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Abstract
The isostatic residual gravity field of the Central Andean Plateau region contains laterally continuous, elongated anomalies that reflect the tectonic and magmatic effects of oceanic subduction in the forearc and continental subduction in the backarc. On the western side of the Andes, the residual anomalies are similar to those found at many "Andean"-type subduction margins. In particular, a high-low residual anomaly pair coincident with the coast and the Peru-Chile trench marks the location when the Nazca plate underthrusts beneath western South America. This high-low anomaly couple is mirrored in the backarc by a similar couple that tracks the location of the eastward-vergent Principal Frontal Thrust and results from the westward, antithetic subduction of the Brazilian shield beneath the plateau. The residual high is situated in the Eastern Cordillera of Bolivia, and is caused by a combination of high-density basement rocks near the surface in the hanging wall of the thrust, and by local crustal undercompensation. To the east, the parallel low coincides with the Subandean fold-thrust belt and foreland basin, and is due to a combination of low-density sedimentary rocks in the foreland basin and the locally overcompensated crust of the downflexed foreland lithosphere. The Eastern Cordillera and the Subandean are structural analogs to the uplifted forearc and the Peru-Chile trench on the western side of the Andes. Along-strike variations in the amplitude and width of the Eastern Cordillera and Subandean anomalies indicate that the foreland lithosphere is strongest at the latitude of central Bolivia, but decreases significantly to the south where the eastern margin of the Andes appears to be nearly in a state of local isostatic compensation.

Introduction
The Central Andes are a prime example of a linear mountain belt related to subduction of an oceanic plate beneath a continental margin. The Andean topography is mostly compensated at depth by a thick crustal root that seismic studies indicate reaches a thickness of up to 70 km (James, 1971a; Schmitz, 1994; Zandt et al., 1994; Beck et al., 1996). Early reconnaissance gravity studies indicated that this thick crustal root was reflected by a negative regional Bouguer anomaly reaching a minimum of over -400 mGals (Dragicevic, 1970). Because of a sparse gravity station distribution, this data was of limited use in interpreting detailed crustal structure. Recent data acquisition in central Chile and northwest Argentina (Götzke et al., 1990; 1996), in western Bolivia (Cady and Wise, 1992), and in southern Peru (Fukao et al., 1989) has greatly improved the station coverage and this new data provides a vast amount of information for the study of crustal structure and tectonic processes.

A common interpretation approach with Bouguer gravity anomalies is to apply iterative forward or inverse modeling in order to derive a detailed 2-D or 3-D crustal density model. Often, other geophysical constraints are used to minimize nonuniqueness in the solutions. Modeling results are useful in that they set numerical constraints on density contrasts and geometries and thus may be used to interpret subsurface lithologies and structures. In most cases, however, very complex models are needed in order to fit the data. With this complexity, geologic insight is often lost.

A more heuristic approach is to view the gravity anomaly map as a generalized, vertically averaged map of crustal geology. The anomalies are then interpreted qualitatively in the context of regional-scale tectonic processes and structures. In
In addition, useful information may be derived from the maps by means of semi-quantitative, “back-of-the-envelope” calculations. Often, results of these calculations provide insights that are lost in multiparameter 3-D models.

In mountainous regions, the Bouguer anomaly map is of limited use for qualitative interpretation because it is dominated by the regional gravity field of the isostatic root. This “regional” is often removed by empirical methods such as graphic smoothing, polynomial fitting, or wavelength filtering, but these methods are deficient in that the regional is not related to any physical model. By assuming a simple isostatic model (typically a local “Airy” model), the gravitational attraction of the root may be calculated directly from the topography. The resulting anomaly is known as the isostatic residual gravity anomaly.

Although the concept of isostasy provides an important tool in interpreting regional gravity data, care must be taken when interpreting the residuals in terms of isostatic imbalances. Terms such as “undercompensation” or “overcompensation” are misleading since density anomalies in the crust that are completely compensated will still result in isostatic residual anomalies (Simpson et al., 1986; Jachens et al., 1989). Effectively, the isostatic correction acts as a high-pass filter, removing most of the long-wavelength anomalies resulting from isostatic compensation. What remains is an anomaly map that emphasizes upper-crustal geology (Jachens et al., 1989).

This paper presents an isostatic residual gravity anomaly map of the Central Andes between 13°S and 29°S, which was compiled from measurements at over 25,000 gravity stations. The map encompasses both limbs of the Bolivian Orogen, including portions of Peru, Chile, Bolivia, and Argentina, and spans the entire width of the Andean Orogen from the Peru-Chile trench to the stable craton of the Brazilian shield (Fig. 1). The results of detailed modeling of some of this data have been presented previously (Couch et al., 1981; Brussel and Wilson, 1984; Fukao et al., 1989; Strunk, 1990; Götze et al., 1994; Kirchner et al., 1996), and the reader is referred to those studies for more detailed models of the crustal structure. The purpose of this study is to discuss the regional significance of elongated anomalies with wavelengths between 100 and 500 km that are laterally continuous over 1000 km or more. These anomalies reflect lithospheric scale structures of the oceanic subduction system, the magmatic arc, and the antithetic backarc foreland thrust system.

**Geologic Setting**

The Andes mountain system is often cited as a prime example of an active orogen formed by subduction of an oceanic plate beneath a continental margin (Dewey and Bird, 1970; James, 1971b). This paper focuses on a segment of the Central Andes in southern Peru, northern Chile, Bolivia, and northwest Argentina, where the subducted Nazca plate dips at around 30° to the east, as well as the transition zones to segments of nearly horizontal subduction to the north and south.

In this segment of the Andean system, the major morphostructural elements are, from west to east—the 5 to 7 km deep Peru-Chile trench, a forearc complex, the Neogene volcanic arc of the Western Cordillera (the Central Volcanic zone), a “hinterland” complex composed of the Central Andean Plateau and the Eastern Cordillera, and a foreland thrust system composed of the Subandes Ranges of Peru and Bolivia and the Santa Barbara Ranges of northwest Argentina (Figs. 1 and 2). In northern Chile, northwest Argentina, and southern Bolivia, the Andean structures trend N-S. To the north of 18° S (the “Arica Elbow”) the coastline and Andean structures trend NW-SE. This bend in the Andes is known as the Bolivian Orocline.

In northern Chile, the forearc is composed of a Coastal Cordillera and a Longitudinal valley (Fig. 2). The modern forearc is composed of rocks corresponding to the Jurassic–Early Cretaceous magmatic arc. This reflects the steady eastward migration of the volcanic arc since the Jurassic (Coira et al., 1982; Scheuber et al., 1994). These rocks are generally basaltic to andesitic lavas and intrusives in the Coastal Cordillera and backarc sediments in the Longitudinal Valley. By the Late Cretaceous, magmatism had shifted eastward to the location of the present-day Longitudinal Valley and continued through the Eocene.

To the north of the Arica Elbow in southern Peru, Jurassic magmatism was less pronounced, because of highly oblique subduction (Jaillard et al., 1990). Instead, basement rocks of the forearc are primarily Precambrian metamorphic rocks of the Arequipa Massif (Schackleton et al., 1979).
North of 16° S, Mesozoic intrusives of the Coastal Batholith predominate along the coast with older rocks offshore (Megard, 1987).

The Western Cordillera forms the western topographic rim of the Central Andes and marks the primary locus of late Miocene and younger volcanism (Figs. 1, 2, and 3). This volcanic cover is generally composed of andesitic to dacitic lavas and dacitic-rhyolitic ignimbrites (Baker, 1981; Coira et al., 1982). Plio-Quaternary volcanism in the Central Volcanic zone is restricted to between 15° and 28° S. The volcanic gaps to the north and south are attributed to flattening of the subducted Nazca plate in these segments (Jordan et al., 1983).

The Central Andean Plateau, comprising the Peruvian and Bolivian Altiplano and the Argentine Puna, occupies a region ~300 km wide and 2000 km long with an average elevation near 4 km (Fig. 2). The Altiplano is a deformed Cretaceous and Tertiary sedimentary basin that has been uplifted to its present elevation since the late Miocene (Isacks, 1988; Gubbels et al., 1993; Lamb and Hoke, 1997). The Altiplano is essentially a single, internally drained, intermontane basin situated between the Western and Eastern cordilleras. In contrast, the Puna of northwestern Argentina is characterized by smaller and more fragmented basins, more extensive constructional volcanism, and greater local and structural relief (Jordan and Alonso, 1987; Whitman et al., 1996).

In southern Peru and Bolivia, the plateau is bounded by the Eastern Cordillera, a complexly deformed section of lower Paleozoic sedimentary rocks (Figs. 1 and 3). In places, these rocks are covered by large Tertiary ignimbrite sheets and shallow Tertiary basins. Farther south in northwestern Argentina, the distinction between the plateau (the Puna) and Eastern Cordillera is somewhat ambiguous.
The Subandes of Peru and Bolivia and the Santa Barbara Ranges of northwest Argentina compose the foreland fold-thrust system of the Andes. In the Subandes, laterally continuous E- and NE-verging folds and W- and SW-dipping thrusts deform an eastward-thinning Paleozoic and younger sedimentary wedge. The style of deformation is largely thin skinned and confined to the overlying sedimentary package (Mingramm et al., 1979; Roeder, 1988; Baby et al., 1989). Farther south, the Santa Barbara system is characterized by asymmetric W-verging folds and high-angle, E- and W-dipping reverse faults that expose Mesozoic and older strata. These structures are less laterally continuous and exhibit a greater degree of basement involvement than do the Subandean folds and thrusts to the north (Mingramm et al., 1979; Allmendinger et al., 1983; Jordan et al., 1983; Grier et al., 1992). The tectonic style of this segment is transitional between the thin-skinned tectonics of the Subandean belt to the north and the thick-skinned Pampenan ranges to the south (Jordan et al., 1983).

Preparation of the Isostatic Residual Map

Data

The data used in this study consist of Bouguer gravity anomaly values that were compiled from three sources (Fig. 4). The largest source of data is from the U.S. Defense Mapping Agency (DMA) file, which contains over 28,000 points between 12° S and 35° S. This data set was compiled by the DMA from numerous sources and includes over 13,000 offshore stations and 15,000 onshore stations. The onshore data generally consist of densely spaced stations (1 to 5 km) situated along widely spaced roads (50 to 200 km). To the south of 22° S, the DMA station distribution is very
sparse, except along the coast and offshore. Between 19° S and 28° S, the DMA data were supplemented by gravity values at 5700 stations in northern Chile, northwestern Argentina, and southern Bolivia, which were collected and compiled by the gravity research group at the Freie Universität, Berlin (Götze et al., 1990, 1996; Kösters, et al., 1997). The station distribution for this data set is considerably better than for the DMA data, with better than 5 km station spacing along lines generally spaced 50 km or less apart. The final source of data consists of 450 stations situated on four widely spaced profiles in southern Peru (Fukao et al., 1989).

The three data sets were merged by the following procedure. First, all data were adjusted, if necessary, to be consistent with the 1967 gravity formula (International, 1967). In order to provide a consistent data set, a Bouguer correction was re-calculated for all onshore data using a reduction density of 2670 kg/m³. For offshore data, a Bouguer correction for the water layer was calculated from reported bathymetric soundings using a reduction density of 1030 kg/m³. Offshore data without reported bathymetric soundings were deleted from the data set. In regions of overlap between the different data sets, only the newer data were used. The DMA data set required considerable additional editing. Several stations on the Altiplano had anomalies that differed by as much as 100 mGals from stations on nearby profiles. In general, these probably erroneous data were situated at isolated stations within the Altiplano, and all stations on the Altiplano not on continuous gravity profiles were deleted. After the initial merging and editing, additional isolated, probably erroneous, gravity values that appeared as short-wavelength anomalies or "bulls eyes" on
preliminary Bouguer and isostatic residual anomaly maps were identified and removed from the data set. The final data set consisted of over 25,000 stations between 12° S and 35° S. To the south of 29°, the station coverage is sparse and only data north of this latitude are presented in this study.

All stations in the Berlin data set included terrain corrections. These corrections consisted of a far field term calculated for distances of 5 to 100 km from a 3′ digital elevation model, and a near field term calculated from topographic maps and station surveying information (see Götze et al., 1990, 1996 for details). For all other data, terrain corrections were calculated from a 30-arc-second digital elevation model (USGS, 1997) for distances of 5 to 100 km from each station by using a line mass approximation. Generally the terrain corrections were less than 2 mGal. However, in extreme cases, the correction was over 50 mGal. Inner-zone corrections for distances of less than 5 km were not calculated because of the limitations of the topographic grid. For stations in regions of extreme topography, potentially large additional terrain corrections exist for this inner zone. These stations, however, correspond to a relatively small percentage of the entire data set and insufficient terrain corrections probably are reflected by short-wavelength perturbation on the larger-scale regional anomalies.

Isostatic reduction

The isostatic residual gravity anomaly was calculated by subtracting a calculated isostatic regional from the observed Bouguer gravity values. First, an Airy-Heiskanen local compensation model was used to calculate an isostatic root from the topography. In the Airy-Heiskanen model, the topography within a vertical column is compensated at depth by a crustal root of equal mass. On land, the topographic load results in a crustal root of thickness, \( w(x,y) \), given by:

\[
  w(x,y) = -\frac{\rho_i}{\Delta \rho} h(x,y),
\]

while offshore, the mass deficiency of the water column results in a crustal anti root,
\[ w(x,y) = \frac{\rho_t - \rho_w}{\Delta \rho} h(x,y), \]  
where, \( h(x,y) \) is the topography or bathymetry, \( \rho_t \) is the topographic density, \( \rho_w \) is the water density, and \( \Delta \rho \) is the density contrast across the bottom of the root.

The isostatic parameters used in this study are shown in Table 1. The topographic density \( \rho_t \) was chosen to be consistent with the Bouguer reduction density, and the density contrast \( \Delta \rho \) and zero elevation crustal thickness \( z_c \) were selected to produce Moho depths consistent with the crustal thickness determined from analysis of wide-angle Moho reflections and seismic travel-time residuals observed near Jujuy, Argentina (Cahill et al., 1992; Whitman, 1994). These parameters predict maximum crustal thicknesses in the high Andes of 65 to 70 km, consistent with seismological studies (James, 1971a; Wigger et al., 1994; Zandt et al., 1994; Beck et al., 1996), and are probably reasonable assumptions for isostatic corrections in other parts of South America (Ussami et al., 1993). In any event, the calculated isostatic gravity regional is relatively insensitive to these assumed parameters (Simpson et al., 1986) and the results of this study differ by only a few percent from the isostatic residual map of Götze et al. (1990), which used values for \( \Delta \rho \) and \( z_c \) of 500 kg/m\(^3\) and 40 km, respectively.

The gravitational attraction of the root was calculated by using a modified version of the program AIRYROOT (Simpson et al., 1983), which employs a Fourier transform expansion developed by Parker (1972). Following Parker, (1972), the Fourier transform of the vertical attraction of the root, \( F(\Delta g_o) \), is given by

\[ F(\Delta g_o(x,y)) = -2\pi G \Delta \rho \exp(-k |z_c|) \sum_{n=1}^{\infty} \frac{k^{n-1}}{n!} F(w^n(x,y)), \]  
where \( F \) denotes the Fourier transform, \( G \) is the gravitational constant, \( k \) is the magnitude of the wavenumber vector: \( k = k_x + k_y \), and \( z_c \) is reference crustal thickness at sea level.

The isostatic root was calculated by application of equations 1 and 2 using values from a 3' topographic and bathymetric data set (Isacks, 1988) subaveraged into 10 by 10 km cells. The topographic grid was extended at least 1000 km away from the boundaries of the study area to avoid edge effects in the gravity calculations. Additional padding and tapering was applied to the calculated root grid and the 2-D Fourier transform was calculated. The regional gravity field was then calculated from the inverse Fourier transform of the first five terms of (3). Equation 3 assumes a flat earth, and no additional corrections for earth curvature were made. Direct calculations of the isostatic regionals for curved and flat-earth models along selected profiles indicate, however, that neglecting earth curvature results in a long-wavelength error of no more than 4 mGal.

The calculated isostatic regional value was then determined at each station by bicubic interpolation from the calculated grid. Since the regional anomaly determined by equation 3 is calculated at a constant datum level (usually sea level), an additional correction was required to adjust the regional field to station level. This correction is especially important in the Central Andes, where the range of station elevations is over 5000 meters, and if neglected, errors of over 10 mGal may result. As a first approximation, the regional at station level, \( \Delta g_o \), is given by:

\[ \Delta g_t = \Delta g_o + t \frac{\partial \Delta g_o}{\partial z}. \]  
where \( \Delta g_o \) is the regional calculated at sea level and \( t \) is the station elevation. The vertical derivative term in (4) is related to the regional field calculated on the x-y grid in (3) by:

\[ \frac{\partial \Delta g_o(x,y)}{\partial z} = F^{-1}[-kF(\Delta g_o(x,y))], \]  
where \( F \) and \( F^{-1} \) denote the forward and inverse Fourier transforms, respectively. The derivative field is then interpolated to each station location and equation 4 is applied. Subtracting the calculated regional from the Bouguer anomaly at each station results in the isostatic residual gravity anomaly at station level. An additional downward field continuation of the anomaly could have been

<table>
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<th>Table 1. Isostatic Parameters</th>
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<tr>
<td>( \rho_t )</td>
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applied in order to reduce all the data to a sealevel datum. In general, this correction requires a
detailed knowledge of the mass distribution be­tween station level and datum level, which gener­ally is not available (Götze and Li, 1996). In order
to avoid contaminating the data, no additional po­
tential field continuations were applied to the
data. Care must be taken to model or interpret this
anomaly at station level.

Two additional corrections were applied to the
gravity data. Since the free-air gravity correction
is typically reduced to the reference ellipsoid
rather than the geoid (station level), long­wavelength biases in the gravity data can occur in
regions of significant geoid relief. This so-called
“indirect effect” (Chapman and Bodine, 1979;
Götze and Li, 1996) amounts to 0.20 mGal/m on
land and 0.27 mGal/m offshore, and must be
added to the gravity anomaly. In the area of this
study, the geoid ranges from 20 to 45 m of relief,
resulting in corrections of as much as 3 mGal. Fi­
nally, after all the isostatic corrections were ap­
plied, the residual field still contained a static off­
set of −20 mGal. Götze et al. (1991b) have
similarly noted that the mean isostatic residual
anomaly in northern Chile and northwest Argent­
a is positive by around 15 mGal, and they
interpret this result to indicate that the Andean
lithosphere is regionally undercompensated. Al­
ternatively, we suggest that this long­wavelength
anomaly reflects subduction­related, uncompen­sated mass in the mantle, and is associated with
degrees 4 through 10 of the spherical harmonic
decomposition of the earth’s gravity field (e.g.,
Bowin, 1985). In order to remove this effect, con­
tributions through degree 10 of the GEM-T1 geo­
potential model (Fig. 5) (Marsh et al., 1990) were
subtracted from the isostatic residual anomaly
values.

**Gridding and interpolation**

To produce the map in Figure 6, the isostatic
residual anomaly station values were interpolated
into a regularly spaced grid. First, station data
were projected into x-y coordinates using a trans­
verse Mercator map projection with a central me­
ridian of 67° W. The data then were averaged into
a set of 10 by 10 km grid cells in order to avoid spatial aliasing and to suppress outliers. Cells
that contained no gravity stations were given a
value of “no data.” Finally, cell values with no
data were calculated by a 2-D tensioned spline
algorithm (ESRI, 1994). This gridding process re­sulted in a 300 x 180 cell grid with 10 km² cells.

**Structural Interpretation**

A map of the isostatic residual gravity field is
shown in Figure 6. The most prominent features
in the residual field are laterally continuous, elon­gated anomalies that are generally parallel to
the strike of the major tectonic structures (Fig. 6).
These anomalies are 200 to 300 km wide and
1000 km or more long. The sections below discuss
the regional significance of these elongated
anomalies related to subduction of the Nazca
plate beneath western South America, late Ceno­
zoic arc magmatism in the Central Andes, and
backarc compressional deformation to the east of
the Andes.

**Anomalies related to subduction of the
Nazca Plate**

Large-amplitude free-air and isostatic residual
gravity anomalies are ubiquitous features of most
oceanic convergent margins, and as a first ap­
proximation, indicate that the bathymetry of oce­
anic trench systems departs significantly from a
state of local isostatic equilibrium (Vening-
Meinesz, 1964; Watts and Talwani, 1974; 1975;
Bowin et al., 1982). In many respects, the iso­
static residual anomaly is superior to the free-air
anomaly for analyzing the structures of continen­tal margins because it accounts for the gravita­
tional effect of the water layer as well as the
“edge” effect resulting from the rapid increase in
crustal thickness beneath the continent.

Adjacent to the coast of western South
America, the isostatic residual gravity field is
characterized by three parallel anomalies (Fig. 6).
A 75 to 100 mGal high is situated around 100 km
seaward of the Peru-Chile trench (OTH in Figs. 6
and 7). This “outer trench high” is analogous to
the positive free-air anomalies observed seaward
of many deep-sea trenches (e.g., Watts and Tal­
wani, 1974) and may be explained in terms of
simple flexural plate models as the upward dis­
placement of the oceanic Moho within the flexural
forebulge of the Nazca plate as it approaches the
trench. Similarly, a −100 mGal low coincident
with the axes of the Peru Chile trench (TL in Figs.
6 and 7) reflects the mass deficiency caused by
the downward bending of the Nazca plate into the
trench, possibly combined with a downward dis-
placement of the trenchward margin of the South American plate caused by coupling of the upper and lower plates (e.g., Wdowinski, 1992). The trench bathymetry is not locally compensated by a crustal anti-root, since the Moho is embedded within the deflected plate.

The source of a 100 mGal residual high that runs along the coast is more problematic (CH in Figs. 6 and 7). Recent work in northern Chile has associated this anomaly with high-density, high-seismic velocity intrusive rocks in the coastal cordillera, corresponding to the Jurassic volcanic arc (Götze et al., 1991a; 1994; Götze and Kirchner, 1997). Farther north, the rocks along the coast change markedly. In southern Peru, the high is coincident with the Precambrian metamorphic rocks of the Arequipa Massif. In central and northern Peru, the high coincides with the Mesozoic granitoids of the Coastal Batholith (Brussel and Wilson, 1984).

Although high-density rocks make a considerable contribution to the positive anomaly, their geodynamic setting cannot be ignored. The coastal high is a remarkably continuous along-strike feature of the subduction margin of western South America. Similar "inner-arc highs" are present along many other subduction margins around the world (Vening-Meinesz, 1965; Watts and Talwani, 1974; Simpson et al., 1986; Abers, 1994). We therefore suggest that the coastal gravity high is a fundamental feature of subduction margins.

In the eastern Aleutians, Abers (1994) recently related the inner-arc high to the uplift of high-density lower-crustal rocks along the coast caused by the mechanical support of the upper plate by the subducting plate. We suggest a similar mechanism for the inner-arc high along the coast of western South America. Along the coast, the subducting Nazca plate flexurally supports the overriding South American plate, such that the coastal topography is locally undercompensated. The high-density and high-seismic velocity rocks observed along the coast of northern
Chile (Wigger et al., 1991, 1994; Götze et al., 1991b, 1994) and the Precambrian rocks of the Arequipa Massif in southern Peru reflect the uplift of lower-crustal rocks to higher crustal levels due to flexural support of the subducting Nazca plate. The coastal high is not simply the result of a specific localization of high-density intrusives in this region. In the context of isostasy, mechanical coupling of the upper and lower plates is probably required to support the mass anomalies associated with these high-density rocks. This model can be applied equally well to the coastal isostatic residual high in the northwestern United States, which usually is attributed to the accretion of high-density oceanic terranes (e.g., Simpson et al., 1986).

High-density rocks associated with the subducted Nazca plate also may contribute to the coastal high. On the basis of thermal and petrological models, Grow and Bowin (1975) estimated the density structure and gravitational contribution of the slab. Kirchner et al. (1996) and Götze and Kirchner (1997) have shown that the gravity contribution of the slab is, at most, 60 mGal near the coast. This effect decreases inland and is partially masked by the overlying asthenospheric wedge.

As previously mentioned, the residual field contained a long-wavelength static offset of ~20 mGal, which was removed by subtracting contri-
FIG. 7. Profiles across the Altiplano and Puna segments of the Central Andes showing averaged topography and point values of the isostatic residual gravity anomaly projected into sections along 100 km wide swaths. Location of profiles is shown in Figure 4. Abbreviations: PFT = Principal Frontal thrust; other symbols are presented in Figure 6.

butions through degree 10 of the earth’s satellite-derived gravity field (Fig. 5). Bowin (1985) has associated these harmonics with convergence zones. In effect, these harmonics are a first-order approximation of the gravity contribution of the subducted plate. Subtracting this contribution from the gravity field can be viewed as an alternative to modeling the density structure of the subducted plate.

Residual lows related to the volcanic arc

Felsic volcanic piles and their subvolcanic intrusives are commonly associated with gravity lows (Kane and Godson, 1985; Simpson et al., 1986). In the Central Andes, two linear belts of anomaly lows correlate with the location of late Cenozoic extrusives of the Central Volcanic zone (Fig. 8). In the Western Cordillera, a -50 to -100 mGal anomaly low is coincident with the main Neogene volcanic arc (WCL in Figs. 6 and 7). This low is most prominent between 21° S and 24° S, coinciding with the Altiplano-Puna volcanic complex, a region of voluminous late Miocene and younger caldera complexes and ignimbrite sheets (de Silva, 1989; de Silva and Francis, 1991). In the Eastern Cordillera of Bolivia, a -25 to -40 mGal low (ECL in Figs. 6 and 7) follows a general trend connecting the late Miocene Morococa and Frailes ignimbrite plateaus in the north with the Cerro Panizos ignimbrite shield on the northern margin of the Altiplano-Puna volcanic complex. This anomaly is a continuous feature coincident with the “inner arc” along the eastern margin of the Altiplano, and may mark the location of a low-density intrusive body in the subsurface.

Anomalies related to underthrusting of the Brazilian Shield beneath the plateau

Similar to oceanic subduction margins, gravity anomalies and the geometry of foreland basins adjacent to mountain ranges demonstrate that mountain ranges and their adjacent foreland basins often are not in local isostatic equilibrium (Beaumont et al., 1982; Karner and Watts, 1983; Molnar and Lyon-Caen, 1988). At the foreland thrust margin of mountain ranges, this is reflected by large, positive isostatic residual anomalies across the high topography, and large negative residual anomalies across the adjacent foreland. As with the oceanic case, these gravity anomalies are most simply explained in terms of flexure of a loaded elastic plate underlain by a fluid substratum. In this context, the foreland lithosphere an-
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**Fig. 8.** Map showing the relationship of the isostatic residual anomaly field to the distribution of late Cenozoic volcanic centers (black diamonds) (after Isacks, 1988) and the shape of the subducted Nazca plate (labeled contours after Cahill and Isacks, 1992). The recent volcanic rocks generally occur over residual lows. The general trend of the gravity anomaly and the shape of the Nazca plate are uncorrelated except along the coast.

tithetically subducts beneath the mountain belt, a process known as “A-type” subduction (e.g., Bally, 1981). The strength of the underthrust lithosphere regionally supports the thrust load such that the high topography is underlain by excess mass, whereas the adjacent foreland basin is underlain by a deficit in mass.

In the Central Andes, this mechanism is best demonstrated at the latitude of southern Bolivia by parallel anomalies in the Eastern Cordillera and Subandes (ECH, SAL, and BSH in Figs. 6 and 7B). These anomalies follow the general trend of the major westward-dipping thrust faults that bound the eastern margin of the Andes, and may be viewed as the continental analogs of the coastal high, trench low, and outer trench high that were discussed in the previous section. In the Eastern Cordillera of Bolivia, a 25 to 75 mGal residual high (ECH) coincides with highly deformed lower Paleozoic rocks in the hanging wall block of a crustal-scale, two-sided thrust wedge, and is caused by a combination of high-density basement rocks in the hanging wall block and local undercompensation of the topography. To the east, a ~25 to ~75 mGal residual low (SAL) reflects both the presence of low-density sedimentary rocks in the Subandean fold-thrust belt and foreland basin, and a locally depressed Moho caused by the downflexed foreland lithosphere. A residual high in northeastern Bolivia (BSH) coincides with the outcrop of Precambrian basement rocks on the Brazilian shield (Fig. 3). Our interpretation of this anomaly is limited by poor station coverage, but a comparison with the isostatic residual anomaly map of Brazil (Ussami et al., 1993) suggests that this high is part of a larger-scale, broad residual high, possibly marking the flexural forebulge of the Brazilian shield.
To aid in the interpretation of the gravity map, a generalized, regional-scale map of the major late Cenozoic structures on the eastern half of the plateau and in the foreland was constructed (Fig. 1). Because of the extensive volcanic cover, no attempt was made to map structures in the Western Cordillera or on the western side of the plateau. In southern Peru and northern Bolivia (12° to 17° S), faults were identified from stratigraphic relationships on existing 1:1,000,000-scale geological maps (Instituto, 1975; Pareja et al., 1978) and generalized from structures identified on the map of Sempere et al. (1990). In southern Bolivia and northernmost Argentina, structures were interpreted from a 1:1,000,000-scale LANDSAT TM and MSS photomosaic. In northernwestern Argentina, faults were selected from a 1:1,000,000 unpublished compilation (R. W. Allmendinger, unpubl., 1981) and from MSS prints.

Major E- and W-dipping crustal-scale faults bound the Eastern Cordillera residual high (Fig. 9). In northern Bolivia, the ECH is bounded to the southwest by the Cordillera Real, Coniri, and Arque-Toracari backthrust systems (CRT and CT in Figs. 1 and 9), which place lower Paleozoic rocks exposed in the Eastern Cordillera over younger rocks in the Altiplano basin (Pareja et al., 1978; Roeder, 1988; Sempere et al., 1988; Sempere et al., 1990). South of the bend in the Bolivian orocline, the ECH is bounded on the west by a major N-S-trending lineament (PL in Figs. 1 and 9). The exact nature of this structure is unknown, but it marks the eastern topographic limit of the Altiplano. The ECH is bounded to the west and northeast by the westward-dipping Principal Frontal Thrust (PFT in Figs. 1 and 9). This fault and a related structure, the Main Andean Thrust (MAT), form a crustal-scale, E- and NE-verging thrust system that marks the western limit of the Subandean fold thrust belt, and juxtaposes highly deformed, lower Paleozoic rocks over upper Paleozoic and younger rocks (Roeder, 1988; Sempere et al., 1988; 1990; Baby et al., 1989; Gubbels et al., 1993).

The ECH is broadest and of greatest amplitude in the elbow region of the Bolivian Orocline between 17° and 20° S. This region is situated where the Andes are the widest and where the greatest amount of shortening is thought to have occurred (e.g., Isacks, 1983). The maximum in the ECH may then mark the area where the largest amount of vertical mass transport on the bounding Andean thrusts has occurred due to the greater amount of shortening. Alternatively, the excess mass associated with this high may be the result of lateral mass transport toward the elbow. In the elbow region of the Andes, the deformation changes from sinistral strike-slip and northeastward thrusting north of the bend, to dextral strike-slip and eastward thrusting south of the bend. This change in kinematics is a geometric consequence of the change in the relative convergence direction across the oroclinal bend of the Andes (Dewey and Lamb, 1992) and results in a net transport of mass in the elbow region toward the northeast.

The axis of the Subandean residual low generally tracks the location of the Subandean deformation front (DF in Figs. 1 and 9), which separates deformed upper Paleozoic through Tertiary sedimentary rocks within the fold-thrust belt from undeformed Tertiary sediments in the eastward-tapering Subandean foreland basin. In Bolivia, this basin forms a classical downflexed foreland trough, and is a continuous along-strike feature for more than 1500 km. From southern Bolivia to northwestern Argentina, the low progressively decreases in both magnitude and width, and south of 26° S no distinct low is present. This lateral variation in the Subandean low coincides with along-strike changes in the Subandean foreland basin that show a progressive southward narrowing and disappearance of the basin into northwest Argentina (Mingramm et al., 1979; Russo and Serriatto, 1979; Jordan and Alonso, 1987).

Regional Compensation of the Foreland Lithosphere

Gravity anomalies across the margins of mountain ranges often are interpreted in terms of the flexure model of isostasy to the gravity anomalies on the eastern margin of the Andes (e.g., Karner and Watts, 1983; Molnar and Lyon-Caen, 1988). In this model, topographic and thrust loads are compensated regionally by a foreland lithosphere of finite strength. Typically, this strength is expressed in terms of an equivalent elastic plate of a given thickness. This “effective elastic thickness” is a convenient measure for comparing the compliance of the lithosphere in different regions (e.g., McNutt et al., 1988).

Numerous recent studies have applied this
Fig. 9. Map showing relationship of the isostatic residual anomaly field to the location of major structures on the eastern half of the plateau and in the foreland thrust belt. Major E- and W-dipping crustal-scale faults bound the Eastern Cordillera residual high. The Subandean low tracks the location of the deformation front.

model to the Central Andes (Lyon-Caen et al., 1985; Roeder, 1988; Whitman, 1994; Watts et al., 1995; Fan et al., 1996; Stewart and Watts, 1997). These studies generally show that the effective elastic thickness of the Andean foreland is greatest in the bend region of Bolivia and decreases to the north and south. These regional variations in strength of the foreland lithosphere are reflected by systemic changes in the ECH/SAL isostatic residual anomaly pair. This high-low isostatic residual anomaly pair is best expressed near 18° S in the elbow region of the Andes, where the width is nearly 400 km (Fig. 6). Both to the north and south of the bend, the anomaly pair decreases in width and amplitude. To the south of 23° S in northwestern Argentina, both the amplitude and the width decrease markedly (Figs. 6 and 7C). By 25° S the anomaly pair is not well expressed and the isostatic residual field is dominated by short-wavelength anomalies that are related to local geological structures (Götte et al., 1990, 1994). At this latitude, the eastern margin of the Andes is very nearly in a state of local isostatic compensation.

The effective elastic plate thickness at continental thrust belts is controlled by a combination of factors, including crustal thickness, the overall lithospheric thickness and thermal structure, the dip and curvature of the underthrust plate, and mineralogy of the mantle lithosphere (Kusznir and Karner, 1985; McNutt et al., 1988). The effective elastic thickness of the foreland is greatest near the bend region where the Andes are closest to the Brazilian shield (Watts et al., 1995; Stewart and Watts, 1997). The north-to-south change from regional to local isostatic compensation of the eastern margin of the plateau is consistent with the abrupt decrease in mantle lid thickness.
inferred from observed patterns of upper-mantle seismic wave attenuation (Q) beneath the plateau and foreland (Whitman et al., 1992). Thermomechanical modeling studies indicate that for relatively thin, thermally young lithosphere, the elastic thickness is dominated by the relatively weak quartzofeldspathic rheology of the crust, while for thick, thermally mature lithosphere, the elastic thickness is largely controlled by the olivine rheology of the mantle (Kusznir and Karner, 1985). This southward decrease in the thickness and strength of the foreland lithosphere may reflect a change from a stable cratonic lithosphere of the Brazilian shield in Bolivia to a weaker, thermally mobilized lithosphere in northwestern Argentina.

Conclusions

In both the forearc and in the backarc, the isostatic residual gravity anomaly of the Central Andes contains parallel elongated anomalies that are characteristic of other subduction margins. The relationship of these anomalies to lithospheric-scale structures is best demonstrated near 20° S (Fig. 10). Because of the depth and their small density contrasts, these structures contribute relatively little to the gravity anomalies. Instead, the anomalies are largely caused by upper-crustal rocks. Their presence, however, is indirectly related to deeper structures in the lithosphere, and thus these anomalies are useful in understanding the relationships between surface geology and plate tectonic processes.

Along the coast a high-low anomaly pair coincides with the uplifted continental margin and oceanic trench typical of “Andean-type” margins. In the backarc, a similar anomaly pair coincides with the overthrust eastern plateau margin and its adjacent downflexed foreland basin. These structures are the direct continental analogs of the oceanic subduction-related structures in the forearc. Gravity lows on the western and eastern margins of the plateau are both associated with the magmatic arcs. In the Western Cordillera, the active...
volcanic arc and its coincident isostatic residual low is clearly related to the eastward subduction of the Nazca plate. By analogy, the parallel late Miocene inner arc in the Eastern Cordillera of Bolivia and its associated isostatic residual low reflect westward subduction of the mantle portion of the Brazilian shield (Fig. 10).

The isostatic residual anomalies corresponding to the trench, forearc, and main magmatic arc are relatively continuous features along the western margin of the Andes. This probably results from the relatively constant dip of the subducted Nazca plate along the inner-plate contact zone in this region. In contrast, the elongated anomalies along the eastern margin of the Andes show marked along-strike variations (Fig. 6). These variations may reflect along-strike variations in the thrust loads of the upper plate and the flexural rigidity of the lower plate.

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