**Large-scale geomorphology in the central Andes of Peru and Bolivia: Relation to tectonic, magmatic and climatic processes**

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## ABSTRACT

Andean topography directly reflects tectonic processes that result from ocean-continent convergence. High topography is supported by a deep thick crustal root which formed due to shortening of a previously extended crust, and also by magmatic addition from below as a result of asthenospheric melting induced by dewatering of the subducting slab. Sedimentological and fission track data has been used to reconstruct evolving Andean palaeogeography since the late Cretaceous, when most of the region was at or near sea-level. Palaeoaltitude has been estimated using mainly morphology of erosion surfaces, some of which are covered by marine limestones. Palaeoflora data is sparse and may indicate climate change as much as uplift. Prior to ca. 25 Ma the data suggest the Andes were probably no more than 1000-2000 m high, with significant areas still near sea-level. Major uplift since then reflects crustal shortening (90%) and magmatic addition from below (10%) and coincides with an increase in rate of subduction. The earliest uplifts inverted preexisting Cretaceous depocentres or terrane boundaries and these still dominate the geography of the northern Andes. In the central Andean plateau region these early uplifts merged to produce a ca. 4000 m high plateau, up to 500 km wide. Major shortening in this region may have been favoured by a strong lithosphere which mechanically favoured the development of a high-shortening thin-skinned foreland thrust belt with low-angel basal décollement. Elsewhere low-shortening basement uplifts dominate. In the central Andes, erosion rates are very low, except on the eastern flank. Although surface volcanism is spectacular it contributes little to crustal volume and average elevation. Dissection in much of the region is very recent, results in short-wavelength, narrow gorges, and probably reflects late Cenozoic climate change rather than rapid recent uplift.

## INTRODUCTION

The Andes (fig. 1) stretch 9000 km along the western edge of South America, are nearly 7000 m high at their highest and 700 km wide at their widest. Pacific oceanic crust is being actively subducted beneath the western margin of the continent and is driving active volcanism and crustal deformation. Along the orogen there are striking changes in structure, geomorphology, kinematics of active deformation, distribution of active volcanism and climate making the region an ideal natural laboratory to study the numerous processes which form a ocean-continent convergent margin orogen. There is a considerable literature on Andean stratigraphy, structure and magmatism in the Andes, but relatively little on geomorphology, and there are almost no published papers using modern digital elevation models to systematically analyse topography. Zeil (1979) amongst others has provided a broad regional review of the bedrock geology while Clapperton (1993) has reviewed South American Quaternary geomorphology. In this paper I will attempt to describe the morphology of the Andes at a very large scale, review the evidence which we can use to infer the location and timing of Andean uplift and discuss the relationship of this uplift to plate tectonic processes. Although I have included some material from the throughout the Andes I have concentrated on the central Andes because more relevant evidence seems concentrated in this region. The bias in part reflects my own field work in the region since 1990.

## RECONSTRUCTING ANDEAN UPLIFT: METHODS

The uplift and erosion history of the Andes is recorded in forearc, intramontane and foreland basin sediments. Provenance and palaeoflow data can be integrated with radiometric and palaeontological ages to map out uplifting and actively eroding regions. Because much Andean sediment is continental only a low resolution chronostratigraphy is available, from dated volcanic intercalations (e.g. Kennan et al., 1995) and a few mammal faunas (e.g. Gayet et al., 1991; Marshall and Sémpere, 1991).

Palaeogeographic reconstructions in this paper are based only on reliably dated sequences (e.g. see Lamb et al., in press; Kennan et al., 1995). In may areas no direct dating is available and there is a danger that superficially similar red beds have often been correlated using the invalid assumption that bounding unconformities are time surfaces traceable over vast distances (e.g. Sébrier, 1988). In reality, both erosion and sedimentation may have been quasi-continuous in many areas. Seismic sections (unpublished oil company data) show widespread synsedimentary thrusting, with associated localised intraformational unconformities.

Fission track ages in apatites and zircons (see Andriessen, 1995 for a recent review) are used to infer the timing and rate of erosional denudation. Denudation rate can be calculated using either samples collected over a wide altitude range or assuming a reasonable geothermal gradient (typically about 25°C/km in the central Andes. e.g. Henry and Pollack, 1988). Although the term "uplift" is usually used when discussing the results, the data only give bulk erosion rates, or movement of a rock body towards the eroding topographic surface. They say nothing about surface uplift or rate of uplift with respect to a fixed datum such as sea-level (see England and Molnar, 1990 for a discussion of the misuse of "uplift" and MacFadden et al., 1994 for a misuse of denudation as surface uplift in the Andes). While there is a general coincidence in the Andes between greatest relief and highest inferred denudation rates, the data cannot be used to infer palaeoaltitude or the rate at which average orogen altitude was increasing. Independent, palaeobotanical data from other mountain belts (e.g. Colorado plateau, Gregory and Chase; 1992) clearly indicate that surface uplift and erosion may be separated by more than 30 Ma. However, because fission track data record the time and rate of erosion, we can infer a minimum surface uplift age and minimum age of clast influx into surrounding basins where there may be no datable tuffs or fossils. Thus the data can be integrated into palaeogeographic reconstructions.

Palaeosurfaces due to marine or fluvial erosion are widespread, especially in the central Andes, and may be used to infer changes in surface altitude. While the presence of

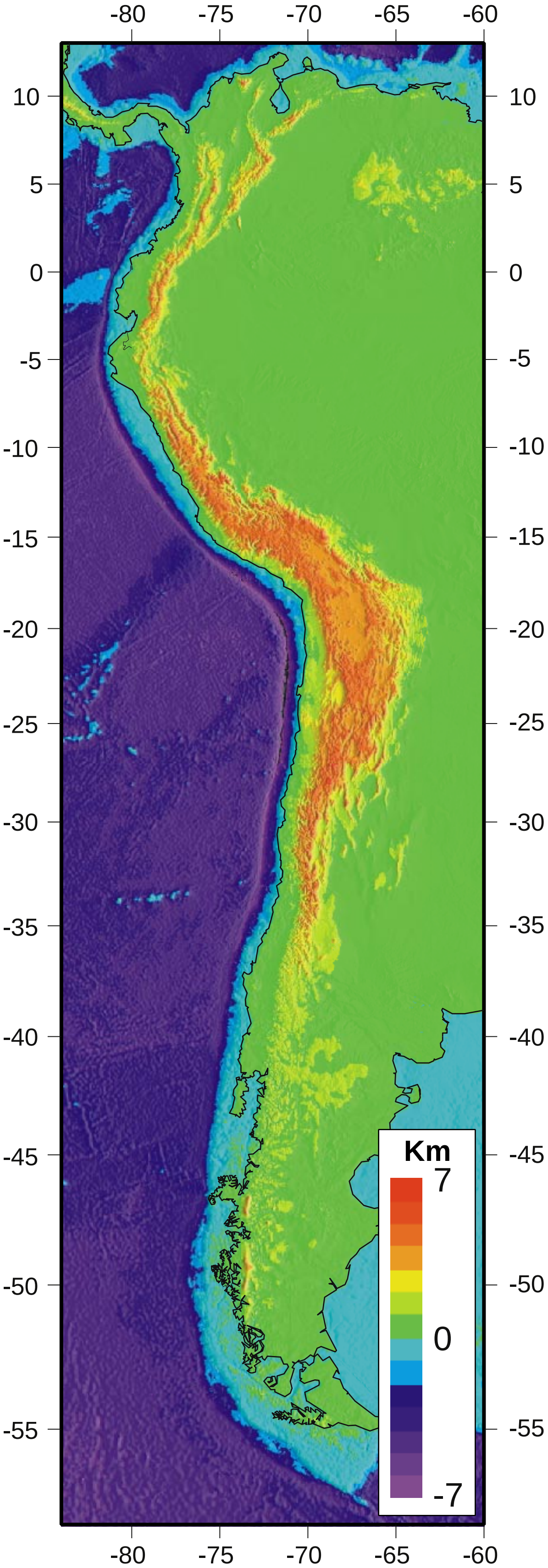


**Fig. 1**. Colour shaded relief map of western South America, illuminated from the east, generated from 3-minute topography data. The pronounced topographic low off the western coast is the Peru-Chile trench, where Pacific oceanic crust is subducting beneath the continental margin. The most prominent feature of the Andes are the “Central Andean Plateau” between 10°S and 30°S, which changes trend abruptly at ca. 19°S. Outlying uplifts are clear to the east of the main Andean range as the plateau tapers to north and south. The quite distinct character of the northern Andes, where high cordilleras are separated by broad, low altitude valley, and the generally low relief of the southern Andes are also apparent.

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marine rocks gives a direct indication of uplift, palaeoaltitude can only be reconstructed from fluvial terraces or pediments if drainage gradient and base-level altitude are known. For instance, in the central Andes of Peru and Bolivia, recent rapid uplift was inferred from the dissection of erosion surfaces now at ca. 4000 m (e.g. Walker, 1949). However, base-level for these surfaces was probably the perched, internally draining Altiplano (author's own observations) which could have been well-above sea-level. I have only quantified uplift of palaeosurfaces where base-level can be clearly inferred (examples discussed below). Molnar and England (1990) have suggested that uplift of such palaeosurfaces could in part be due to isostatic rebound during or following deep dissection. However, Gilchrist et al. (1991), giving an example from the Alps, suggest that in real orogens this effect can only account for a small fraction of the apparent uplift and it cannot account at all for the considerable surface uplift of undissected or internally draining regions such as the Andean Altiplano or Tibetan Plateau.

Surface uplift of mountain belts is commonly inferred from changes in flora. Plants are sensitive to temperature and because temperature varies with altitude palaeofloras in mountain belts can, in principle, be used to deduce the altitude at which they grew. Palaeoflora species or communities can be compared with nearest living relatives, although there is the possibility of adaptation to a new environment. Alternatively, potentially temperature-dependant leaf characteristics such as smoothness of margin may be used to infer palaeotemperatures (e.g. see Forest et al., 1995; Gregory and Chase., 1992). Unfortunately, palaeoaltitude can only be inferred from palaeotemperatures if the change of temperature with altitude at the time the plants were deposited is known.

In many orogens, there is evidence of late Cenozoic increase in denudation rates (see Ruddiman et al., 1989). Deep Pliocene dissection is seen in the Himalaya, in the Colorado Plateau and in certain parts of the Andes (e.g. see Kennan et al., 1995) and many authors have used it to infer Plio-Pleistocene uplift of 1500-4000 m (e.g. Servant et al., 1989; Walker, 1949; Harrison, 1943). However, although both dissection and floral changes could result from surface uplift it is possible that climate change involving cooling and an increase in precipitation could also be responsible (see Molnar and England, 1990). The middle Pliocene (ca. 3 Ma) was apparently a time of major climate change; oxygen isotope trends around the world show a dramatic decrease in d18 O and the first northern hemisphere sea ice is noted (e.g. Raymo and Ruddiman, 1992). Closer to the Andes, Pliocene changes in d18 O in eastern Pacific sediments (Shackleton and Hall, 1995) indicate more local temperature changes as ocean currents changed, possibly in response to the completion of the Panama land bridge (Marshall et al., 1979). It has been proposed that widespread orogen uplift could have changed world climate (e.g. Raymo and Ruddiman, 1992; Raymo et al., 1988) but much of the floral and geomorphological evidence used to infer uplift could equally be a result of climate change. Andean examples are discussed below (see Molnar and England, 1990 for examples from other orogens).

## OVERVIEW OF PLATE TECTONIC SETTING AND KINEMATICS

The Andes define the boundary between the South American continent and four distinct oceanic plates (fig. 2). Relative motions between the Nazca, Antarctic and South America plates between ca. 60 Ma and the present are relatively well-known (DeMets et al., 1990; Pardo-Casas and Molnar, 1987; Pilger, 1984). Since ca. 50 Ma convergence between the plates has been towards ca. 080° but there have been considerable changes in convergence rate (fig. 3) which may broadly correlate with periods of more intense volcanism and shortening in the continental margin. Present day Nazca-South America convergence is towards 080° at ca. 78 mm/yr at 10°S and ca. 84 mm/yr at 25-40°S while, south of the Chile Ridge spreading centre, the Antarctic Plate is subducting towards 100° at ca. 20 mm/yr. Major changes in continental margin orientation with respect to plate convergence have produced very different strain patterns in different parts of the Andes (e.g. Dewey and Lamb, 1992).

The E-W trending Andes of northern Colombia and Venezuela are dominated by obduction of oceanic Caribbean crust onto the South American plate and by large magnitude (displacement >ca. 100 km) dextral faults which link subduction on the Caribbean and central American (Panama) trenches (Kellogg and Vega, 1995; Burke et al., 1984; Mattson, 1984). To the southwest of Panama, the very oblique plate convergence is highly partitioned between subduction zone slip, active dextral slip on NE-SW trending faults within the Andes ( Dewey and Lamb, 1992; Winter and Lavenu, 1989; Pennington, 1981) and foreland thrusting (fig. 2). Recent plate reconstructions indicate a prolonged complex history for the Panama triple junction (Pindell and Tabutt, 1995; Kellogg and Vega, 1995; Restrepo and Toussaint, 1988; Aspden and McCourt, 1986; Burke et al., 1984; Kennerley, 1980). Cretaceous and later collision and accretion of island arc and oceanic crust fragments was followed by ca. 1500 km northeastward migration of Panama towards its present position.

In contrast, central Andean kinematics is dominated by ENE-directed shortening. Present day seismicity suggests approximately 90% of interplate motion is taken up by slip on the subduction zone and 10% by shortening and thickening in the overlying continental margin (Assumpção, 1992; Assumpção and Suarez, 1988; Dewey and Lamb, 1992; Dorbath et al., 1986, 1990, 1991; Chinn and Isacks, 1983; Stauder, 1975; Suarez et al., 1983; Harvard and ISC earthquake catalogues). Shortening directions of most subduction and Andean earthquakes are close to 080° indicating that little strain partitioning is taking place. There is no evidence of terrane pile-up in the Arica Bend region at 19°S to suggest any long-term partitioning of trench slip. Studies of young faulting and active continental margin seismicity between 6°S and 36°S show that only a small fraction of total continental margin strain is partitioned (Kennan, 1993, 1994; Dewey and Lamb, 1992; Cabrera et al., 1987, 1991; Sébrier et al., 1985, 1988; Lavenu, 1979; Lavenu and Ballivian, 1979). Displacements on orogen parallel strike-slip faults are modest (all <<100 km) and relate to accommodation of local changes in fold-thrust directions. The N-S trending Atacama Fault of northern Chile accommodated significant sinistral and dextral slip during the Cretaceous (Brown et al., 1993) but at present shows only

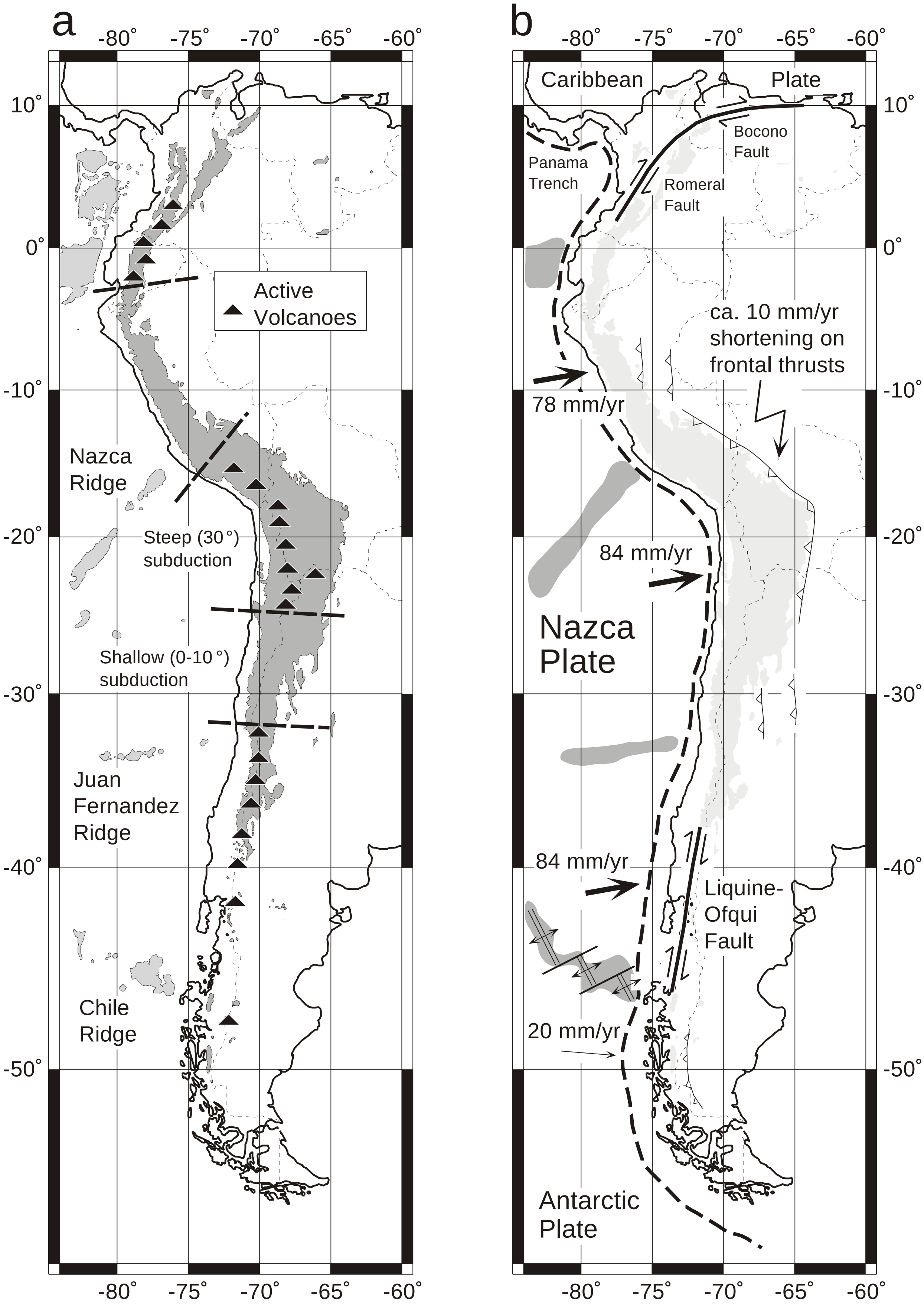



**Fig. 2.** Outline maps of western South America. a, Sea floor -3000 m and surface +1500 m contours outline the main bathymetric and topographic features of western South America. Offshore ridges are named. Note that the locations of recent volcanoes correlate with zones of steep subduction (ca. 30°). The shallow subduction zone between 3°S and 15°S may reflect difficulty subducting the more buoyant Nazca (south) and Carnegie (north) ridges. b, Shows the major plate-kinematic features of the plate margin. Note that partitioning of intraplate slip is strong in the northern Andes and nearly absent in the central Andes. Approximately 10% of Nazca-South America slip is taken up by shortening in the continental margin, and 90% by subduction zone slip. Active earthquakes suggest continental deformation is concentrated in the eastern Andes.

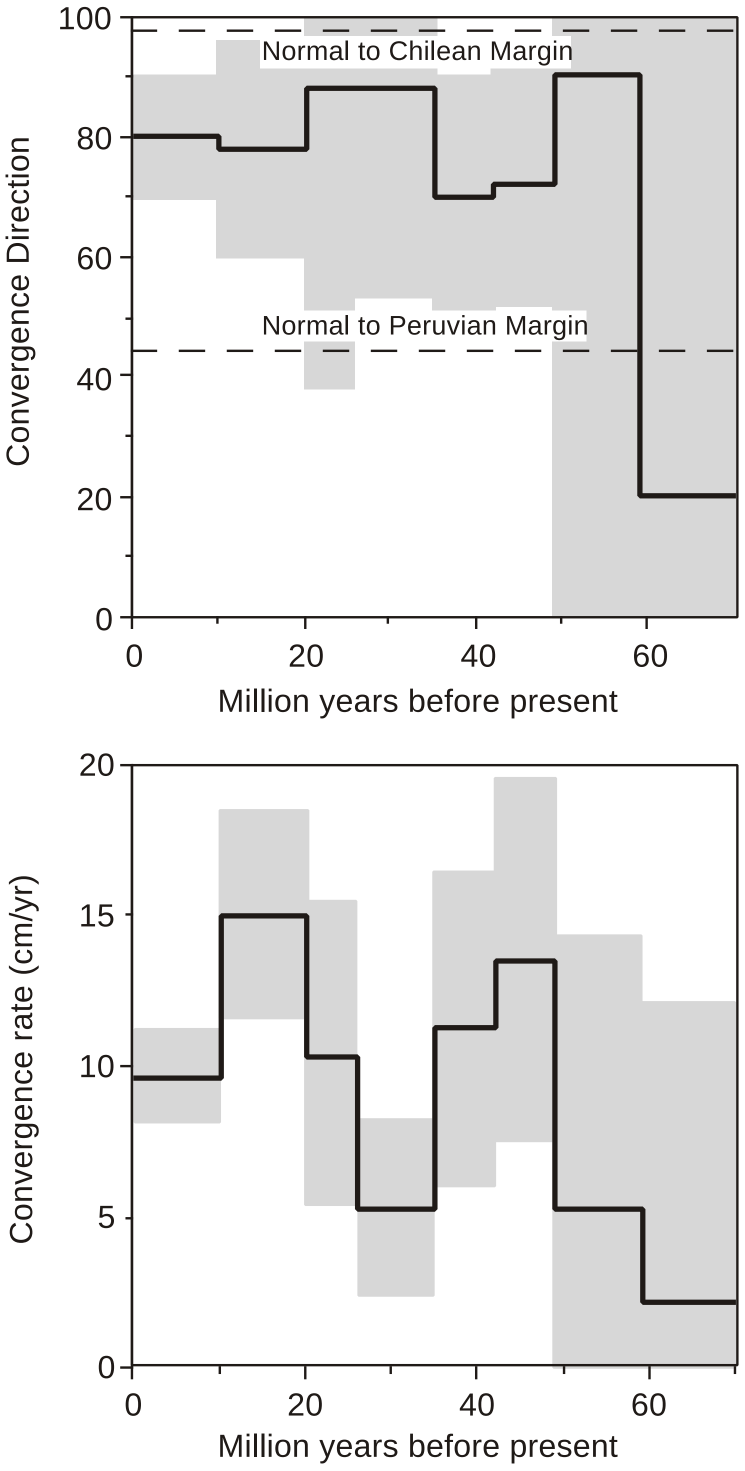



**Fig. 3.** Graphs of rate of Nazca-South America plate convergence and convergence direction plotted against time for 20°S (data from Pardo-Casas and Molnar, 1987). Note that since the early Cenozoic convergence direction has been nearly constant at ca. 080° but there have been important increases in convergence rate in the Eocene and early Miocene which appear to correspond to periods of increased deformation in the continental margin, although convergence rate appears to have been dropping as major late Cenozoic shortening occurred in the Subandean fold-thrust belt in the Eastern Andes. Con vergence has been oblique in Peru and Chile but there is no clear major strike-slip in these regions.

small normal displacements (Dewey and Lamb, 1992).

From about 36-46° S the active Liquine-Ofqui fault appears to be partitioning subduction zone slip at present, resulting in maximum dextral slip at rates of <3 cm/yr (Dewey and Lamb, 1992) and E-W directed shortening. The long term (say 50 Ma) behaviour of the Liquine-Ofqui fault is not well known but Eocene and Miocene basins, possibly pull-aparts, are present along the fault zone (Pankhurst et al., 1992; Forsythe and Prior, 1992). To the south of 46°S, intraplate convergence drops to only ca. 20 mm/yr, normal to the subduction zone, with no continental margin strike-slip (Flint et al., 1994; Murdie et al., 1993; Cande and Leslie, 1986). The area is not very seismically active but Miocene to recent deformation (e.g. Ramos, 1989) suggests a similar ratio of trench-slip to continental margin shortening to farther north.

Along the Andes, subduction is segregated into zoned of shallow (0-10°) and steep (ca. 30°) dip (e.g., Comte et al., 1992; Lindo et al., 1992; Isacks, 1988; Grange et al., 1984). The changes in slab-dip are gradual and do not reflect sharp tears in the subducting slab. Zones of shallow subduction may be due to the buoyancy of younger oceanic crust or of the Carnegie, Nazca and Juan Fernandez aseismic ridges. Zones of steep dip correlate with active volcanism (fig. 2) and the presence of an asthenospheric wedge. Dehydration of sediments on the subducting slab promotes melting within this wedge and during ascent the resulting basaltic magma is modified by fractionation and assimilation of crustal material to a more andesitic average composition (Thorpe, 1982; Davies and Bickle, 1991; Peacock, 1993).

## UPLIFT OF THE NORTHERN ANDES (11°N TO 4°S)

The Andes of central Ecuador and south central Colombia comprise distinct parallel cordilleras with typical elevations of 2500-3500 m, with the highest peaks reaching ca. 5-6000 m. These ranges are separated by the deep intermontane Cauca and Magdalena valleys and flanked by coastal and foreland plains (fig. 4). To the northeast, the Eastern Cordillera splits into the Perijá and Merida ranges which ring Lake Maracaibo, while the triangular isolated Santa Marta range lies farther west. The narrow, E-W trending Caribbean Andes of Venezuela locally reach 2000 m.

The Western Cordillera and coastal plains of Ecuador and Colombia are an oceanic crust and island arc terrane accreted to the continental margin in the middle Cretaceous. The other Cordilleras are cored with continental crust, and share a similar thick, deformed Mesozoic cover indicating that they accreted no later than the Jurassic and possibly during the Precambrian or Palaeozoic (e.g. Restrepo and Toussaint, 1988; Aspden and McCourt, 1986; Kennerley, 1980; Irving, 1975). Cenozoic to recent volcanics are restricted to the Western and Central Cordilleras between 2°S and 6°N, where stratovolcanoes define all the higher peaks (e.g. CGMW, 1978, Arango et al., 1976). The higher peaks in the Santa Marta, Perijá and Merida ranges consist of unroofed metamorphic basement with no volcanic cover.

### Palaeogeographical development

Sedimentary sequences up to 6000 m thick in the foreland, intramontane and foreland basins record the uplift and unroofing of the distinct cordillera. The earliest uplift of the Western and Central Cordillera was caused by middle Cretaceous arc accretion. Major rejuvenation of the Central Cordillera in the Eocene is reflected in the first appearance of conglomerates in the Cauca and Magdalena basins. Farther north, there is some Eocene erosion from the Santa Marta massif (Irving,1975). Only in central Ecuador is Eocene uplift of the Eastern Cordillera apparent. Palaeogene K-Ar ages record metamorphism in Eastern Cordillera schists (Feininger, 1982) and conglomerates appear in the 600 m thick Eocene Tiyuyacu formation in the foreland basin (Dashwood and Abbots, 1990; Kennerley, 1980). At the same time oil was expelled from organic-rich sediments in the Eastern Cordillera (now graphite schists, Feininger, 1975) into actively growing structures.

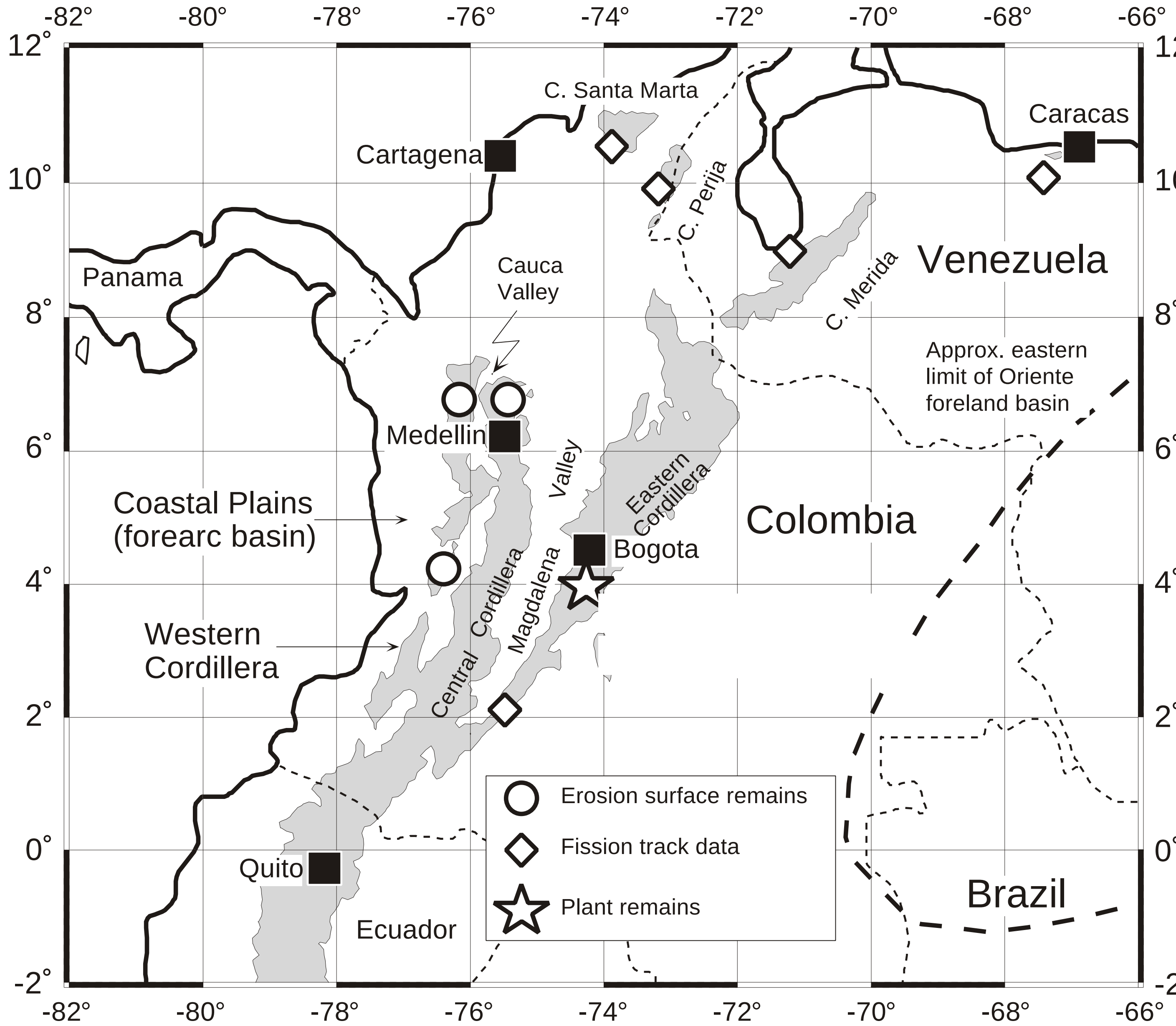
By the Oligocene relief and erosion rates in the Colombian Western and Central Cordilleras were low. Fine-grained facies with marine incursions are noted in surrounding basins and transgress across the Eastern Cordillera region. Major rejuvenation by ca. 18 Ma of the Central Cordillera resulted in conglomerates being across the Magdalena Basin (Honda formation). The first influence of a rising Eastern Cordillera is noted at ca. 12 Ma (van der Wiel and van den Bergh, 1992a,b). Major middle to late Miocene thrusting of the Eastern Cordillera to the west over the Magdalena and east over the foreland (Roeder and Chamberlain, 1995) was accompanied by deposition of this sedimentary wedges (e.g. Cooper et al., 1995). This uplift caused redirection of palaeo-Amazon or Orinoco drainage away from the western Caribbean towards their present easterly courses (Hoorn et al., 1995). Folding and thrusting in the easternmost ranges of Ecuador and Colombia is still active. Major thin-skinned shortening, deforming thick middle to latest Miocene sequences, is also noted in the Cordillera de Perijá, farther north (Kellogg, 1984).

### Fission Track Data

Limited data from the Eastern Cordilleras (sites shown on fig. 4) are consistent with the palaeogeographic outline, indicating that erosion accelerated markedly in the mid-late Miocene. Zircon (Andriessen, 1995) and apatite ages (van der Wiel, 1990) from southern Colombia indicate average 100-12 Ma unroofing rates of ca. 60 m/Ma and post-12 Ma rates of ca. 300 m/Ma. From the Santa Marta, Perijá, Santander and Merida ranges ringing Lake Maracaibo (Kohn et al., 1984a; Shagam et al., 1984; Kroonenberg et al., 1990) show similar low rates of unroofing until the middle Miocene. Subsequent rates were ca. 300-500 m/Ma over wide areas of these ranges with very localised rates of > 800 m/Ma on narrow, fault slivers. Farther east, the coastal ranges near Caracas have much younger zircon ages of 17-24 Ma suggesting that erosion rates of >300 m/Ma were sustained for longer, since the earliest Miocene (Kohn et al., 1984b).

### Palaeobotanical evidence

Pollen from Pliocene sediments in the Sabana de Bogota (fig. 4), an intramontane basin now at ca. 2500 m in the Colombia Eastern Cordillera show tropical floras at ca. 4 Ma changing rapidly to high altitude floras by 3 Ma (Kroonenberg et al., 1990; Andriessen et al., 1993). Tropical floras are now found below 500 m but ca. 2000 m of rapid mid-Pliocene surface uplift can only be inferred if there was no regional climate



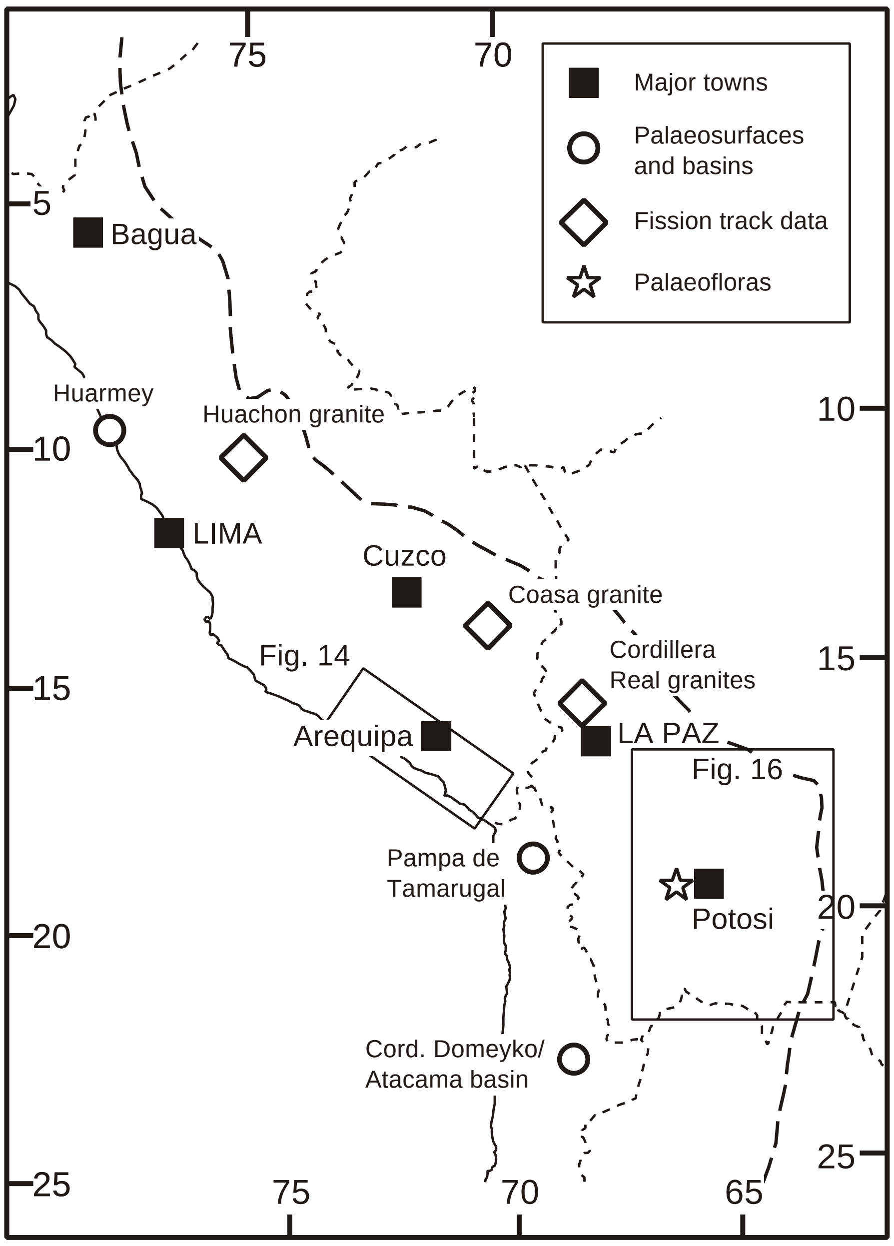




**Fig. 4**. (above) Outline topographic map of the northern Andes of Ecuador, Colombia and Venezuela showing areas above 1500 m, major cities and sites discussed in the text. Here the Andes consist of several parallel Cordillera separated by deep intramontane valleys with thick Cenozoic sediment fills. These valleys are bounded by active thrusts and by large-displacement dextral faults. The Eastern Cordillera/Perijá block uplifted most recently (ca. 10 Ma), sepa rating the Magdalena valley from the foreland and diverting palaeo-Amazon/Orinoco flow from the north towards the east.

**Fig. 5**. (left) Location map for Central Andes. Sites for fission track data, palaeoflora, palaeosurfaces and major towns are shown, in addition to box outlines of the locations of figures 14 and 16.

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change at the time. There is no apparent sudden overdeepening of the foreland basin at this time which might result from greater topographic load on the Guyana shield (e.g. Cooper et al., 1995). There are no reports of palaeofloras from the region that predate possible Pliocene climate change.

### High-altitude palaeosurfaces in Colombia

Low relief erosion surfaces between ca. 2200 m and 3000 m are reported from the Western and Central Cordilleras of Colombia (Page and James, 1981; Padilla, 1981. see fig. 4). The higher ca. 3000 m surfaces near Medellin appear to underlie the early to middle Miocene Honda formation (e.g. van der Wiel and van den Burgh, 1992a) which lies at ca. 6000 m below sea-level in the Magdalena valley, indicating vertical displacement of 8000-9000 m. Uplift of the surfaces from near sea-level (present Magdalena valley floor altitude is only ca. 100 m) must have started at ca. 18-15 Ma because the rejuvenated Central Cordillera sourced the Honda formation (van der Wiel and van den Bergh, 1992a) suggesting an average surface uplift rate of ca. 200 m/Ma. The presence of a lower plain at ca. 2200 m may indicate that this uplift was pulsed. Deep gorges dissected the surfaces in Pliocene to recent time. The isostatic rebound due to this unloading is probably no more than ca. 100 m.

## UPLIFT OF THE CENTRAL ANDES (4°S TO 46°S)

The Large-scale morphology of the central Andes is strikingly different from farther north. The most prominent feature is the "Bolivian Orocline", a abrupt change in topographic trend from NE-SW in Peru to N-S in northern Argentina and Chile. In the core of this orocline lies the central Andean Plateau (11°N to 28°S), up to 500 km wide and with extremely constant average altitude of ca. 4000 m (figs. 1 and 5). Plateau width, but not elevation, drops symmetrically to the northwest and south. Average elevations along the western and eastern margins of the plateau are slightly higher, with peaks locally reaching near 7000 m (glaciated above ca. 4800 m). Much of the plateau is occupied by the Altiplano (Peru/Bolivia) and the Puna (Argentina), a compound internal drainage basin interrupted by low relief bedrock ridges (fig. 6) and much of the distinctive very flat relief of the region comprises the youngest basin fills. The surrounding regions are low relief undulose highlands cutting across Palaeozoic-Mesozoic bedrock (fig. 6). The plateau margins have very steep topographic gradients (up to 1 in 15) and are dissected by rivers draining into the Pacific and Atlantic, but there is almost no dissection deep into the plateau. In northern Chile, the western Andean slope is disrupted by a narrow coastal range reaching ca. 2500 m, a 50 km wide sediment-filled "central depression" and the discontinuous "Precordillera" (comprising the Cordilleras de Morena and Domeyko). Dissection of the eastern slopes is strikingly more intense north of 18°S coincident with slightly higher peaks around the plateau rim (Masek et al., 1994).

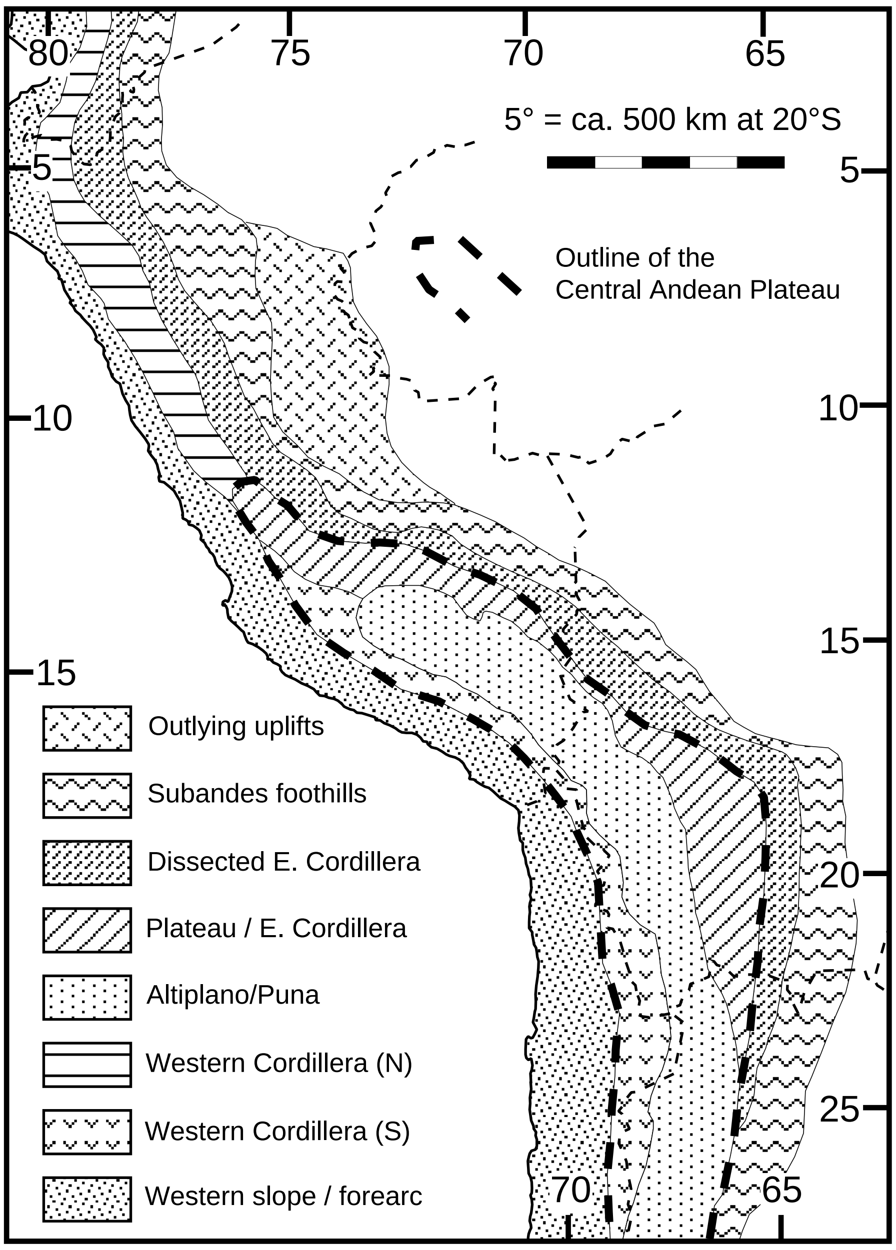
Average elevation and width of the Andes drops dramatically north and south of the plateau region. By ca. 5° N and 45°S elevations are typically 500-1000 m with only isolated peaks reaching 2000-3000 m. These regions are entirely externally drained (fig. 7) and river valleys are typically very deep and narrow. For example, the bed of the Marañon valley in Peru is up to 3000 m below surrounding peaks for much of its ca. 500 km length. In contrast to the plateau region, there are distinct outlying ranges up to 1000-2000 m high up to 250 km beyond the main Andean range. In both northern Peru and northern Argentina these ranges trend north-south.

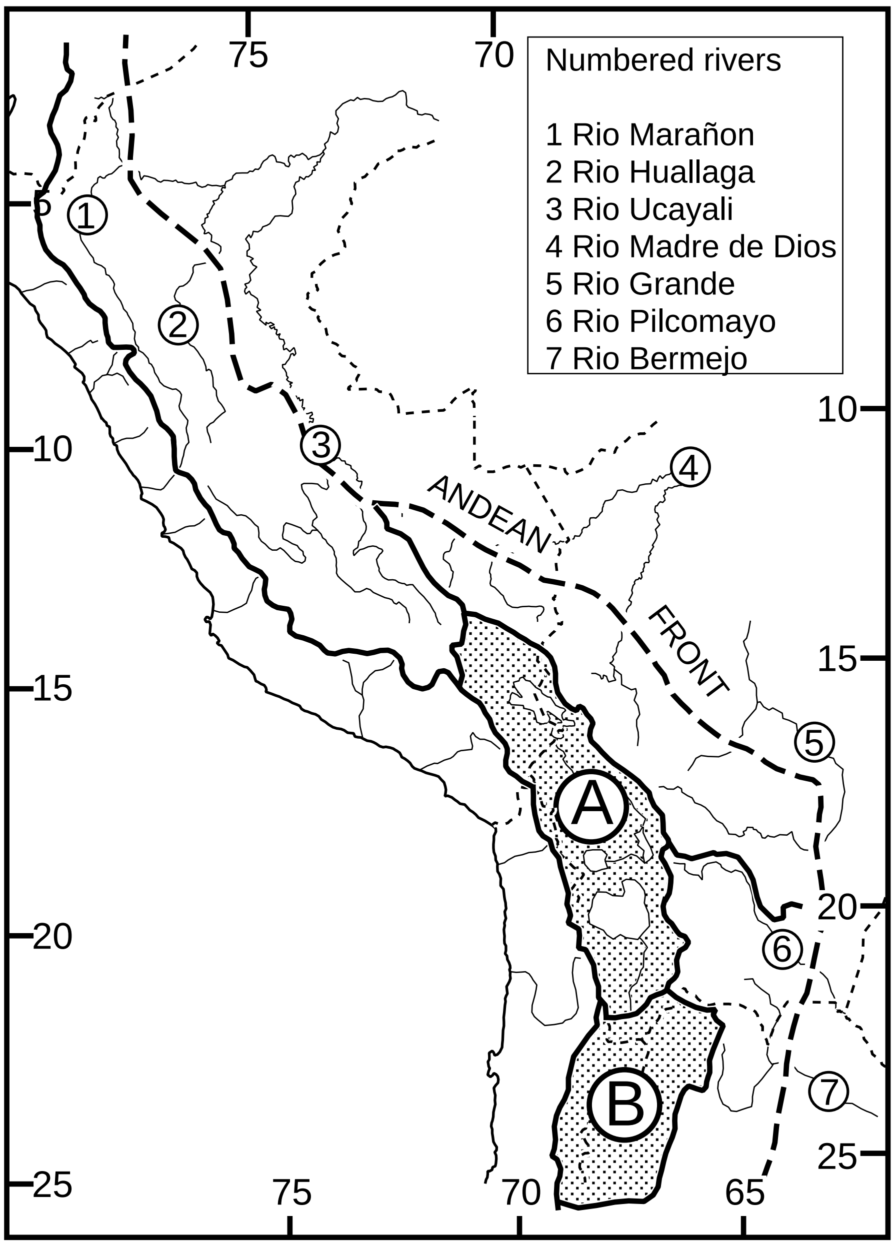
Bedrock geology of the central Andes can be conveniently described in terms of orogen-parallel zones which broadly parallel the topography (fig. 8). Coastal regions of Peru and Chile are dominated by the granites of the coastal batholith (e.g. Mukasa, 1986; Pitcher et al., 1985; Coira et al., 1982) which intrude the Precambrian basement of the Arequipa Massif (Shackleton et al., 1979) and deep extensional marginal basins formed during the Mesozoic and which predate Andean uplift (e.g. Flint and Turner, 1988; Cobbing, 1985).

Offshore, between the coastal batholith and the Peru-Chile Trench, there are elongate overlapping forearc basins filled with Eocene to recent sediments (e.g. von Huene et al., 1987). Where the Carnegie and Nazca ridges collide with the continent, these fore-arc basins have been uplifted above sea-level (e.g. Dunbar et al., 1990). In northern Chile, the Cordilleras de Domeyko and Morena comprise deformed basement to Mesozoic rocks flanked by the deep Tamarugal (central depression) and Atacama basins (e.g. Buddin et al., 1993; Flint et al., 1993; Jolley et al., 1990). The western edge of the central Andean Plateau is dominated by Eocene to recent volcanic rocks, mainly andesite flows and rhyolitic ignimbrites related to subduction zone dehydration and melting (e.g. Thorpe et al., 1982). The prominent peaks of the Western Cordillera are Pleistocene stratovolcanoes spaced fairly regularly at ca. 50 km intervals. East of the Western Cordillera the Altiplano-Puna is a complex of Cenozoic sedimentary basins filled with up to ca. 10 km of fine and coarse-grained red-beds with intermittent volcanic intercalations most common in the west (e.g. Lamb et al., in press; Kennan et al., 1995).

In contrast, the dominant rocks of the Eastern Cordilleras are Precambrian and Palaeozoic sediments and greenschist to amphibolite facies metamorphic rocks with Mesozoic-Cenozoic strata widely preserved in thrust footwalls and syncline cores (e.g. Pareja et al., 1978; INGEOMIN, 1978). North of ca. 12°N, lower Palaeozoic rocks are absent and Carboniferous and younger strata lie directly on late Precambrian schists and gneisses (Dalmayrac, 1978). In the Macusani (14°S), Morococala (18°S) and Los Frailes (19°-20°S) distinctive ignimbrite shields with highly peraluminous composition were erupted along the western edge of the Eastern Cordillera at ca. 12 Ma to 4 Ma (Cheillitz et al., 1992; Ericksen et al., 1990; Bonhomme et al., 1988; Grant et al., 1979). In contrast to the main volcanic arc these rocks are not related directly to the subduction zone, but possibly reflect melting of crustal material.

Seismic sections (unpublished oil company data) indicate that, throughout the Cenozoic, the Eastern Cordillera has been overthrusting the eastern basins of the Altiplano, although average elevations in the Cordillera are only slightly higher. The eastern margin of the cordillera, approximately coincident with the edge of the central Andean plateau, is also defined geologically by major overthrusting over the Subandean ranges, a thin-skinned fold and thrust belt involving Palaeozoic, Mesozoic and late Cenozoic strata (e.g.



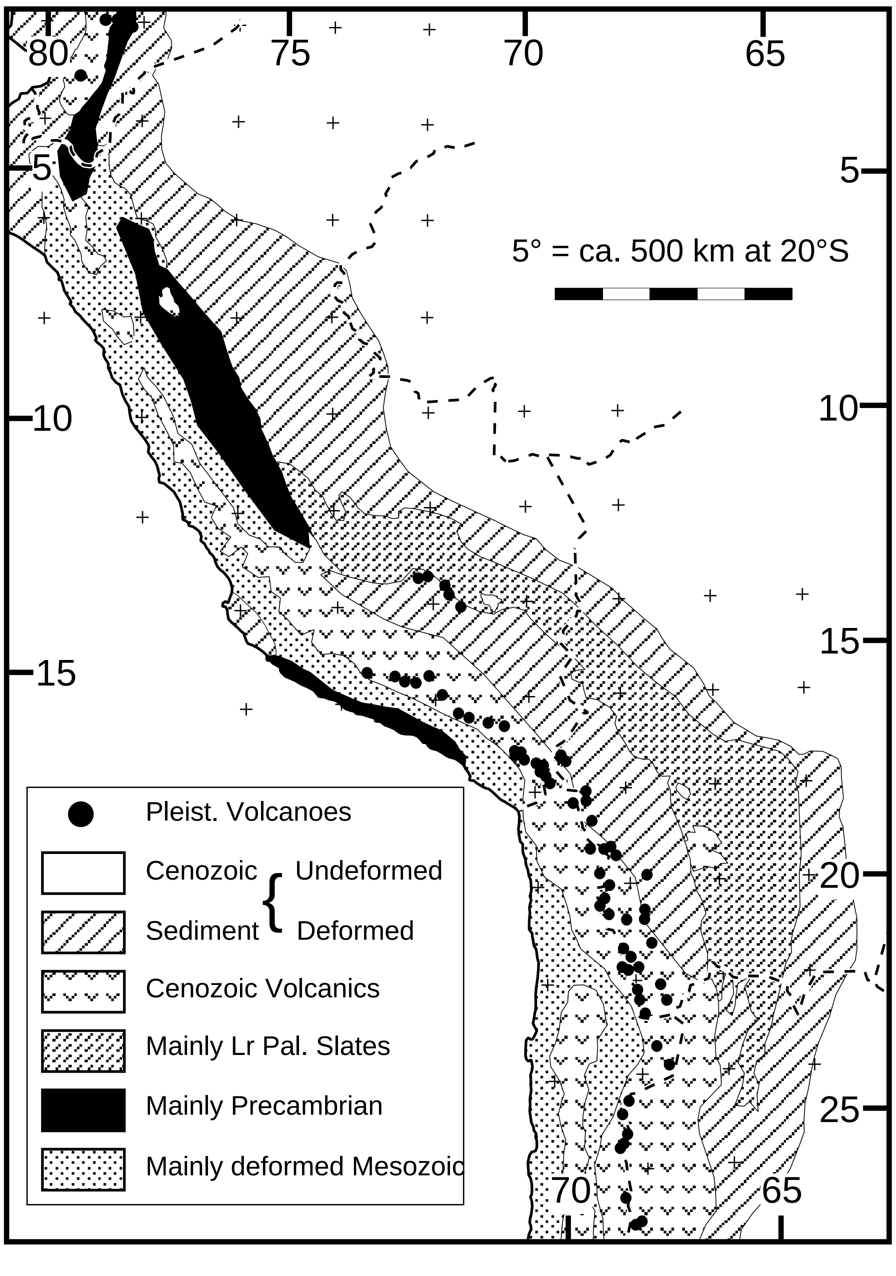




**Fig. 6.** (top left) Geomorphological zonation of the central Andes. The most prominent feature of the region is the central Andean plateau, the centre of which is a compound internal drainage basin. The Eastern and northern margins are low relief, little dissected undulose surfaces cutting across Palaeozoic to Cenozoic bedrock, while the western margin is defined by the active volcanic arc. The steep eastern slope consists of deeply dissected lower Palaeozoic rocks while the lower gradient Subandes is a deforming fold-thrust belt. Rainfall is high in these regions north of ca. 18°S. In northern the Peru the Western Cordillera consists of deeply dissected Mesozoic rocks with a reduced volume of Cenozoic volcanics. The western slopes are extremely arid throughout the region, consisting of salt pans, local ranges of hills and deep gorges cut by ephemeral streams draining from the Western Cordillera.

**Fig. 7.** (top right) Major rivers of the central Andes. Note that much of the central Andean plateau region is an area of internal drainage, with ephemeral rivers draining into lakes and salt pans in the (A) Altiplano and (B) Puna-Atacama basins. North of this region the main watershed coincides approximately with the volcanic Western Cordillera. Northern parts of the plateau are externally drained and deeply dissected. In Subandean regions three main watersheds divide waters draining directly into the upper Amazon (rivers 1,2,3), joining the Amazon in central Brazil (rivers 4,5) and draining into the Rio de la Plata in Argentina (rivers 6,7). The main east-west watershed and the boundaries of the internal drainage basins have been fairly constant since at least 25 Ma.

**Fig. 8.** (bottom left) Simplified geological map of the central Andes. The deep Mesozoic volcanic and sediment filled basins of the western Andes are intruded by coastal batholith granites (not shown). The eastern Andes are mainly composed of Precambrian schists and granites (north of 12°N) and low grade Palaeozoic sediments (south of 12°N). Mesozoic-Cenozoic sediments are preserved locally in syncline cores. The thick Cenozoic sediments of the Altiplano and Subandean basins were deformed mainly in the last 10 Ma. Note that Cenozoic volcanics are more widespread that recent volcanoes. This may indicate shallowing of subduction angle where they are absent occurred in the last 10-5 Ma.

Baby et al., 1992; Roeder, 1988). These Cenozoic strata are a deformed foreland basin fill, continuous with sequences in the present-day foreland. Active deformation in the central Andes is concentrated in this foreland zone. The importance of thrusting and folding in the eastern foothills declines to north and south of the plateau, where basement-cored anticlines, bounded by steep reverse faults (Allmendinger et al., 1990, 1983) define the outlying uplifts of the Sierras Pampeanas (Argentina) and the Sierras de Shira, Contaya and Moa (northern Peru).

**Palaeogeographic development**

Middle to late Cretaceous marine sediments cover much of the central Andes (Jaillard, 1994; Riccardi, 1988) except coastal regions (fig. 9). In Peru, a narrow fold-thrust belt emerged and shed coarse sediments east into basins in the western Altiplano (e.g. Vicente et al., 1979; Noble et al., 1990) while transpression on N-S faults in northern Chile uplifted the Cordillera de Domeyko while localised extension or transtension formed the Atacama Basin (Flint et al., 1993). In the late Cretaceous to Palaeocene the sediments in these early clastic basins coarsened and clastic sediments became more common that carbonates farther east (e.g. Jaillard, 1994; Flint et al.,1993; Gayet et al., 1991). During the Palaeocene and Eocene deformation in this western belt intensified and resulting unconformities are sealed by a major volcanic flare-up (after a 50 Ma period with little eruption) dated at ca. 42-36 Ma in west-central Peru (e.g. Noble et al., 1990, 1979). The unconformities die out rapidly to the east into the Altiplano where sharp influx of west-derived sands is recorded (e.g. Bagua basin, Naeser et al., 1991, Cuzco and Sicuani basins, Noblet et al., 1987; Lopez and Cordova, 1988; Jaillard et al., 1993; Chavez et al., 1994). Pre-late Eocene sands were also shed west into forearc basins (e.g. von Huene et al., 1987). In northern Chile, the coarser upper members of the Purilactis group in the Atacama basin record continued unroofing in the Cordillera Domeyko (Hartley, 1993).

Parts of the Eastern Cordillera also first rose at this time. From southern Peru to southern Bolivia a proto-Cordillera (fig. 10) rose bounded to east and west by active thrust belts, and the Mesozoic cover of the region eroded into the Altiplano (e.g. Kennan et al., 1995; Ellison et al., 1989; Rodrigo and Castaños, 1975). Remnants of a narrow foreland basin, which pinched out west of the Subandes, are preserved in synclines in the eastern part of the present Cordillera (e.g. Kennan et al., 1995; Kennan, 1994). Sediments in Altiplano and foreland basins show no sign of Eastern Cordillera unroofing between ca. 6°S and 10°S (e.g. Noblet et al., 1987; Koch and Blissenbach, 1962). In northernmost Peru (fig. 10), foreland basin sequences older than ca. 40 Ma thicken up to >1000 m (e.g. Rosenzweig, 1953) and distinctive schist clasts in sediments just above the Cretaceous indicate that in places 4000 m of Mesozoic cover was eroded very quickly (e.g. Tafur, 1991).

Between ca. 36 Ma and 26 Ma, coincident with a drop in intraplate convergence rate (fig. 3) there was little or no deposition in onshore or forearc basins, and almost no volcanic activity. In the latest Oligocene to middle Miocene (fig. 11), ignimbrites and andesites were erupted throughout the entire Western Cordillera (e.g. Lahsen, 1982; Noble et al., 1990, 1985, 1979b; McKee and Noble, 1982). Multiple episodes of coarse-grained sedimentation, folding or tilting, and erosion are recorded in the Western Cordillera of Peru and Chile (e.g. Flint, 1985; Mortimer and Saric, 1975), Altiplano basins (Lamb et al., in press; Kennan et al., 1995; Naeser et al., 1991; Mégard et al., 1984) and in the forearc (e.g. von Huene et al., 1987). By the early Miocene the entire Eastern Cordillera from 5°S to >23°S was emergent and eroding. Early-middle Miocene subsidence of >>1000 m (Koch and Blissenbach, 1962) was probably driven by overthrusting and flexure. In the Bolivian Subandes, thin clastic sequences (400 m deposited from ca. 27 Ma to 10 Ma) end a 40 Ma hiatus as sedimentation spread east of the Eastern Cordillera (Marshall and Sémpere, 1991; Sanjines and Jimenez, 1976). As deformation progressed sedimentation in the Altiplano and parts of the Eastern Cordillera tended to concentrate in relatively narrow fault-bounded basins, some of them very deep (e.g. Lamb et al., in press; Chavez et al., 1994; Sémpere et al., 1994; Hérail et al., 1993; Laubacher et al., 1988). Lower Palaeozoic debris is seen for the first time in many basins (authors own observations).

Shortening in the Eastern Cordillera continued until ca. 10 Ma when several, flat-lying ignimbrite shields were erupted. Sedimentation and deformation continued in the Altiplano basins and rapid (>600 m/Ma) deposition of sandstones and conglomerates started in the Subandes, as the Eastern Cordillera thrust east onto the Brazilian shield (fig. 12). By ca. 5 Ma internal shortening had more or less ceased on the Altiplano (Kennan et al., 1995) and thin-skinned deformation of the foreland basin sequences in the Subandes started (e.g. Baby et al., 1993). At present the eastern range front of the Andes is a major watershed. Erosion is concentrated in this region and almost absent from the plateau region to the west.

**Fission track and other cooling ages in Peru and Bolivia**

A few fission-track and K-Ar/Ar-Ar mica ages are available from three areas of the Eastern Cordillera (see fig. 5). In the Huachon region (ca. 10°S) Triassic and older granites intrude probable Precambrian schists. Apatite fission track ages (Laubacher and Naeser, 1994) suggest that denudation rates were ca. 70 m/Ma between 12 Ma and 22 Ma and about 300 m/Ma between 12 Ma and the present. Much older zircon ages suggest average cooling rates of less than 1.25°C/Ma between 100-20 Ma, equivalent to denudation of only 5 m/Ma at 25°C/km geothermal gradient. From the data given, there is no clear evidence of a significant early Cenozoic event, consistent with the thin, fine-grained sequences seen in the foreland.

Data from southern Peru (ca. 14°S) is harder to interpret. Here, the Triassic Coasa batholith intrudes low-grade Palaeozoic schists (Kontak et al., 1990a). The easternmost granites show clear 40 Ma K-Ar and Ar-Ar biotite ages and younger, 20-30 Ma fission track ages. Kontak et al., (1990b) have interpreted these ages as due to a cryptic thermal perturbation at ca. 40 Ma which reset the biotites. However, such a long-lived upper crustal thermal anomaly seems unlikely. For instance, a major high-level vein-forming and stock intrusion episode only 25 km to the southwest shows concordant K-Ar and fission-track ages of ca. 25 Ma indicating rapid cooling (Kontak et al., 1987). Alternatively, the granite age data could indicate cooling from ca. 250°C to 100°C between 40 Ma and ca. 25 Ma due to unroofing. At typical regional geothermal gradients this would require ca.

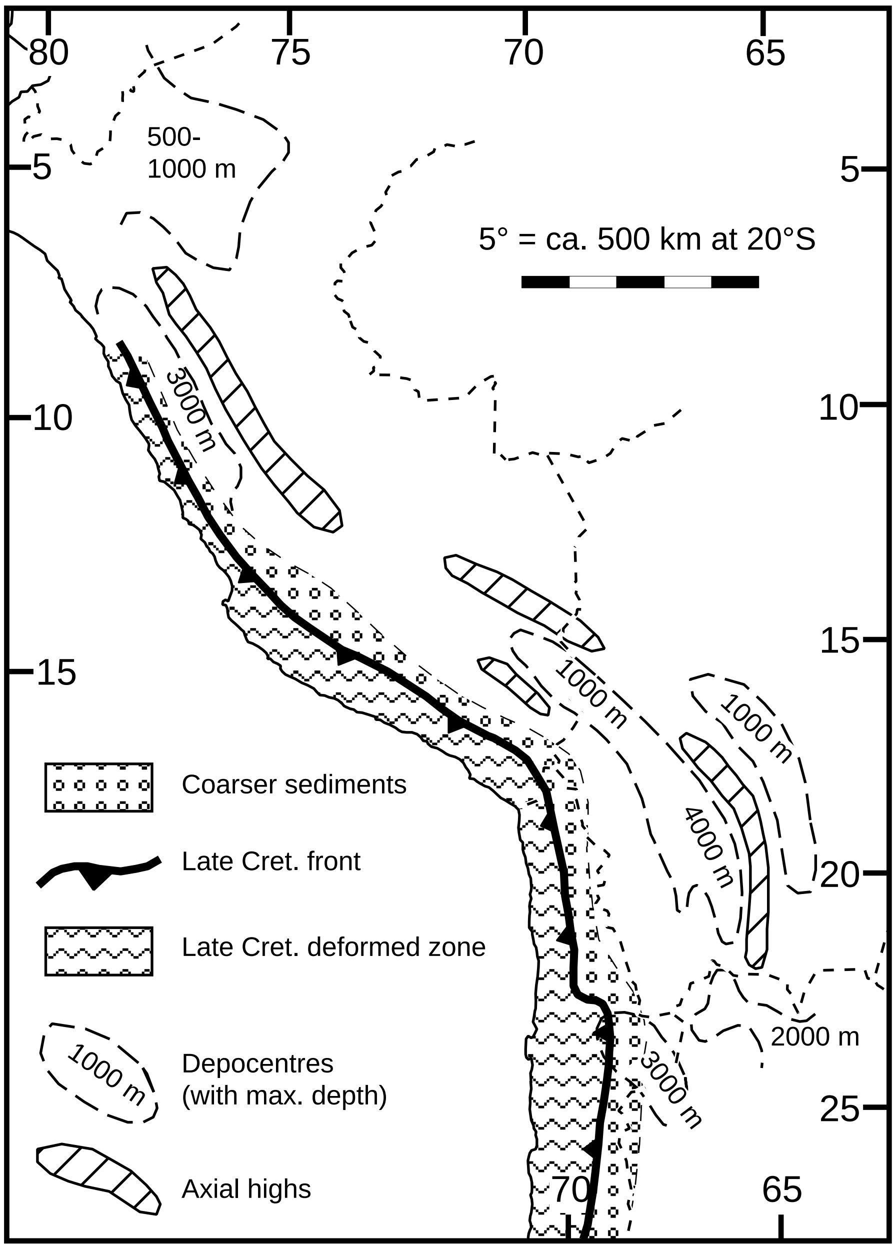



**Fig. 9.** (top left) Late Cretaceous-early Cenozoic Palaeogeography generalised from sources quoted in the text. No attempt has been made to palinspatically restore the effects of Andean E-W shortening. Note that deformation was restricted to immediate coastal regions with a narrow apron of coarse sediments to the east. Over almost all the Andes and foreland there are widely correlatable sequences of fine-grained clastic and carbonate rocks. Also shown are the main Cretaceous depocentres and axial highs. These were zones of reduced or no subsidence from the Albian onwards where locally the Cretaceous sequence is condensed to a few metres of nodular limestones (e.g. near Aiquile, Bolivia). The depocentres were bounded by normal faults overlapped in most cases by the latest Cretaceous.

**Fig. 10.** (bottom left) Palaeocene-Eocene (ca. 50-35 Ma) palaeogeography. Note that the limits of observed eroding regions coincide with the Cretaceous depocentres suggesting a basin inversion origin. The entire western Cordillera region was deformed and shed sediment to the west into forearc basins and east into the Altiplano. Eruption of thick Andesite and dacite flows at ca. 40 Ma sealed many associated unconformities. The Eastern Cordillera between 14°S and ?23°S started to unroof at about 50 Ma, and was bounded to east and west by active reverse faults. The eastern foreland basin was very narrow: no Palaeogene sediments are seen in the Bolivian Subandes. Although much of the Peruvian Eastern Cordillera was not eroding at this time sediments up to 1000 m thick are widespread in the foreland. The northernmost uplift (5°S) resulted in coarse Precambrian clasts shed to the east in the earliest Cenozoic.

**Fig. 11.** (bottom right) Early to middle Miocene (ca. 25-10 Ma) palaeogeography. By ca. 20 Ma the entire Eastern Cordillera was uplifting and eroding shedding metamorphic and granite clasts into the foreland basin. The Altiplano had become a closed internal drainage basin with volcaniclastic sediment sourced from the active Western Cordillera volcanoes and Palaeozoic and Mesozoic clasts coming from the Eastern Cordillera. The Eastern Cordillera spread to the east and sedimentation started in the Bolivian Subandes. Within the Cordillera there were numerous narrow fault bounded basins (e.g. C, Crucero basin and B, Bagua basin). The Atacama basin (A) and Tamarugal (T) forearc basins in northern Chile filled with volcanics and clasts derived from the Western Cordillera and older uplifted blocks, such as the Cordillera Domeyko. In southern Peru, a marine incursion at 25 Ma deposited limestones up to 40 km inland. Until ca. 12 Ma volcanism was restricted to the west, but from 12 Ma several ignimbrite shields formed along the Eastern Cordillera-Altiplano boundary.

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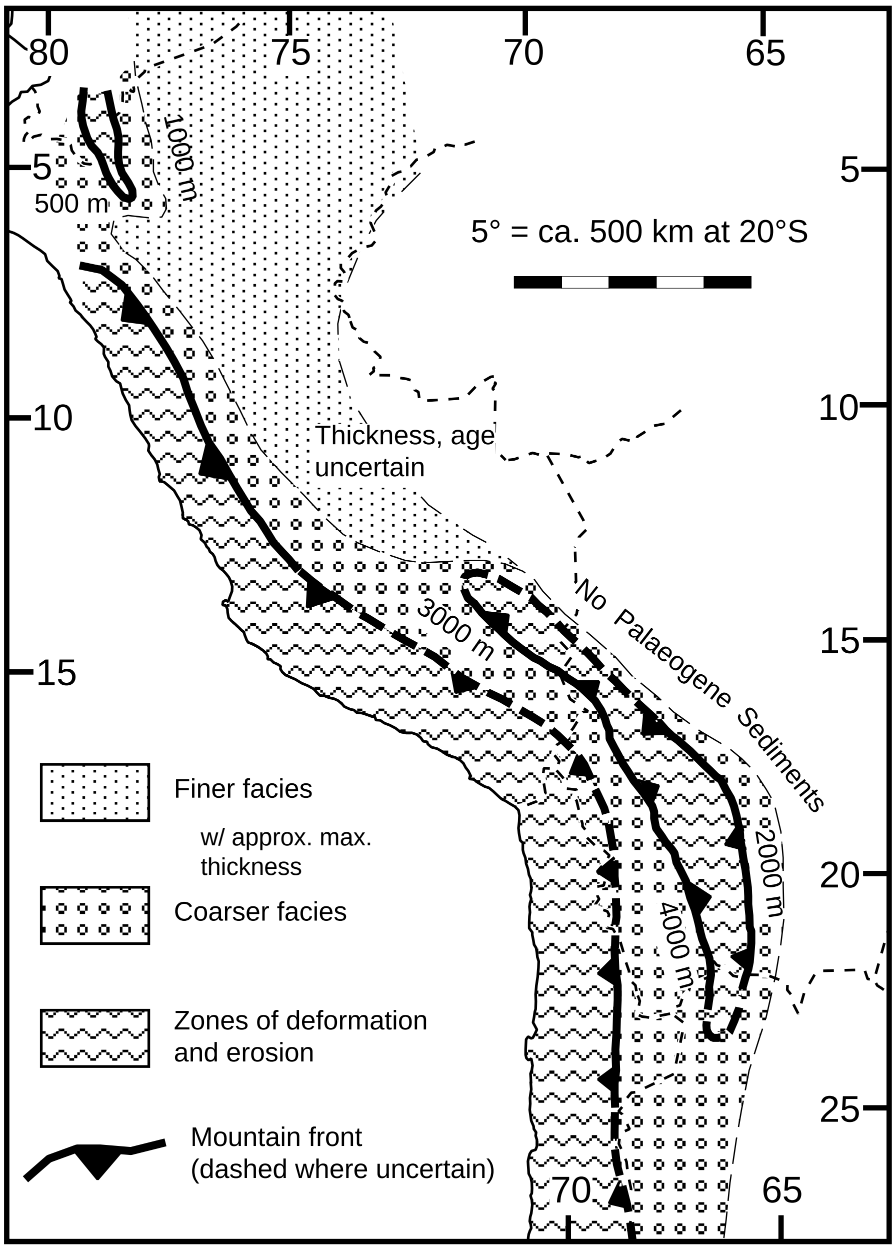


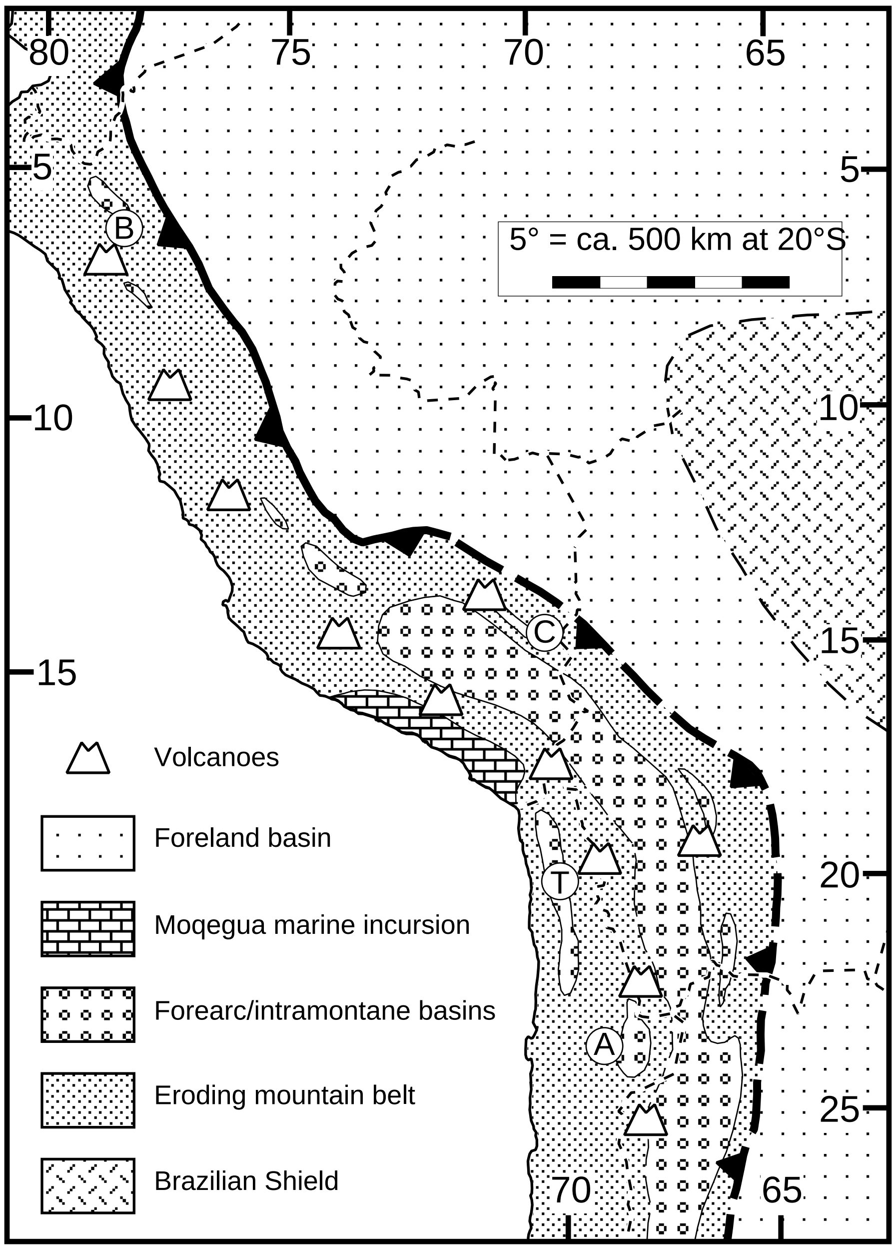

10 km of erosion. The hypothesis can be tested by looking for distinctive clasts in foreland basin sediments of this age (Kennan, work in progress). Miocene and younger cooling rates may have averaged no more than ca. 3-6°C/Ma, equivalent to unroofing rates of 120-240 m/Ma if the geothermal gradient was ca. 25°C/km. No zircon data are available for this region.

The most detailed, and most easily interpreted, data are available from the Cordillera Real of northern Bolivia. Apatite and zircon ages (Benjamin et al., 1987; Crough, 1983) indicate clearly that little unroofing occurred before ca. 50 Ma. Between 50 Ma and 30 Ma unroofing rates were ca. 100-150 m/Ma increasing to 300-400 m/Ma between 30 Ma and the present (fig. 13). This is entirely consistent palaeogeography discussed above. No fission track data are available from farther south.

## Palaeobotanical evidence

Berry (e.g. 1939 and references therein) has described some palaeofloras from several localities in Ecuador, Peru and Bolivia and has compared them to nearest living relatives. Unfortunately, many of the localities are poorly described and cannot dated accurately using more modern isotopic data. Plants found at Potosi, Bolivia (ca. 4500 m in the Eastern Cordillera) may be as old as 22 Ma and compare with living floras found at altitudes 2000-3000 m. Some of these plants are intolerant of freezing (Lena Stranks, pers. comm.) and are not now seen above ca. 2500 m in Bolivia. In the absence of climate change a maximum Miocene to recent uplift can be inferred but the true figure may be somewhat lower. There are no published leaf morphology studies on Andean palaeofloras.

## High altitude palaeosurfaces in Peru, Chile and Bolivia

Palaeosurfaces seem to be more widely preserved, and better described, in the central Andean plateau region than elsewhere and they can be tied to marine deposits to that surface uplift can be reliably estimated.

In southern Peru (15-18°S. figs. 5 and 14) a prominent seaward dipping erosion surface is covered by a few metres of marine limestone dated at 25 Ma (Noble et al., 1985; Tosdal et al., 1984, 1981). These overlie gently folded Eocene conglomerates and are overlain by conglomerates and ignimbrites. This marine erosion surface is at about 400-1000 m close to the coast and can be traced at least 40 km inland to an altitude of ca. 2000 m giving an averages, defining an amazingly unbroken tilted slope with a dip of 2-2.5°. Northwest of the marine incursion a slightly steeper degradational surface rises to the Western Cordillera watershed, mantled by Miocene to recent volcanic rocks. There seems to be little offset of the aggradational and degradational parts of this surface. Thus at ca. 25 Ma it seems a low relief forearc rose northeast to a watershed at perhaps 1000-1500 m. Subsequent simple tilting suggests about 2-3000 m of uplift of the Western Cordillera. As uplift progressed a series of nested lower surfaces and valleys, with successively lower present-day slopes, were cut. Although it is not possible to quantify tilting rates from such dynamic features are river profiles it is interesting that the most rapid change in slope occurred between about 14 Ma and 9 Ma. Myers (1976) has described a superficially similar series of surfaces from near Huarmey (10°S, central Peru) which were also nested during the Miocene with the most recent phase of valley incision being post 5 Ma.

Similar seaward tilting also appears to have occurred in northern Chile (19-22°S). The Pampa de Tamarugal is a half-graben basin bounded by the Atacama fault and Coastal Cordillera (fig. 15b). The floor of the basin in an erosion surface of early-middle Cenozoic age mantled by earliest Miocene ignimbrites which define a simple monocline rising east to the volcanic arc (Galli-Olivier, 1967). Faulting and tilting started in the early Miocene (Paskoff and Naranjo, 1983; Mortimer and Saric, 1975; Mortimer et al., 1974) and post-Pliocene offsets are known (e.g. fig. 7 of Dewey and Lamb, 1992). Sediments ponded against the Coastal Cordillera as the basin deepened but by about 10 Ma the basin had filled. Rivers overtopped the Cordillera gaining a direct connection with the sea. The subsequent incision of gorges (Mortimer, 1980) thus is related to a change from internal to marine baselevel and not to rapid uplift of the coastal block. South of ca. 22°S drainage had a direct marine connection and low-altitude late Miocene pediments (Mortimer, 1973) suggest little coastal uplift of this region, separated from the Cordillera Domeyko by young thrusts (e.g. Jolley et al., 1990). In both southern Peru (Tosdal et al., 1984) and northern Chile (Mortimer and Saric, 1975; Hollingworth, 1964) substantial seaward tilts are seen in Quaternary lacustrine deposits which suggest that tilting is still active. Pleistocene coastal terraces are uplifted in northern Peru (De Vries,1988) and in southern Peru (Macharé and Ortlieb, 1992) but are probably local transient effects and not widely applicable.

In contrast, erosion surfaces related to drainage into the foreland basin in the eastern Andes are well preserved in south-central Bolivia (fig. 16). Servant et al. (1989), Gubbels et al. (1993) and Kennan et al. (in press) have described extremely flat surface remnants up to 10 km x 30 km over an area of ca. 600 km x 100 km, always east of the Altiplano-foreland drainage divide. These surfaces lie at 2200-3800 m and cut directly onto folded Palaeozoic to Cenozoic bedrock. Clastic cover is usually thin or absent but reaches ca. 100 m where sediments have banked against protruding ridges. The surfaces define flat-bottomed valleys commonly with steep side scarps. These valleys trend N-S, controlled by regional structures, and valley bottom altitude systematically drops as they meander towards the east. Downstream gradient drops downstream from ca. 1 in 125 to as low as 1 in 200-250. The lowest surfaces, at ca. 2200 m are found immediately west of the Subandes fold and thrust belt. The surfaces cut rocks as young as 15 Ma and are mantled by 9 Ma to 3 Ma tuffs (Kennan, in press and references therein) suggesting that they were cut in the mid-late Miocene. Headward erosion by gorges up to 1000 m deep post-dates 3 Ma. The volume of sediment eroded during surface cutting must have been ca. 1-2 x 104 km3 (Kennan et al., in press) but no sufficiently large sediment sinks exist in the Eastern Cordillera suggesting that the ultimate sink and baselevel was the late Miocene foreland basin (now deformed as the Subandes) which was at or near sea-level (the ca. 10 Ma Yecua formation contains marine fossils). Even allowing for significant structural shortening, the lowest remnants are unlikely to have been more than 500 m above sea-level suggesting that ca. 2000 m of surface uplift has in the Eastern Cordillera of Bolivia since ca. 10 Ma (fig. 17).

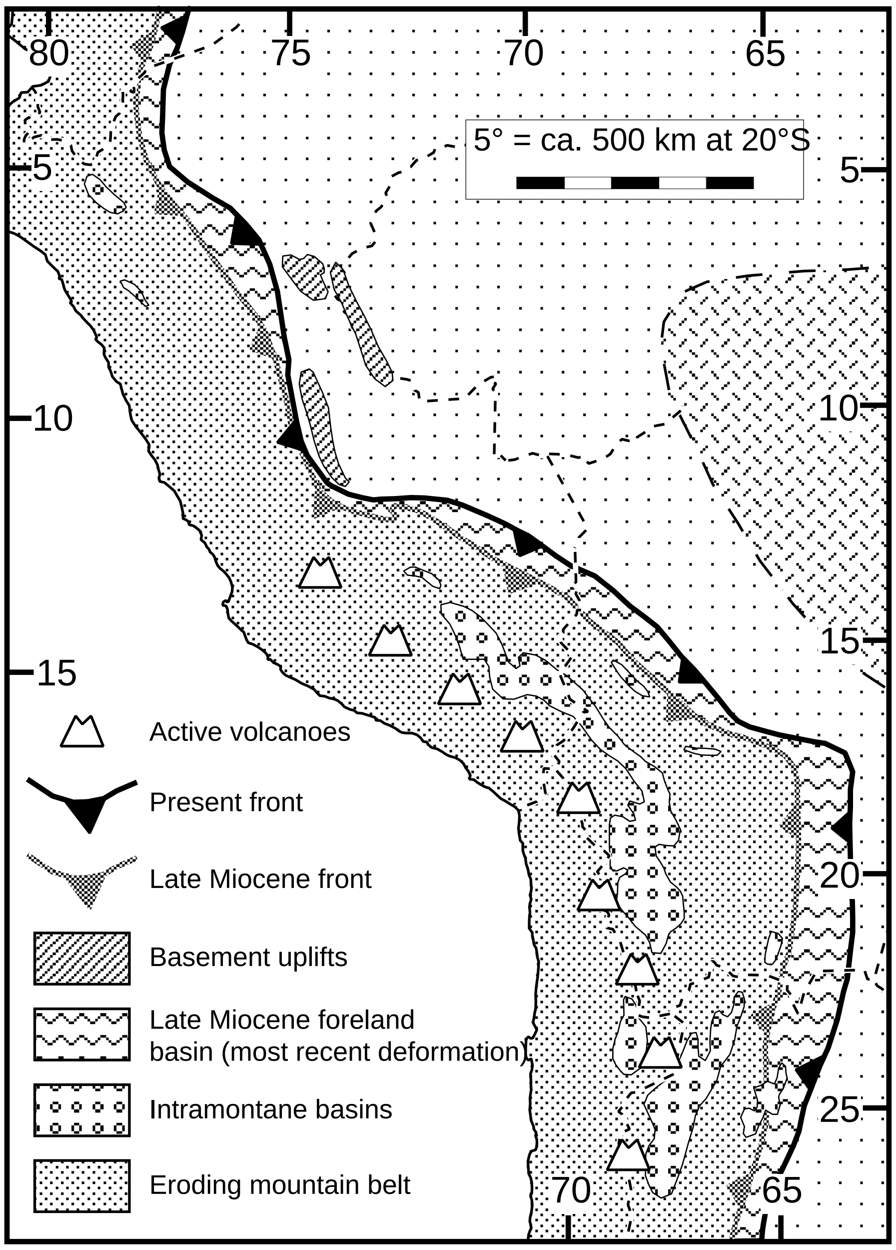


**Fig. 12**. (left) Late Miocene (ca. 10 Ma) to recent palaeogeography. By the late Miocene active volcanism became restricted to the western margin of the Andean plateau. The Eastern Cordillera was thrust eastward onto the Brazilian shield producing a deep flexural foreland basin. Between 10 Ma and 6 Ma shortening in parts of the Altiplano reduced considerably the areas still accumulating sediment and during the late Miocene to Pliocene deformation advanced into the foreland basin, producing an imbricate fold-thrust belt. Thin-skinned tectonics become less important north of the plateau, where thick-skinned uplifts occur. During the middle Pliocene much of the northern Andes and eastern plateau region became externally drained and the area of intramontane basins dropped.

**Fig. 13**. (below) Fission track ages versus normalised altitude for the Cordillera Real, Bolivia (data points from Benjamin et al., 1987 and Crough, 1983). Because mean altitude, and thus depth to a given geotherm, drops to the northeast along the sampling profile I have normalised altitudes to 4025 m - (depth below mean surface altitude). This results is a more conservative estimate of denudation rate and prevents an apparent exponential increase in recent denudation rates being inferred in the younger samples, which lie deepest below mean surface (e.g. see Benjamin et al., 1987 figure 3). Unroofing in the Cordillera Real started at about 50 Ma, during the inversion of Cretaceous depocentres. The dramatic increase in unroofing at ca. 25±5 Ma coincides with the main period of surface uplift in the central Andean plateau region but the deduced rates give no indication of rate of increase of mean surface elevation. The high Miocene to present unroofing rates in this region, which have exposed granites and andalusite-cordierite schists, contrast with the very low erosion south of 18°S where Cretaceous and Cenozoic sediments are widely preserved. Note that the youngest apatites eroding into the foreland at present are as young as 5-4 Ma, as a result of deep dissection.

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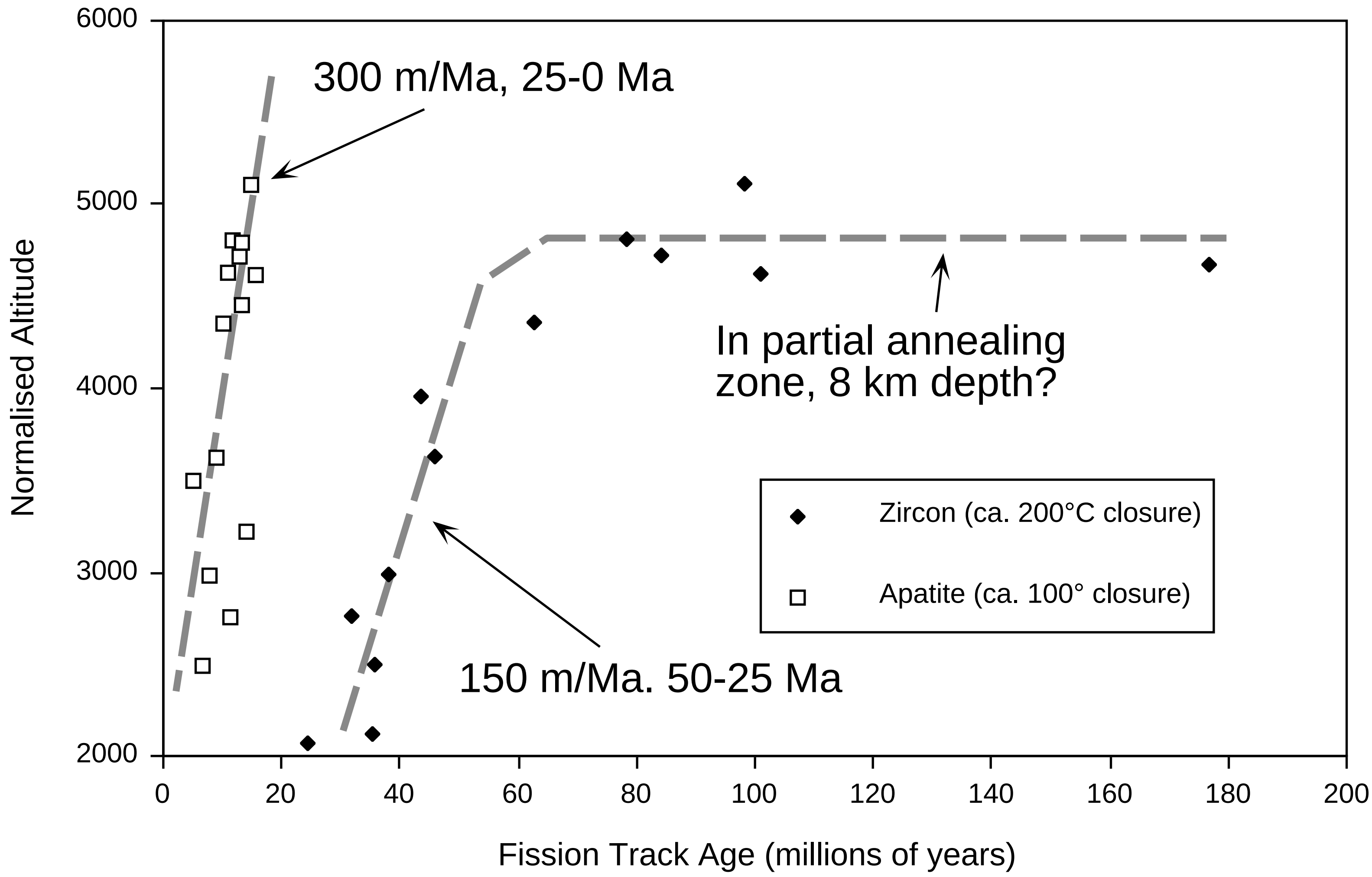

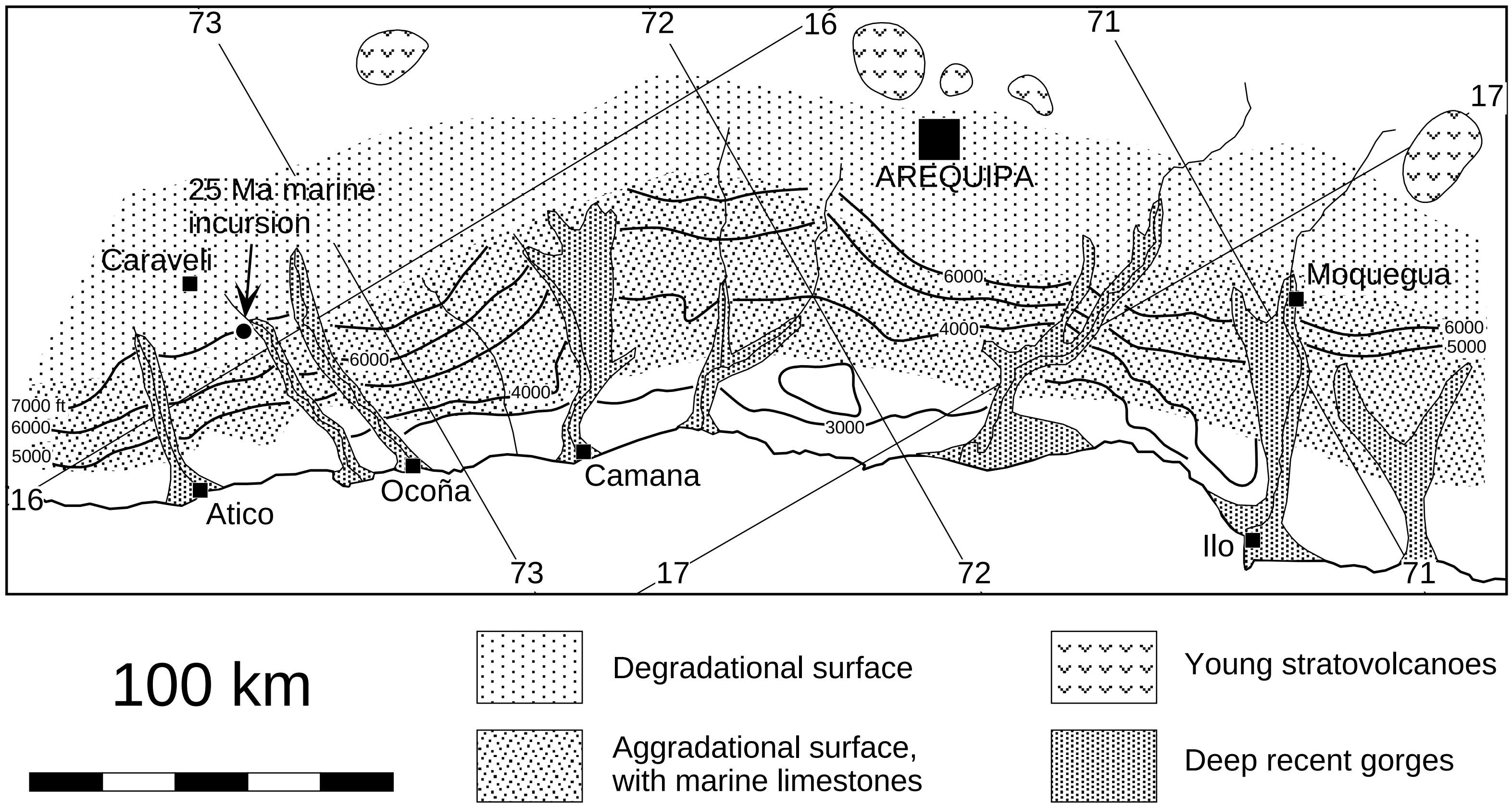


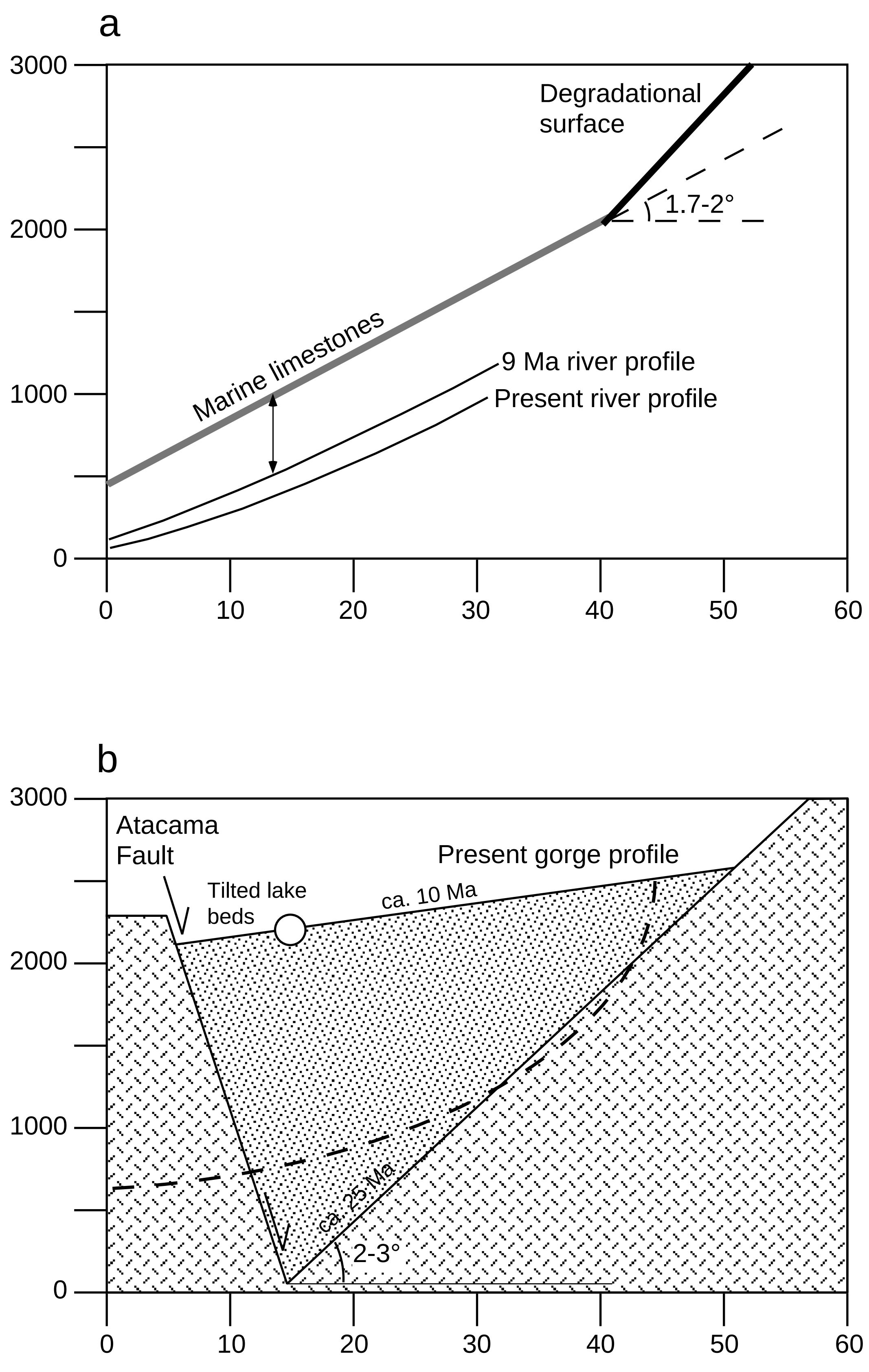


**Fig. 14.** (above) Map of the coastal region of southern Peru showing the known limits of a late Oligocene erosion surface transgressed by marine limestones. The contours (data source Operational Navigation Chart sheet P-26, scale 1:1,000,000, contours in feet) show the interpolated present elevation of the marine transgression, which is now tilted to the southeast at 2-3°. To the northeast a degradational surface rose to an altitude of perhaps 1500 m above sea-level and is the basement on which the Miocene arc erupted. Lower palaeosurfaces (not shown) and deep canyon incision have not removed enough material to account for more than a few percent of the uplift of the surface.

**Fig. 15.** (left) Schematic cross-sections illustrating inferred uplift magnitude and timing in western Peru and Chile. a, Moquegua region (see fig. 14) showing the attitude of the 25 Ma marine incursion surface and subsequent valley profiles (data from Tosdal et al., 1984). By ca. 9 Ma limestone outcrops, now at 1000 m, were elevated at least 500 m above dated river profiles near Locumba, suggesting that at least half the tilting predates ca. 9-10 Ma. b, Pampa de Tamarugal region, northern Chile. The Pampa overlies a half-graben basin, bounded by the Atacama fault. Tilting and sedimentation started at ca. 20-25 Ma as the Andes started to rise in this region. By ca. 9 Ma the basin was filled and drainage gained a direct outlet to the sea, initiating deep gorge dissection. Dissection was thus a result of base-level change and directly of uplift. In both regions Plio-Pleistocene lake terraces show evidence of continued tilting.

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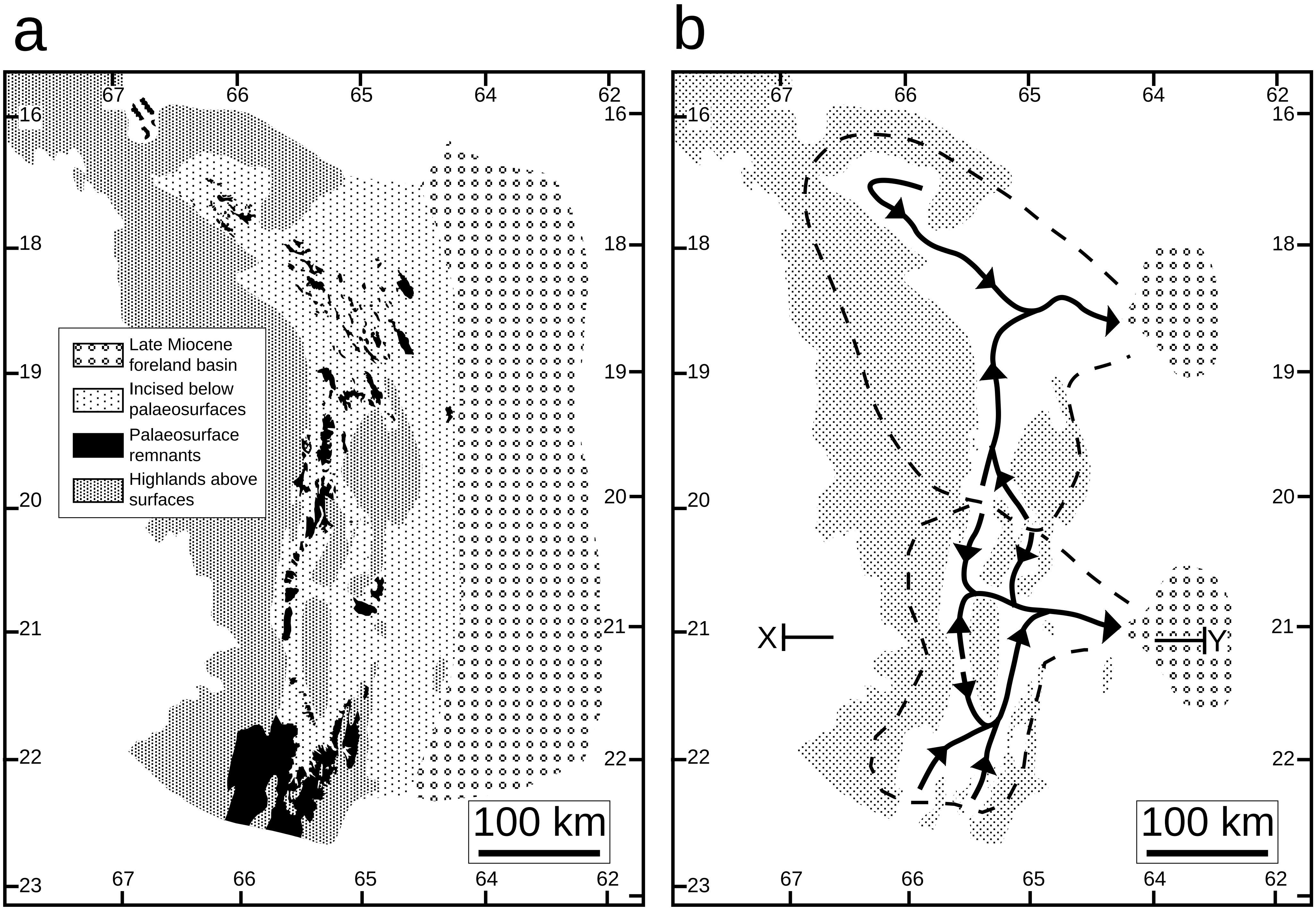


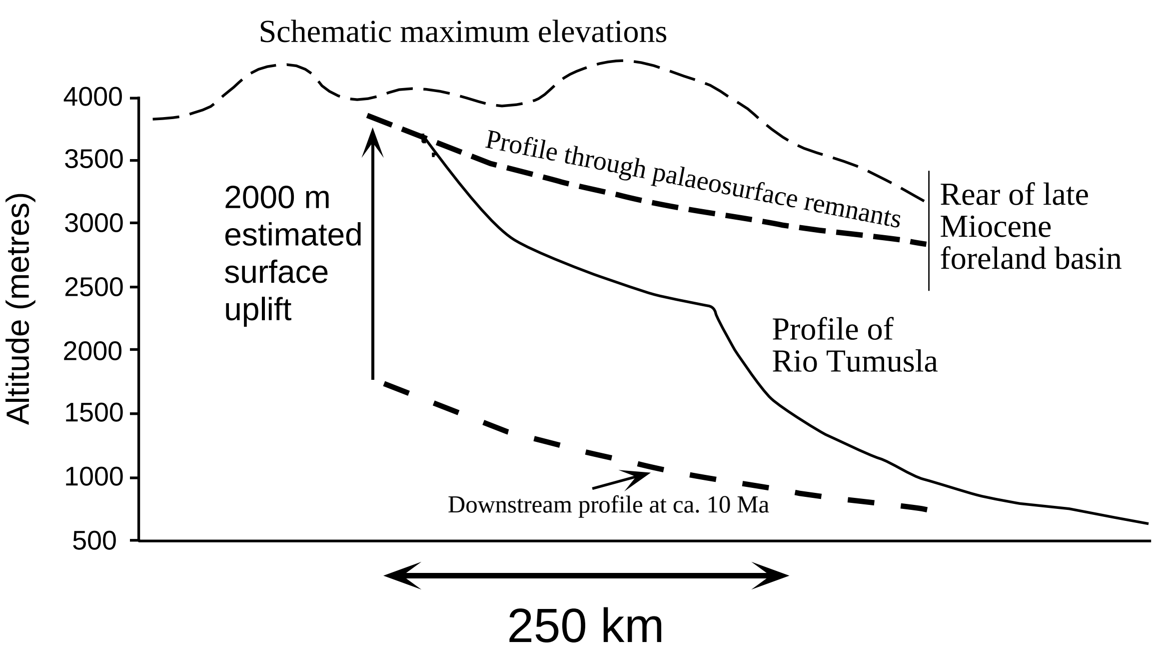




**Fig. 16.** (above) High altitude late Miocene palaeosurfaces in the Eastern Cordillera of south-central Bolivia. **a**, shows the location of all mapped surface remnants, areas of higher altitude and the position of the foreland basin at ca. 10 Ma, now the Subandes fold-thrust belt. **b**, interpretation of palaeodrainage basins with lines of drainage inferred from palaeosurface attitude and elevation. The palaeosurfaces define a mature low-gradient system of broad, flat-bottomed valleys which drained into the late Miocene foreland basin, which was at or near sea-level. Thus the present surface altitude is the result of ca. 2000 m of vertical uplift since ca. 10 Ma. The lack of significant regional tilts on any surface remnants suggests they are not underlain by a major crustal ramp active since the late Miocene, during Subandean thrusting.

**Fig. 17.** (left) Schematic cross-section through the high altitude palaeosurfaces (approximately X-Y on fig. 16b) showing the inferred uplift of the surfaces since 10 Ma. Also shown are approximate maximum elevations above the palaeosurfaces, suggesting that prior to ca. 10 Ma the Eastern Cordillera was probably not much higher than 1000-2000 m. The present drainage profile of the Rio Tumusla is shown. Although dissection is present all along the length of the river it is most prominent downstream of a major knickpoint, not coincident with a marked lithological change, which may be migrating upstream.

## UPLIFT OF THE SOUTHERN ANDES

South of the intersection of the Chile Ridge with the coast at 46°S there is a dramatic change in plate kinematics, geomorphology and geology (see figs. 1 and 2). The mountain belt is wider and higher than to the north, although only local peaks exceed 1500-2000 m. Core bedrock (the Mesozoic Patagonian Batholith extends from ca. 40°S to 54°S) is the same to the north and south of the change. Uplift and deformation can be dated from the Cenozoic marine to continental transition and from the youngest deformed rocks and clearly shows that deformation moved from south to north. In southernmost Chile, the Cretaceous Rocas Verdes back arc basin, floored with oceanic crust, was inverted between 100 Ma and 80 Ma (de Wit and Stern, 1978; Dalziel,1986). Initial unroofing rates were high but dropped by ca. 60 Ma (Nelson, 1982). The youngest deformation in the extreme south is Eocene (Winslow, 1981). In contrast, farther north significant deformation postdates 15 Ma marine rocks and very dramatic crustal shortening occurred in the late Miocene (Flint et al., 1994; Ramos, 1989). The highest marine rocks are found at ca. 500-1000 m above sealevel. Sediment age and subsidence rates in the Patagonian foreland basin show a similar north-south diachroneity (e.g. Biddle et al., 1986). There is little Cenozoic to recent arc andesite volcanism south of ca. 44°S (e.g. Thorpe et al., 1982). In foreland regions, however, alkaline plateau basalts are common, although absent to the north (e.g. Ramos and Kay, 1992). These are also younger in the north. The intersection of the Chile Ridge with South America has moved north through the Cenozoic (e.g. Cande and Leslie, 1986). The subcrust presence of an active spreading ridge may have resulted in higher heat flow and a softer upper crust allowing easier shortening although plate convergence rate is lower than to the north (e.g. Ramos, 1989). Overall crustal shortening estimates (ca. 50 km, Ramos, 1989) suggest a similar continental/subduction zone slip ratio to farther north.

## DISCUSSION

Andean topography is clearly correlated with a thick crustal root. Geophysical evidence (e.g. Watts et al., 1995; Zandt et al., 1994; Whitman, 1994; James and Snoke, 1994; De Matos and Brown, 1992; Introcaso et al., 1992; Schmitz et al., 1990; Kono et al., 1989; Götse et al., 1988; Wigger, 1988; Lyon-Caen et al., 1985; Wilson, 1985; James 1971) suggests that the crust under the highest parts of the Andes reaches 60-70 km thick, or nearly twice the thickness in fore-arc and foreland regions (35-40 km) and that large-scale Andean geomorphology is likely to be intimately linked with crustal thickening processes such as magmatic addition or horizontal shortening. Lavas younger than ca. 25 Ma show more crustal contamination than older lavas (Boily et al., 1990), suggesting that the crust through which they rose became markedly thicker from the early Miocene.

Widespread evidence of simple tilting or vertical uplift of erosion surfaces (figs. 14-17), and the preservation of flat-lying thick Cenozoic volcanic and sedimentary sequences over much of the Altiplano, with only localised shortening, has prompted many authors to suggest that uplift in the Andes has been caused by vertical block-movements (e.g. Myers, 1975; Cobbing et al., 1981) with the deep crustal root mainly a result of magma addition from below (e.g. James, 1971; Gough, 1973, Thorpe et al., 1981).

We can test this idea with a mass balance calculation for the central Andes at ca. 18°S where the total cross-sectional area is about 45-50000 km$^2$ and the trench to foreland width about 900 km (Kennan, 1994). This is an ideal area to choose because the cover of middle to late Cenozoic rocks over much of the plateau indicates that overall erosion rates are very low and that the topography is a primary signal of tectonic processes. The significant unroofing in the Cordillera Real of Bolivia accounts for less than 5% of the crustal cross-section and much of this may have been recycled into Altiplano or Subandes basins included in the calculation. A cursory examination of the digital elevation data (fig. 1) for much of the central Andes shows that deeply dissected regions are relatively narrow so the assumption of little overall erosion is reasonably valid.

The total volume of central Andean extrusive rocks is apparently relatively small. Although locally thick, total volume added is probably not more than 100 km$^3$/km arc length (Baker and Francis, 1978), mainly over the last 25 Ma. The volume of related cumulates left within the crust, due to high-level fractionation of primary basaltic melt, may be 2 to 4 times extruded volume (e.g. Thorpe et al., 1982, Boily et al., 1990) so extrusive volcanism can account for perhaps 500 km$^2$ of the cross-section. It is not so clear how to estimate the volume of magma intruded or accreted to the base of the crust. Magma addition related to widespread granite intrusion (fig. 18) may account for 4-8000 km$^3$/km arc length, including cumulates (estimate from Thorpe et al., 1981) but most were intruded between ca. 180 Ma and 80 Ma, during a major period of extension, high heat flow and space creation, and predate Andean surface uplift (e.g. Atherton, 1993; Petford and Atherton, 1992; Bonhomme and Carlier, 1990; Farrar et al., 1990; Kontak et al., 1990a; Soler and Bonhomme, 1988; Mukasa, 1986; Beckinsale et al., 1985; Noble et al., 1984; Lancelot et al., 1978; Capdevila et al., 1977). Assuming that magma production rates have been similar since then, perhaps 4000 km$^3$/km arc length basaltic magma has ponded at the base of the crust since about 50 Ma. Thus, total magma addition since 50 Ma may be ca. 5000 km$^3$/km arc length, accounting for only ca. 10% of the crust at 18°S.

Seismic velocities and densities in the lower crust (references given above) support the accretion of a basaltic root mainly beneath the Western Cordillera, while the lower crust in the Altiplano and farther east has a more felsic composition. He isotope studies, however, indicate that modest mantle melting is occurring right across the Altiplano, and that lithospheric mantle is thin in these regions (Hoke et al., 1994). Ponding of mantle melt near the base of the crust may have contributed to crustal melting and extrusion of peraluminous ignimbrites in this region.

At 18°S the shortening required to explain the observed crustal thickening is about 250-350 km depending on the initial crustal thickness (ca. 35-40 km). In addition to surface observations, seismic data (unpublished oil company data) shows significant shortening within apparently undeformed basins in the Altiplano. Altogether ca. 150 ± 30 km of shortening can be accounted for across the relatively gentle

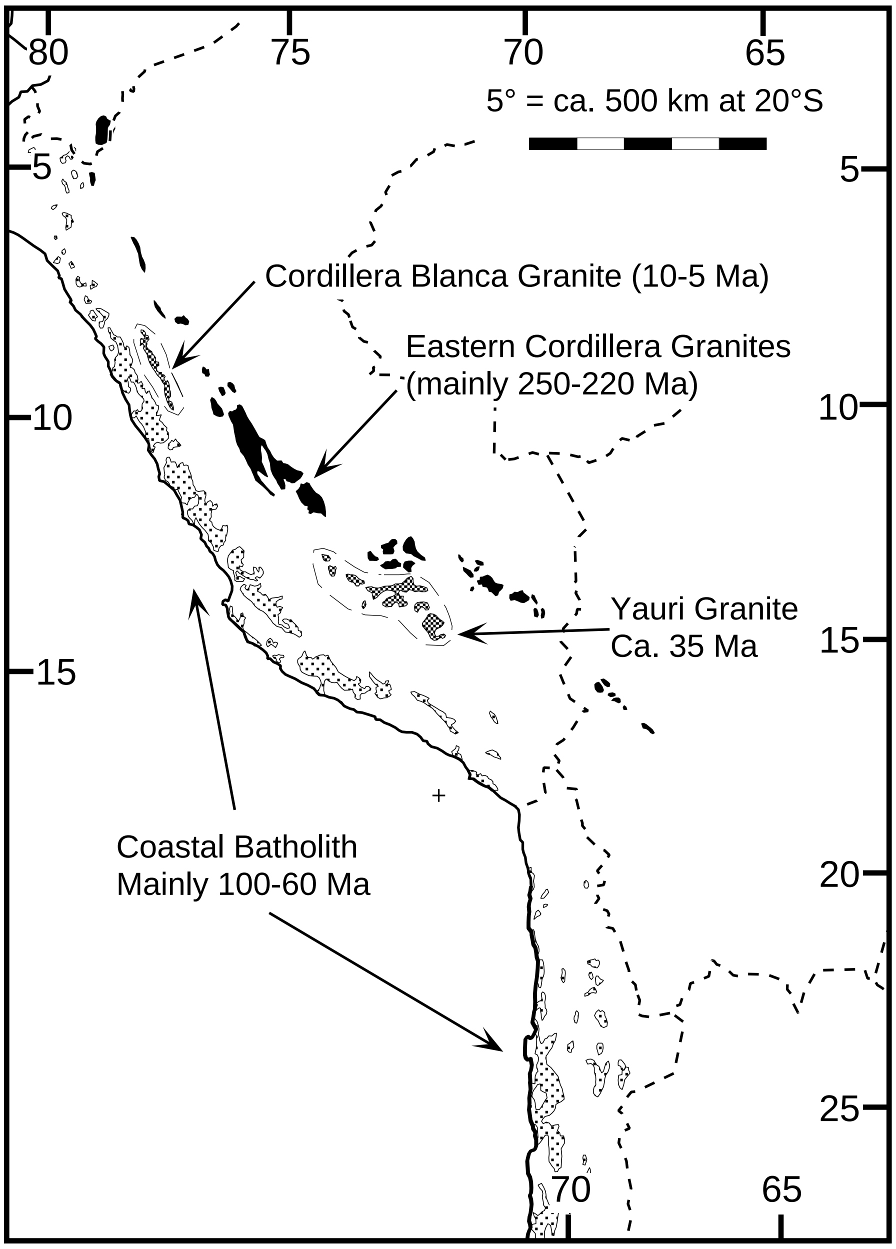
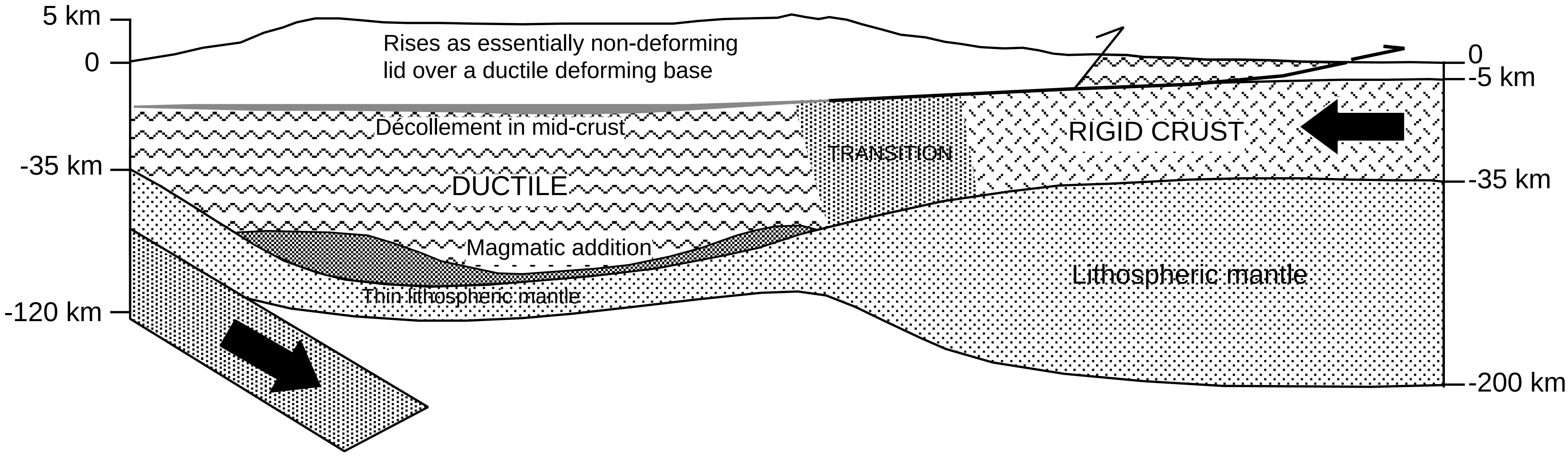



**Fig. 18.** (left) Granitoid bodies in the Central Andes (generalised from CGMW, 1978). Although large volumes of granite occur in the Central Andes most pre-date significant uplift. The Coastal batholith was mostly intruded between 180-60 Ma and the Eastern Cordillera granites are older at ca. 250-220 Ma. In both cases intrusion at a very high-level (8-4 km depth) accompanied major basin extension, injection of basaltic magma at a high level in the crust and high heat-flow. Almost no large granites have been intruded since the onset of compression, crustal thickening and uplift: later felsic intrusions are mainly small, high-level stocks. The exceptions are the late Oligocene Yauri batholith and the late Miocene Cordillera Blanca batholith. The former lies in a poorly known zone of complex strike-slip faulting which may have allowed space creation and the latter lies in the footwall of a major extensional/sinistral fault system. Because erosion is not sufficiently deep we do not know if granites are being intruded beneath the Miocene to recent arc or whether basaltic magma is ponding at the base of the crust due to a lack of space creation mechanisms.

**Fig. 19.** (below) Proposed Andean crustal thickening model (see Isacks, 1988). The apparent vertical uplift over much of the Andes is offset by about 3-500 km from the major shortening in the Subandes. The two are reconciled by using a mid-crustal décollement to transfer lower crustal shortening beneath the Andean plateau. This is consistent with numerical models of crustal shortening. Also limited seismic data show no sign of the Brazilian shield subducting steeply to the west in the mantle. In this model almost all the bulk of the crust can be accounted for using measured surface shortening, with perhaps 10% due to magma addition, mainly below the Western Cordillera volcanic arc.

*Lorcan Kennan, Andes Geomorphology PREPRINT, in press, Wiley.*



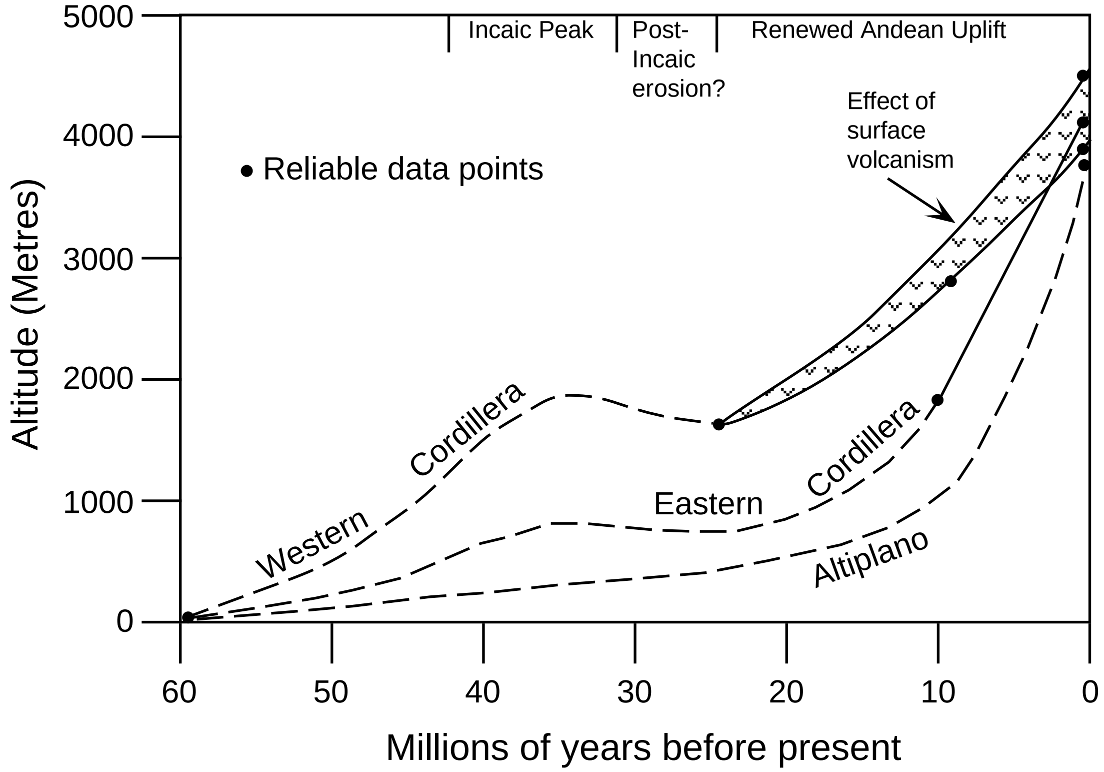
structures of northern Chile, the Bolivian Altiplano and the Eastern Cordillera at 18°S (Lamb et al., in press; Kennan, 1994) leaving perhaps 100-200 km which must be concentrated in the narrow Subandean fold-thrust belt in the east. Seismic reflection studies (unpublished oil company data) clearly show a well-developed thin-skinned style of deformation between 12°S and 25°S. Palaeozoic to Cenozoic sequences are imbricated above Lower Palaeozoic shale décollements and total shortening at ca. 18°S estimated at least 150 km (e.g. Baby et al., 1992; Roeder, 1988). The low altitude of the Subandes clearly indicates that large magnitude shortening of the middle and lower crust beneath the décollement is not occurring in that region. Instead, essentially undeformed Brazilian shield is being thrust beneath the Eastern Cordillera and Altiplano (fig. 19). Isacks (1988) has proposed that this produces distributed ductile shortening and thickening beneath the central Andean plateau, with passive uplift of a less-deformed upper crustal "lid". How much uplift occurs and when depends closely on the thickness and width of the deforming layer. As a preliminary estimate, the ca. 200 x 30 km of Brazilian shield thrust beneath the central Andes would thicken the crust beneath the ca. 400 km wide plateau by an average 15 km, which would produce approximately 2 km of late Miocene to recent surface uplift, consistent with geomorphological observations. This model is supported by gravity studies, which show clear flexure of the Brazilian shield beneath the load of the Eastern Cordillera (Watts et al., 1995; Lyon-Caen et al., 1985). Strong lithosphere extends 150-200 km east beneath the Eastern Cordillera. This underthrusting is constrained to have occurred since 10 Ma because only then did significant flexural subsidence start in central Bolivia. Allmendinger et al. (1990) report that at ca. 30°S, where late Miocene magmatic addition to the crust is negligible, measured crustal shortening can account for the entire observed crustal cross-section.

In addition to modest surface erosion, the subducting oceanic plate may scrape off lower crustal slices from the fore-arc region (e.g. Karig, 1974), with this crustal thinning causing dramatic subsidence of fore-arc basins below sea-level (e.g. von Huene and Lallemand, 1990). Estimated maximum erosion rates are ca. 100 $km^3$/km arc length (Sosson et al., 1994) but it remains unclear whether this material is accreted to the base of the crust deeper down (resulting in high pressure metamorphism) or carried deep into the mantle. The error introduced in the shortening calculations above is unlikely to be greater than ca. 10%.

Although crustal shortening explains the crustal volume, it does not explain why shortening is distributed as it is or why the result is plateau-shaped. Numerical modelling of Andean type scenarios provides some insights into possible preconditions for the formation of a plateau region with steep margins (e.g. Harry et al., 1995; Wdowinski and Bock, 1994a,b). In order to focus deformation relatively close to the subduction zone, a thermal anomaly and/or a thick upper crustal sedimentary column are required, both consistent with the observation that early Andean shortening was focused into regions of Mesozoic extension. Assuming initially thickened or uniform crust produces a much broader, wedge-shaped orogen. The presence of strength discontinuities in the models also results in décollement formation as shortening increases, so that maximum shortening in the upper crust shifts towards the foreland while lower crust and mantle shortening remains closer to the trench. This results in the offset of regions of maximum upper crustal shortening and surface uplift as seen in the Andes.

Surface uplift in the central Andes started at different times and proceeded at different rates, and uplift has not been "plateau-like" (fig. 20a,b). However, through a combination of shortening and magmatic addition (Western and Eastern Cordilleras) and locally thick sedimentation (Altiplano) the region seems to have converged towards the plateau height of ca. 4000 m relatively recently (?<5 Ma). Numerical models also show this convergence, with the plateau height depending on the strength of the thickened crust. Essentially vertical buoyancy forces reach a balance with horizontal compressional forces (e.g. England, 1996; Froidveaux and Isacks, 1984) and additional shortening is accommodated by widening of the plateau region. If bounding horizontal stresses drop plateau collapse towards a slightly lower equilibrium height may occur, and this may account for observed high-altitude normal faulting in north central Peru (e.g. Cordillera Blanca fault, Deverchère et al., 1989; Dalmayrac and Molnar, 1981. Quiches fault, Bellier et al., 1991; Doser, 1987) where average altitudes are slightly higher than in the plateau region to the south. Boundary stresses are likely to depend on plate convergence rate, subduction zone dip, age of subducting ocean crust and other factors. The mechanisms for maintaining a plateau region with constant height despite remarkable differences in observed surface shortening and thickness of young sediments remain unclear. Surface smoothing mechanisms such as landsliding, erosion and redeposition, extensional faulting (e.g. Green and Wernicke, 1986) and some adjustments in thickness of the crustal root must be occurring.

Plateau evolution appears also to be intimately linked with the development of the "Bolivian Orocline" (fig. 21). Isacks (1988) has pointed out that crustal cross-sectional area drops to north and south, possibly as a result of reduced shortening. Greater shortening in the core of the orocline could result in fore-arc rotations of 10-15°, anticlockwise in Peru and clockwise in northern Chile, which are consistent with palaeomagnetic data (e.g. Macedo-Sanchez et al., 1992). This is consistent with the relatively few accurate shortening estimates across the central Andes. Subandean shortening of >150 km at ca. 18°S drops to ca. 70 km at 22°S (Baby et al., 1992) and perhaps as little as 15-30 km at ca. 12°S (author's own field observations). It is still not clear why shortening should be more intense in the core of the orocline. There is a close spatial correlation with the dip of the subduction zone. The restriction of the volcanic arc to areas of steep subduction dates from the late Miocene, and may date the shallowing of the subduction zone (e.g. Jordan et al., 1983). However, this is exactly the time of intensified shortening and it is not clear which is cause and which is effect. There is also a spatial correlation with better developed shale décollements in the core of the orocline (e.g. Allmendinger et al., 1993; 1983) which may allow greater thin-skinned shortening. Recently, Watts et al. (1995) have pointed out a strong correlation with a much larger scale, and longer lived, feature: lithospheric strength. Studies of flexure clearly show that the lithosphere is strongest in the core of the orocline at 18°S. Possibly, strong lithosphere is a prerequisite for developing a thin-skinned belt which can take up large-amounts of shortening. In contrast, weaker lithosphere leads to steeper décollement angles, and thick-skinned structures, which cannot take up



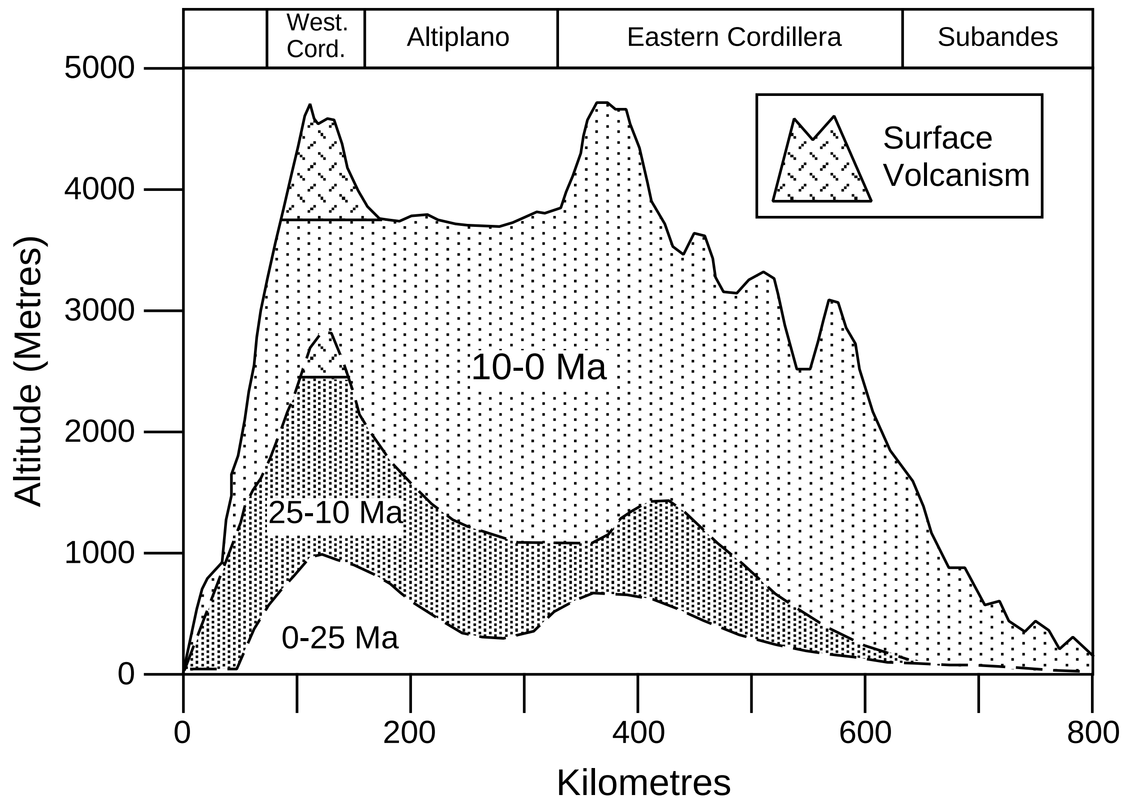




**Fig. 20.** (above) Central Andean uplift in space and time. **a,** (top left) Estimated elevation through time for the Western Cordillera, Altiplano and Eastern Cordillera, showing reliable data points. **b,** (top right) schematic west-east cross-section across the Andes at ca. 18°S. The few available data suggest that Andean uplift was no “plateau-like” and that different regions uplifted at different times, only recently converging towards ca. 4000 m. This altitude may represent a limit imposed by overall lithospheric strength.

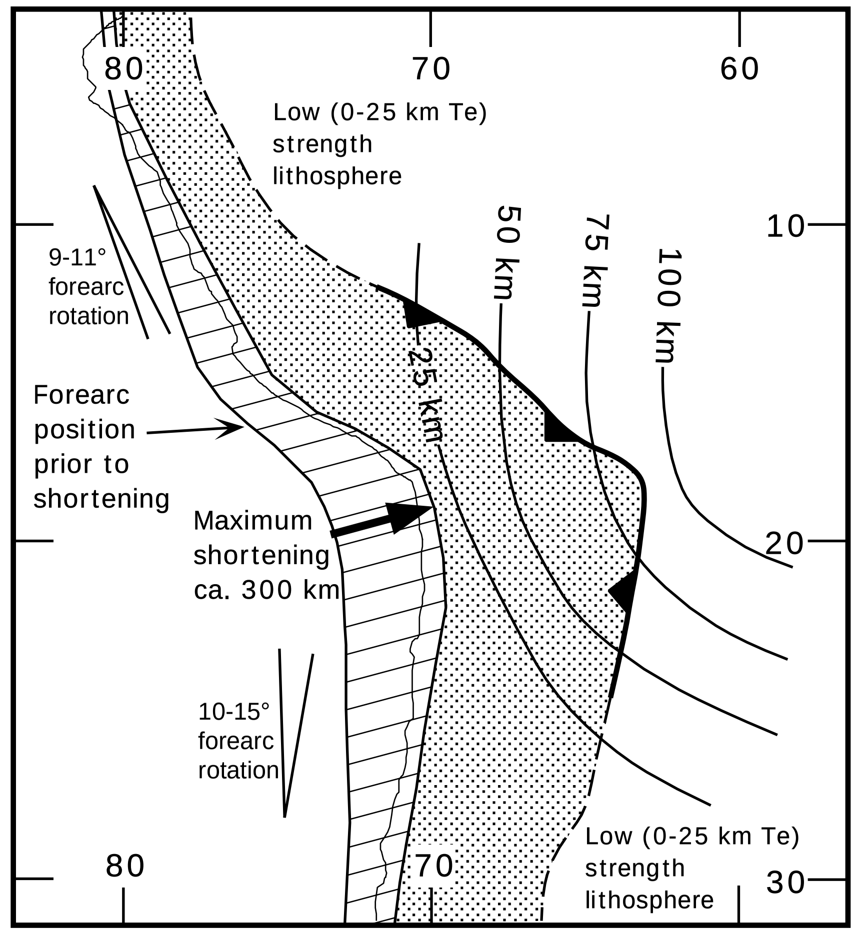



**Fig. 21.** Outline map of the central Andes showing interpreted forearc position prior to significant shortening. Because observed shortening appears to increase dramatically into the core of the Bolivian Orocline forearc rotations have rotated, anticlockwise in Peru and clockwise in north Chile. The area “removed” by shortening is shown hatched. The increased shortening is thus responsible for both the curvature of the Orocline and for the central Andean plateau. The contours east of the Andean thrust front show elastic thickness of the lithosphere (after Watts et al., 1995). There is a strong coincidence of strong lithosphere with well-developed foreland fold-thrust belts (barbed line) and wide foreland basins. To north and south, where Te is 0-25 km, foreland basins show seep narrow depocentres close to the mountain front and basement cored uplifts which accommodate little shortening dominate foreland structures.

large magnitude shortening.

I have already pointed out that an understanding of climate change through time is vital in order to correctly interpret evidence of palaeoflora change and geomorphological features such as dissection. Cause-effect links between climate and Andean tectonics may be important but are as yet poorly known. Climate simulations (e.g. Lenters and Cook, in press) suggest that Andean topography plays a critical role in concentrating precipitation on the Andes' eastern slopes and keeping regions to the west arid.This concentration of precipitation may be why, in contrast to other parts of the Andes, these regions are very deeply unroofed. For instance, from ca. 0°S to 18°S Precambrian and Palaeozoic metamorphic rocks have had up to ca. 8 km of cover removed during the Cenozoic, and significant shortening appears to have been focused into these regions. Masek et al. (1994) point out the close coincidence between this unroofing and present day precipitation rates and note that, in addition to focusing deformation and unroofing into narrow regions, deeply cut valleys at the margin of the central Andean plateau will lead to some isostatic rebound, and curling up of the lip of the plateau, raising the peaks of the Cordillera Real. To the south of 18°S the climate is more arid, dissection is much lower, and without this isostatic rebound the plateau edge drops more gently to the Subandes. Peaks raised in this manner may become glaciated and erosion and discharge patterns may change. This would not, however, indicate raising of mean elevation above the snowline and illustrates that major mean elevation change, and time of change cannot be inferred from first appearance of local glaciation.

Studies of foreland basin sequences and how they relate to erosion in the hinterland may play a vital role in understanding large-scale morphology of the Andes, in the absence of detailed palaeoaltitude data. For instance, Kooi and Beaumont (1996) have pointed out that mountain belt morphology is strongly sensitive to time lags between uplift and erosion processes and it seems unlikely that Andean uplift and erosion have reached any kind of steady state. Thus flexural subsidence, due to the load of the orogen on the Brazilian Shield, and sediment supply were probably not in equilibrium. Burbank (1992) has suggested that if loading outstrips sediments supply, deposition will migrate intowards the mountain front, and if erosion becomes dominant mean elevation and flexural loading may decrease, resulting in basinwide rebound and unconformities. Although a lot of seismic data exists for the Andes, no-one has yet tested if this has happened in the past. It may be happening at the present. A brief examination of LANDSAT MSS images held at Oxford shows that almost the entire Peruvian foreland region is slightly dissected, and that sediment is being eroded to further downstream in the Amazon Basin (see also Rasanen et al., 1992). Only close to the thrust front in northern Bolivia (14°-18°S) is there evidence of foreland subsidence. Here, meanders in tributaries of the Beni river migrate into the mountain front and there are enormous flooded areas (Pete Burgess, pers. comm.). This is the more intensely shortened part of the Cordillera Real and Subandes, where the topographic load is higher and narrower, where there is a deeper foreland basin and steeper basal décollement.

In addition to the provenance studies of foreland basin sediments outlined above, fission track dating of detrital grains may also give important clues to palaeorelief in the orogen (see Cerveny et al., 1988 for an example from the Himalayas). Consider the Cordillera Real of Bolivia (see fig. 13). Here, late Cenozoic unroofing rates of ca. 300 m/Ma suggest it would take apatites about 10-13 Ma to reach the surface after track annealing stopped, if there was no dissection. However, dissection digs well below the mean surface, and as a result the youngest apatites being released into the foreland are 5-4 Ma old. This time lag is proportional to the depth of relief. In older foreland basin sediments we may also expect to see a time lag between depositional and detrital apatite age. If unroofing rates can be well constrained in source regions it should be possible to directly estimate palaeorelief, which in turn yields a minimum estimate for maximum palaeoaltitudes. This requires a careful integration of fission-track dating, provenance and regional structural studies.

## CONCLUSIONS

1) The large-scale geomorphology of the Andes, especially the central Andes, reflects crustal thickening, mainly the result of crustal shortening. Although surface volcanism is widespread and defines prominent smaller scale geomorphological features, the contribution of magmatism to crustal thickening is relatively small.

2) The earliest Andean uplifts were defined by major terrane boundaries or by inverted Mesozoic extensional basins, indicating that preexisting weak-crust and higher heat-flow was probably important. Continued deformation merged these early uplifts into a single mass in the central Andes, while northern Andean uplifts have remained separated by externally drained intermontane basins.

3) Over much of the Andes, net Cenozoic erosion rates have been relatively low. Large-scale topography represents the primary signal of crustal thickening and can be modelled successfully using numerical methods in which erosion is not considered. Initially thinned crust with a weak sedimentary cover, with modest thermal anomaly, appears to be a prerequisite for plateau development. The models also predict that strain in the crust is vertically partitioned by major décollements which allow the sites of major surface shortening and of surface uplift to be separated, accounting for the apparent vertical uplift seen in the interior of all Andean ranges.

4) The mechanical properties of the Precambrian shield regions to the east of the Andes also seem to be important. In areas of high lithospheric strength, such as Bolivia, a foreland fold and thrust belt with major shortening above low angle décollement, can develop. Where the lithosphere is weak, thick-skinned basement uplifts, which can accommodate little shortening, form.

5) In the central Andean plateau, poorly understood processes of lithosphere thinning have maintained a thermal anomaly to the present. He isotope studies suggest that mantle melting is occurring beneath the whole plateau region. Higher heat flow may be aiding lower crust or lithosphere flow, focusing high shortening into the region.

6) As shortening increased, separate areas uplifting at different rates converged towards a plateau height of ca. 4000 m which appears to be be an upper limit defined by crustal strength.

Localised drops in bounding horizontal stresses in northern Peru and northern Argentina may have allowed some collapse of over high topography. The reason for the stress drop is not clear.

7) Erosion plays an important role in cutting smaller scale (wavelength 10-100 km) topographic features. Over large areas of the Andes erosion is concentrated in narrow valleys and volume removed is unlikely to cause much isostatic rebound at typical lithosphere elastic thicknesses of 25-75 km. The timing of this erosion seems to coincide with a major global? climate change. Uplift may have occurred well before dissection, which may be a result of increased precipitation.

8) However, climate models suggest that growing Andean topography may have focussed rainfall in the Andes eastern flank. By ca. 10 Ma, the Andes may have been high enough to cause the onset of aridity on the plateau and in the coastal deserts of Peru and Chile. Thus there seems to be a positive feedback between uplift and concentration of precipitation which may lead to more efficient dissection.

9) The focusing of precipitation and dissection on the Andes eastern flank north of ca. 18°S may have kept topographic gradients steep in this region, enhanced peak uplift of the Cordillera Real in Bolivia, and even focused increased long term (ca. 10-20 Ma) deformation into a narrow region.

10) Most of the links proposed between observed deformation, crustal structure and history, geomorphology and climate remain somewhat speculative. Any approach to understanding Andean topography needs to be multidisciplinary, and to be aware of potential shortcomings of any one line of evidence. For instance, climate change, palaeoflora change and mountain uplift or unroofing are not independent of each other. Key issues for future research also include better constraints on surface and deep crustal structure, and surface field studies can define how important are surface smoothing mechanisms such as sediment redistribution, or local faulting. Fission track studies of dissected mountain regions and basin sediments may give direct independent estimates of palaeorelief and thereby constrain altitude, but inferred unroofing rate does not relate directly to surface uplift rate. Analysis of Andean digital terrain models may provide important insights in lithosphere scale processes and smaller scale geomorphological processes. Analysis of foreland basin seismic data may reveal basin-wide rebound unconformities related to periods during Andean uplift when unroofing was greater than rock uplift, and the load on the flexed shield was decreasing.

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