 1963; Currey, 1982, 1990; Nakiboglu and Lambeck, 1982; Bills and May, 1987, Bills and Currey, 1993).

Though much of the literature on rheology of the crust and upper mantle is concerned with response to major glacial loads (Peltier, 1974, 1976; Wu and Peltier, 1983; Mitrovica and Peltier, 1991, 1992; Lambeck, 1990), lake loads have several advantages over glacial loads. Primary among them is that the complex spatio-temporal pattern of the load can be easily and accurately reconstructed from two ingredients: present-day topography and a history of lake level fluctuations. In contrast, reconstruction of an ice sheet load requires at least three ingredients: geologic constraints on the temporal evolution of the ice margin, observations (or guesses) to constrain the variable boundary conditions at the ice-till interface, and a complex simulation of the non-linear flow of ice (Payne et al., 1989; MacAyeal, 1989; Boulton and Clark, 1990). Lakes are also much better recorders of climatic history than are glaciers and ice-sheets (Currey, 1990).

The early work of Agassiz (1876), Steinmann et al. (1904) and Bowman (1909) established that a series of large lakes have, at various times throughout the Pleistocene, occupied the presently semi-arid basins of the Bolivian altiplano. The northern basin, with a total area of 57.1 10^3 km^2, presently contains lake Titicaca (8.7 10^3 km^2 in area, 3810 m surface elevation) and previously held a somewhat larger lake, (13-14 10^3 km^2 area, ~3900 m surface elevation) which Bowman (1909) called lake Ballivian. The southern basin, with a total area of 134 10^3 km^2, presently contains only the shallow lake Poopo (3-4 10^3 km^2 area, 3660 m surface elevation) and the playas of Empexa, Coipasa and Uyuni, but previously held a single large lake (48-50 10^3 km^2 area, 3760 m surface elevation) which Steinmann et al. (1904) called lake Minchin. It is this southern basin lake which is the focus of our present study.

The chronology of fluctuations in the level of lake Minchin is still only poorly known. Early workers suggested that the large altiplano lakes may have resulted from melting of glaciers in the surrounding mountains. However, Hastenrath and Kutzbach (1985) showed that there was insufficient water stored in the glaciers to be a major source for the lake. The work of Servant and Fontés (1978) provided the first radiocarbon dates on shorelines of lake Minchin. They estimated that the highest stand of the lake (at close to 3800 m) occurred prior to 28 kyr ago. More recent work, summarized by Lavenu et al. (1984), suggests an age range of 22-27 kyr.

Servant and Fontes (1978) also pointed out that there was a late episode of pluvial activity, which they called the Tauca stage, that culminated near 3720 m elevation at about 12 kyr BP. This roughly coincides in age with the Younger Dryas climatic event, which was a brief return to full glacial conditions, and is seen in ice core records at both poles (Jouzel et al., 1987; Dansgaard et al., 1993) and in global sea level variations (Fairbanks, 1989). The Gilbert shoreline of Lake Bonneville also reflects this same global climatic event (Currey, 1990; Oviatt et al., 1992).

During much of the last deep-lake cycle on the altiplano, global sea-level stood ~120 m lower than at present (Fairbanks, 1989). The large-scale and long-term net effect of a ~120 m deglacial rise in sea-level is to depress the ocean basins relative to the continents by ~40 m. Interpretation of relative sea-level histories requires accurate computation of deflection at the coastline (Peltier, 1976; Lambeck, 1990), but there are relatively few direct observational constraints on how the sea-level induced deflection decays with distance from the coast. The spatial and temporal scales over which this deformation occurs are determined by the strength of the crust and upper mantle. For any significant ocean loading signal to be present in the Lake Minchin shorelines, which lie at distances of 100-300 km from the coastline, would require an effective lithospheric thickness of order 100 km.

Though many models of the tectonic evolution of the central Andes exist (Lyon-Caen et al, 1985; Isacks, 1988; Kono et al., 1989; Wdowinski and O'Connell, 1991; Baby et al., 1990), understanding of the spatio-temporal pattern of vertical motions associated with the subduction of the Nazca plate and the accompanying deformation of the South America plate is really quite limited. Existing observational constraints are few and mostly indirect. One of the most compelling observations is that the dip of the subducting slab (as indicated by patterns of earthquake hypocenter depths) varies significantly along the strike of the trench (James, 1971; Barazangi and Isacks, 1979; Hasegawa and Sacks, 1981), and the pattern of relief is significantly different over the steeply and shallowly dipping segments of the slab (Jordan et al., 1983; Dewey and Lamb, 1992). Marine terraces along the Pacific coastlines of Chile and Peru provide some constraints on uplift near the trench, and indicate that rates are spatially and temporally variable (DeVries, 1988; Hsu et al., 1989; Machare and Ortlieb, 1992). Fission track records of exhumation rates of plutonic rocks farther from the trench indicate temporal variability (Crough, 1983; Benjamin et al., 1987), but are often quoted as though they were useful measures of *the* uplift rate of the central Andes, rather than single point samples of a complex spatial pattern.

Risacher and Fritts (1991) recently presented evidence for regular variations in thickness of the upper salt layer in the Salar de Uyuni, which presently contains most of the salt delivered to lake Minchin by its tributary rivers during the last deep-lake cycle. The top of the salt unit is maintained as a level surface by annual episodes of rewetting and hydro-aeolian planation (Ericksen et al., 1978; Currey, 1990). The bottom of the salt unit occurs at depths that increase nearly linearly from <1 m on the west margin to ~10 m near the east edge. The variations in thickness of this salt unit could be taken as an indication of 10 m of tilting down to the east over the 100 km width of the salar since the end of the deep-lake cycle roughly 10-12 kyr ago. This tentative interpretation first suggested to us that we might see amounts of tectonic tilting comparable to the lacustrine hydro-isostatic rebound signal.

Method

Our topographic surveys of the shorelines of Lake Minchin were conducted using three Trimble 4000 SSE P-code GPS receivers, corresponding antennas and ancillary equipment. Each of the receivers played a different role. The "fiducial" receiver and its geodetic antenna occupied our two fiducial sites, one for 10 days, the other for 4 days. Data were collected at the fiducial sites for 10 hours each day. The "base" receiver and its geodetic antenna were transported to each survey locality, and remained fixed for the 1-2 hour duration of the topographic traverse of the suite of shorelines. The "rover" receiver and its compact dome antenna were mounted in a back-pack and were transported across the terrain to be measured.

Before starting a topographic traverse, the position of the "base" receiver was roughly estimated (horizontal position was taken from a 10-20 minute GPS solution, vertical was estimated from a 1:50,000 scale topographic map with 20 m contours). During the course of the topographic traverse, real-time estimates of the position of the "rover" antenna were obtained using the L1 signals from 5-8 GPS satellites and an RTCM signal transmitted from the "base" receiver. This allows differential correction of the pseudo-range signals from the satellites, and enables the position of the "rover" to be estimated, relative to the "base", with horizontal and vertical accuracies of 30 and 70 cm, respectively. Distance between "base" and "rover" never exceeded 2 km during these surveys.

After returning to the fiducial site, we processed the data collected by the "fiducial" and "base" receivers to obtain an improved estimate of the position of the "base" antenna during the surveys conducted that day. Differences between the improved estimate and the initial estimate were then added to the real-time estimates of "rover" locations. Distances between "fiducial" and "base" systems ranged up to 160 km. As the shorelines were initially formed on level (equipotential) surfaces, we corrected the ellipsoid heights initially obtained from the GPS data to orthometric (geoidal) heights.

At each of the traverse sites, we measured elevations of numerous (10-30) shoreline features. However, the task of correlating most of these features throughout the basin will likely prove difficult. The one conspicuous exception is the highest shoreline, which in every case was easily discernible by the complete lack of coastal geomorphic features at higher elevations on the surrounding terrain. As a result, we will limit our discussion to the pattern of elevations seen on this highest shoreline.

Results

Table 1 displays our measurements of the heights of the high shoreline of lake Minchin at 15 locations which are reasonably well distributed around the basin. Figure 1 locates the survey sites and also shows the configuration of the 3800 m topographic contour, which gives a rough idea of the shape of the lake. The contour information was extracted from 1:250,000 scale maps published in the early 1970's jointly by the U.S. Defense Mapping Agency and the Bolivian Instituto Geografico Militar.

We have obtained radiocarbon dates on samples collected from several shoreline contexts. Two of our samples pertain to the high shoreline. Both were obtained on gastropod shells collected from fine sand units lying 1-2 m below the crest of the high shoreline. One of the sample localities is near our survey site Yonza, the other sample was taken near the village of San Pedro de Quemez (20° 44' S, 68° 04' W). Age estimates for both sites were identical: (13,790 $\pm$ 70) radiocarbon years. Using the recent radiocarbon calibration of Bard et al. (1990), this corresponds to 16.8 kyr. This suggests that the high shoreline was occupied rather more recently than previous published estimates would indicate.

One of the primary incentives for our study is a desire to use the shoreline elevation patterns to refine the rheological model of the crust and upper mantle in the central Andes. However, attainment of that goal must await further progress in two areas: additional clarification of the spatio-temporal pattern of deformation, and improved definition of the lake level fluctuation history. At present, the best we can achieve is a simple exploration of the sensitivity of our measurements to the rheological stratification and a test of the hypothesis that the observed pattern of elevations can be reproduced by a linear superposition of two effects: a simple rebound model and a planar tilt.

Though the viscosity of crustal and upper mantle rocks is a complex function of (at least) composition and temperature, we confine our present analysis to two simple cases, involving either inviscid or uniformly viscous substrates, each overlain by an elastic lithospheric plate. In both cases the hydro-isostatic rebound was computed using the algorithm of Bills and May (1987), applied to an Earth model consisting of an elastic plate at the surface and a Maxwell visco-elastic half-space with a viscosity of either zero, or a value in the range 10^{18}-10^{22} Pa s. The surface load was computed from the digital elevation model of the central Andes compiled by Isacks (1988) and the lake surface elevation history was based on our estimates of the high shoreline age and the published lake level curve of Servant and Fontes (1978).

The inviscid case has the advantage of requiring no knowledge of the loading history, and the elastic plate thickness obtained in this case in a firm upper bound for the finite viscosity cases. We simply compare the observed deflection to that computed from the water load at the highest shoreline. In the finite viscosity case we are obliged to specify a lake surface elevation history, which we have taken to be asymmetric and piece-wise linear, with a slow rise from zero to maximum over the interval 28-17 kyr, and a rapid decline back to zero over the interval 17-14 kyr. Though the actual history is doubtless more complex, this simple model is broadly consistent with known constraints. Small adjustments to the starting and ending times of the loading interval will have little impact on estimated viscosities, since the age of occupation of the high shoreline is now known.

During initial exploration of these models it became evident that the observed shorelines contained a signal that is not directly related to rebound. In fact, the residuals displayed a distinctively planar trend. As a result, we simply postulated that the observations might be approximated by the linear combination

$$\Delta z_{obs} = A + \Delta z_{calc} + S_x * \Delta(\text{east}) + S_y * \Delta(\text{north}) \qquad (1)$$

where, Δz_{obs} is observed geoidal height of the high shoreline elevation, referred to an arbitrary standard value of 3750 m, A is an adjustment to the reference surface elevation, Δz_{calc} is the calculated deflection at the relevant location, S_x and S_y are the (east and north) slopes of the planar regression surface, and Δ(east) and Δ(north) are the distances east and north from an arbitrary reference point located near the center of the lake.

For the inviscid model, we examined plate thicknesses from 10 to 75 km. The best fitting value was 38 km, where the residual variance was 32.8% of the data variance. All plate thicknesses from 26 to 63 km produced residual variances below 40% of data variance. While it is clear that the inviscid substrate model is overly simplistic, this modeling exercise does clearly demonstrate that the mechanical response of the crust and upper mantle in the central Andes to applied surface loads reflects a "strong" layer with thickness less than that of the crust alone.

Values for the other model parameters, obtained via a simple least-squares solution, are: $A = (9.19 \pm 0.71)$ m, $S_x = (6.8 \pm 0.4)\ 10^{-5}$, and $S_y = (-5.3 \pm 0.3)\ 10^{-5}$. We also considered a model which included quadratic terms $[S_{xx}*(\Delta x)^2 + S_{xy}*(\Delta x)(\Delta y) + S_{yy}*(\Delta y)^2]$, but it did not provide a significantly improved fit to the data.

In the viscous substrate model, with observational constraints from only one shoreline, there is an unresolved trade-off between plate thickness and substrate viscosity. We can clearly reject substrate viscosities higher than 5 10^{20} Pa s. For lower viscosities, our observations define a curve in the two dimensional parameter space along which the residual variance is essentially constant. Representative points on that curve are: $\{(38, 0), (36, 10^{18}), (35, 10^{19}), (30, 10^{20})\}$, where locations are given in terms of plate thickness in km and substrate viscosity in Pa s.

Taken together, these observations imply that: (a) the lake surface elevation (corrected for tilt and rebound) at the high shoreline was 3750 + A = (3759.2 ± 0.7) m, (b) the computed deflection profile was essentially correct, (c) the net motions, other than lacustrine rebound, have a remarkably planar aspect, and (d) net tilt in the central altiplano over the last 16-17 kyr has been up to the east by 6.8 cm/km and down to the north by 5.3 cm/km. The magnitude of the observed tilt is comparable to our *a priori* expectations (from Salar de Uyuni salt wedge arguments) but is oriented roughly to the northwest rather than to the east.

Interpretation

The observed rebound signal is very similar to that computed from any of a suite of Earth models with elastic plate thicknesses of (40 $\pm$ 10) km and half-space viscosities less than 5 10^{20} Pa s. The history of lake level fluctuations is still inadequately known to allow detailed inferences of crust and mantle rheology. However, it is already clear that the effective elastic plate thickness is less than the 60-70 km crustal thickness in the central Andes, and therefore presumably insufficient to allow a sea level signal this far inland.

The direction of observed tilting is somewhat surprising, as the most easily anticipated axis of rotation is essentially parallel to the Peru-Chile trench (as would be expected for either ocean loading or subduction related tectonism). The sign of the E-W component of rotation (up to the east) is consistent with either an ocean loading signal (down on the oceanic side) or the observation that the primary locus of active tectonism is on the eastern margin of the Altiplano. Less clear is the origin of the N-S component. Furthermore, it is not clear that such a planar tilt signal should have been anticipated. The best fitting linear-plus-quadratic model only displays ~1 m of curvature induced deviation from the best fitting linear-only model. If the observations reflect a mixture of ocean loading (which would be strongest in the west) plus tectonism (presumably strongest in the east) this planar aspect of the tilt must represent a coincidental matching of the components. Alternatively, the lack of significant curvature implies a very strong elastic plate, which is countered by the observed response to the lake load.

A surprising result is the magnitude of the net tilt signal (almost 35 m of tilt across the extent of the basin, versus 27 m of rebound). This rate is higher than can plausibly be continued for even a million years, so if it is all tectonic, it must represent episodic tectonism, as might be expected from phase changes in the subducting slab (Cloos, 1993).

Present observations do not appear to allow a unique explanation for this tilting, but we can list and briefly discuss possible contributing processes and mechanisms. The processes can be broadly categorized as either external (climatically driven) or internal (tectono-magmatic). Climatic effects which could contribute to the observed tilting include: (a) oceanic loading accompanying sea-level rise, (b) lake loading on a laterally variable substrate, (c) asymmetric glacial or ice sheet loading, and (d) erosional unloading of the eastern flanks of the Altiplano. Tectono-magmatic effects that may have contributed include: (e) changes in the subduction process (convergence rate, buoyancy of slab) at the Peru-Chile trench, or (f) magmatic inflation of the Altiplano-Puna volcanic zone.

It is clear that resolution of these issues will be greatly benefited by further modeling results, and most especially by observations of the patterns of deflection on the lower shorelines. Although a single shoreline suffices to demonstrate the existence of multiple signals, it will require several separate views of the vertical motions to separate them according to cause. The basic lacustrine signal is easy to separate because it has a distinctive geometry. The multitude of other potential influences with geometry related to subduction will only be separable through their differing temporal signatures.

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Table 1. Lake Minchin High Shoreline

feature		observations		
	site number & name	latitude (S) dd mm ss.ss	longitude (W) dd mm ss.ss	elevation (m)
1.	Cerro Tres Cerrillos	20 33 00.79	66 48 07.85	3781.2
2.	Loma Capilla	20 28 58.42	66 30 55.96	3787.6
3.	Cerro Paco Kkollu	20 05 07.28	68 13 58.73	3774.8
4.	Cerro Campanani	20 10 32.90	67 56 57.84	3784.0
5.	Cerro Agua Castilla	20 52 36.96	67 03 47.09	3779.2
6.	Cerro Colorado	21 05 38.68	68 04 28.97	3777.6
7.	Cerro Caldami	19 37 10.07	67 45 35.91	3786.0
8.	Cerro Khasoj Saya	19 33 45.29	67 52 08.47	3782.8
9.	Cerro Maraga	19 26 45.72	67 35 49.48	3789.2
10.	Cerro Salli Kkollu	19 23 36.83	67 30 46.43	3783.3
11.	Cerro Piquina	20 13 43.06	66 54 45.27	3787.4
12.	Cerro Ochkha	18 48 59.95	68 26 01.24	3753.0
13.	Cerro Chacacheta	18 52 18.77	66 45 26.60	3768.1
14.	Cerro San Pedro	18 06 51.90	67 00 11.98	3772.3
15.	Yonza	20 37 16.33	68 01 51.57	3783.7

Figure Caption

Fig 1. Lake Minchin Survey Sites and Deflection Contours.
Solid circles are survey sites. Irregular solid line is 3800 m elevation contour, which approximates the maximum extent of Lake Minchin. Smooth lines are countours of net vertical motion, including lacustrine rebound and planar tilting. Distances are measured from an arbitrary reference point with UTM coordinates 783^{0000} N and 63^{0000} E. This corresponds to 19° 37' 19" S, 67° 45' 37" W.