Erosion surfaces and ignimbrite eruption, measures of Andean uplift in northern Peru

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Remnants of five mature erosion surfaces occur on the rugged western flank of the Andes, and each is deeply incised. They indicate alternation of arid with more humid conditions, and that the western Andes were raised in five major pulses from a subdued early Tertiary landscape of volcanic flows. Each pulse of uplift was accompanied by south-westward tilting of the older erosion surfaces. The surfaces developed before an ignimbrite was erupted down the ancestoral *Rio Fortaleza* valley 6 m.y. ago. Since then, renewed uplift has led to further incision of the main valleys and the formation of canyons.

1. Introduction

The Andean mountain chain of Peru is divided into western, central and eastern ranges, called the Cordilleras Occidental, Central and Oriental. In northern Peru, the Cordillera Occidental is itself divided by the longitudinal valley of the Rio Santa into the Cordillera Blanca, which forms the Andean crest with peaks over 6000 m in altitude, and the Cordillera Negra to the west with an average summit level of 5000 m (Fig. 1). The top of the Cordillera Negra is a gently undulating landscape made up of dissected remnants of an old erosion surface, and its western flank is a rugged scarp which descends steeply in a multitude of sharp ridges and deep narrow valleys to the Pacific Ocean. The climate is arid on the western flank of the Cordillera Negra and remnants of five mature erosion surfaces are preserved which provide reference surfaces for measuring pulses of Andean uplift. This paper describes the sequence of erosion surfaces in a sector of the western flank of the Cordillera Negra between latitudes 10° and 10° 30′ south (Fig. 1). It demonstrates how uplift occurred and that the Andes in northern Peru were already established as a major mountain range before the end of the Miocene.

The region is composed of plutonic and volcanic rocks of the Coastal Batholith and its envelope. Geological structures trend NW–SE, parallel to the coastline and the mountain ranges of the Andes. The batholith occupies a belt between 10 and 60 km inland from the coast and consists of a number of plutons which range in composition from gabbro to granite (Myers 1975a). Both individual plutons and the batholith as a whole have steep walls and flat roofs. The batholith is flanked by volcanic rocks which it intruded and which were erupted from magmas rising slightly in advance of the

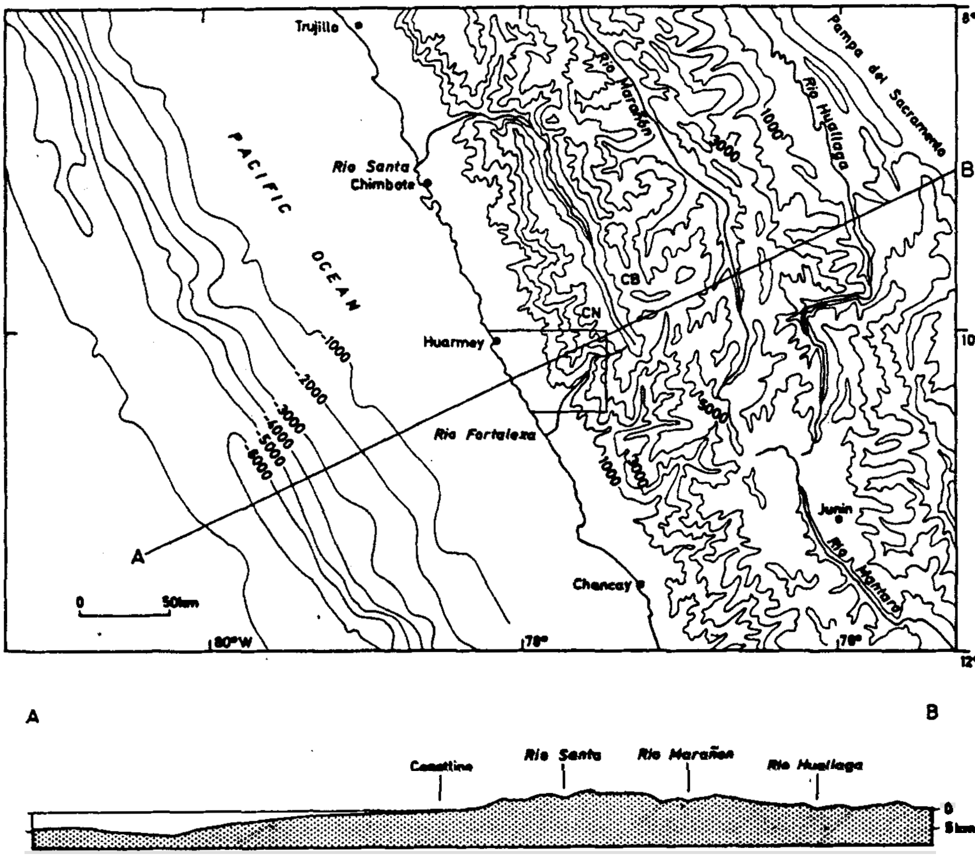


Fig. 1. Topographic map and section of part of the Andes in northern Peru, and submarine contours showing the Lima Trench. The area described is outlined. CN—Cordillera Negra, CB—Cordillera Blanca, AB—line of section. Contours are at 1000 m intervals; horizontal scales of map and section are equal.

main plutons of the batholith. The stratigraphic ages of fossils interbedded with the volcanic rocks (Myers 1974) and the isotopic ages of minerals from the batholith (P. A. Wilson personal communication) suggest that the volcanic rocks range in age from about 105 to 30 m.y. and the batholith from about 100 to 30 m.y. The oldest volcanic rocks represent submarine eruptions whereas the younger volcanic sequences are terrestial deposits. These rocks indicate emergence of the landmass during the intrusion of the batholith, and the absence of younger marine deposits suggests that the region then remained above sea level to the present day.

The older volcanic rocks were folded before most of the batholith was intruded, and most folds are open and have steep axial surfaces which trend NW–SE. The volcanic rocks were most strongly deformed in belts of tight and isoclinal folds on both flanks of the batholith. Regional metamorphism of the volcanic rocks was generally in zeolite facies, but greenschist facies assemblages developed syntectonically in the belts of tight folds which flank the batholith. The younger metamorphic aureole of the batholith is narrow and is in hornblende hornfels facies. Joints are numerous and most are steep and strike either NW–SE, NE–SW, N–S or E–W, but there are few faults in the region (Myers, in press).

Dollfus (1965) described the general geomorphology of a section across the whole Andean mountain range immediately to the south of this region, and Dresch (1958) gave an outline of the geomorphology of southern Peru. Notes on the geomorphology of smaller regions were also given by Jaén (1965) and Mendívil (1965) in southern Peru, and by Wilson *et al.* (1967) and Cobbing (1973) in northern Peru. Most of these authors considered that the various altitudes of the erosion surfaces were caused by tectonic uplift and that the greatest uplift occurred in the Plio-Pleistocene. Garner (1959) discussed the effects of climatic variations on the geomorphology of the Peruvian Andes.

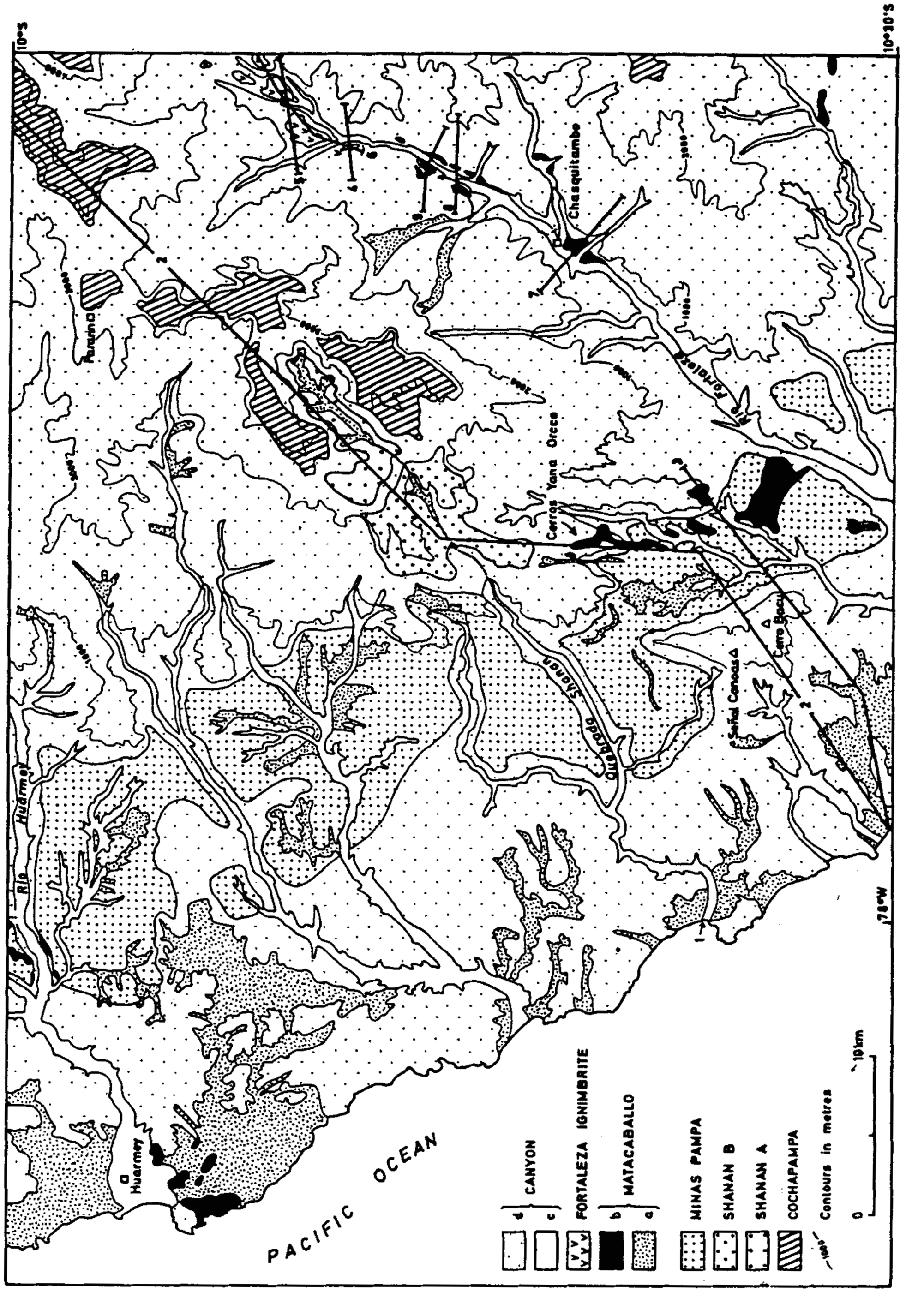
The pioneer work of McLaughlin (1924) in central Peru forms the basis of present knowledge of the erosion surfaces in that region and of their relationship with Andean uplift. McLaughlin (1924) recognised that the gently undulating topography of many high regions of the Andes represented an old erosion surface which predated elevation of the modern Andes, and he called this the Puna surface. McLaughlin (1924) described how the Puna surface was elevated slightly and then suffered a prolonged episode of erosion, the Junin erosion stage, which formed the broad pampa of Junin (Fig. 1) and similar high pampas by the filling in of depressions on the undulating Puna surface. More marked uplift followed, and valleys were incised to a depth of 300–600 m below the Junin surface during the Chacra erosion stage. During a third major episode of uplift, which McLaughlin (1924) called the Canyon erosion stage, these valleys were further incised and became narrow canyons with remnants of the former broader valley floors preserved as benches along their sides.

2. Erosion surfaces

2a. Cochapampa surface

The Cochapampa surface is a dissected undulating pediment which occurs between altitudes of 2800–3500 m, and summits to the north and east which rise to altitudes of 4600 m (Fig. 2). The surface lies entirely on volcanic rocks which form the roof of the Coastal Batholith. The folded volcanic rocks are Cretaceous to Palaeogene in age; they are truncated by the Cochapampa surface and are deeply weathered. To the northeast, the Cochapampa surface is continuous with concordant summit levels of the Cordillera Negra, and so it may be part of the extensive Puna surface of McLaughlin (1924). It is the oldest landform of the region.

The Cochapampa surface slopes southwestwards from a general altitude of 4300–4600 m in the northeast part of the region. This can be seen by plotting the heights of points on this surface against their distance from the coast measured along a line perpendicular to the coastline (Fig. 3 section 2). The summits of Cerro Bacu (1500 m) and Señal Canoas (1477 m) which are the highest points on a ridge of coastal hills, lie on a line which is the projection of this sloping surface, and they may also be remnants of the Cochapampa surface (Fig. 2 and Fig. 3 section 2), although this surface may only have approximated to a straight line. The gradient of the Cochapampa surface is 1 in 20 and if the surface formed with a sub-horizontal attitude then the gradient implies that the northeastern part of the region was elevated more than the southwestern part. Major faults parallel to the Andes which could have broken up and raised the Cochapampa surface in a number of segments are lacking in the deeply dissected western flank of the Andes in this region. Therefore, in this sector of the western Andes the Cochapampa, ?Puna, surface was raised in one piece by westward tilting rather


10°S
10°30'S
77°30'W
78°W
PACIFIC OCEAN
Huarmey
Rio Huarmey
Chasquitambo
Rio Fortaleza
Cerros Yana Orcco
Quebrada Shanan
CANYON
FORTALEZA IGNIMBRITE
MATACABALLO
MINAS PAMPA
SHANAN B
SHANAN A
COCHAPAMPA
Contours in metres
0
10 km

than by level uplift in fault-bounded blocks as McLaughlin (1924 p. 624) tentatively suggested.

2b. Shanan A erosion surface

The Shanan A erosion surface occurs between altitudes of 2300–2800 m as a broad pediment. It is mostly overlain by thick alluvial deposits, locally derived from volcanic rocks forming the roof of the Coastal Batholith. Some pebbles in these deposits are derived from volcanic formations which occur to the northeast, partly below the Cochapampa surface, and some of these alluvial deposits may therefore be derived from the weathered deposits of the Cochapampa surface. The Shanan A surface indicates a first uplift of the Cochapampa surface of about 500 m (Fig. 2 and Fig. 3 sections 1 and 2).

2c. Shanan B erosion surface

This forms an undulating sloping pediment between altitudes of 1700–2200 m (Fig. 2 and Fig. 3 sections 1 and 2). It is bounded to the east by a scarp which rises steeply to the Shanan A surface, and to the west by another scarp between altitudes of 1200–1800 m through which Quebrada Shanan descends in a series of precipitous dry waterfalls. This erosion surface marks a pause after a second uplift of the Cochapampa surface by another 500 m.

2d. Minas Pampa erosion surface

The Minas Pampa surface is a moderately dissected pediment which forms a strip of low relief, parallel with the coastline, between 15 and 25 km inland from the coast (Fig. 2 and Fig. 3 sections 2 and 3). It mainly occurs between altitudes of 600–900 m, except in the north where it slopes towards the Rio Huarmey valley and lies between altitudes of 300–600 m.

The Minas Pampa surface developed on the western part of the Coastal Batholith as a longitudinal plain, parallel with the batholith and ranges of the Andes. It is bounded to the west by a range of hills (Fig. 3 sections 2 and 3) which rise to an altitude of 1500 m. These hills are made up of relatively fine-grained volcanic rocks which are more resistant to weathering and erosion than the coarse-grained tonalite, granodiorite and granite of the Coastal Batholith.

The Minas Pampa erosion surface shows the first major development of subsequent drainage in the region. Prior to this time the main drainage was consequent, flowing southwestwards, normal to the slope of the tilted Cochapampa surface, with tributary streams following the major joint pattern. The rocks on which the Cochapampa surface developed were almost all volcanic and showed little regional variation in resistance to erosion. During the Minas Pampa erosion stage, the fairly flat top surface of the batholith was exposed for the first time in the west, and the coarser-grained plutonic rocks were more rapidly eroded than the volcanic rocks.

Weathering of the bedrock of the Minas Pampa surface is relatively deep, it is second in amount to the bedrock weathering of the Cochapampa surface, and indicates a major pause in uplift at this time, after a pulse of marked uplift and arid aggradation.

Fig. 2. Map showing the distribution of erosion surfaces, Cochapampa, Shanan A, Shanan B, and Minas Pampa; alluvial deposits, a. sand and silt, and b. gravel, of the Matacaballo erosion stage; the 6 m.y. old Rio Fortaleza Ignimbrite Formation; and Canyon erosion stage valleys c. and eroded hillsides d. The lines of topographic sections 1–9 of Figs 3 and 4 are also shown.

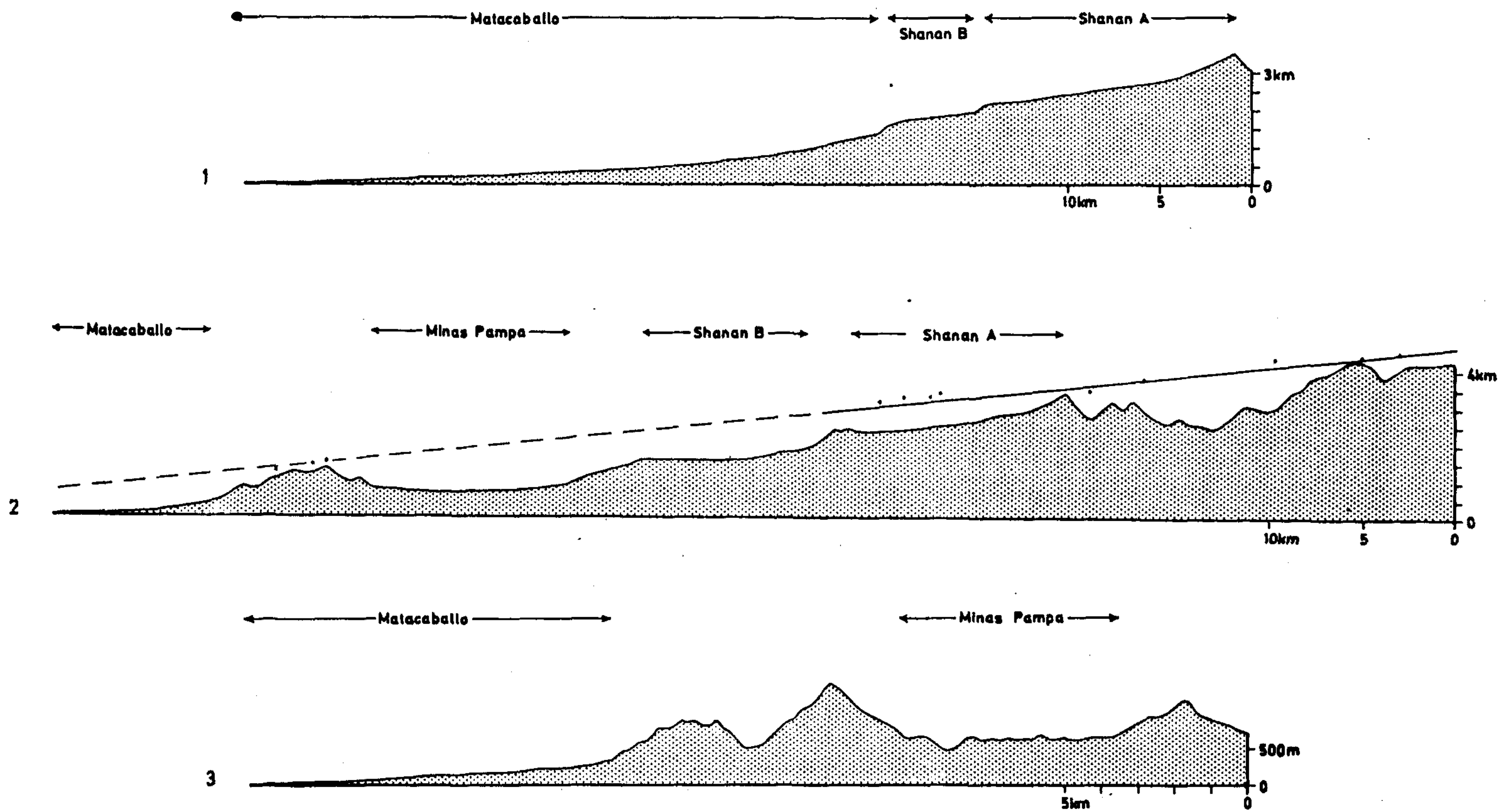


Fig. 3. Topographic profiles, 1. of Quebrada Shanan, and 2 and 3 along lines shown on Fig. 2, showing the Matacaballo, Minas Pampa, Shanan A, and Shanan B erosion surfaces. The line above profile 2 may approximately indicate the slope of the extensively eroded Cochapampa erosion surface, marked by summit levels shown as dots, projected along paths normal to the profile.

2e. Matacaballo erosion surface

The Matacaballo erosion surface is a moderately dissected strip of low relief parallel with the coastline. It extends from the coast inland for up to 10 km and mostly lies between altitudes of 50–200 m (Fig. 2 and Fig. 3 sections 1, 2 and 3). Much of the coastline is bounded by joints and faults which are still intermittently active and abruptly truncate the Matacaballo surface. To the west, the gentle slope of the continental shelf may represent a downfaulted extension of the Matacaballo surface. The Matacaballo surface slopes southwestwards from the foot of the range of hills which form the southwestern edge of the Minas Pampa surface (Fig. 3 sections 2 and 3). Large basin-like depressions, partly filled by silt, sand and gravel, which occur on the Minas Pampa surface in the northern and central parts of the region probably developed during the Matacaballo erosion stage.

The Matacaballo erosion surface is mostly overlain by alluvial deposits (Fig. 2) which accumulated during a late phase of the Matacaballo erosion stage, and were incised by Canyon stage erosion. The deposits mainly consist of alternations of gravel, sand and silt. They are generally fairly well sorted and were probably derived by water transport from the adjacent deeply weathered Minas Pampa surface, and were deposited in a flood plain environment when the Matacaballo surface was still connected to sea level.

South of Huarmey (Fig. 2) these alluvial deposits are overlain by a dissected sheet of gravel up to 50 m thick, made up of well rounded, limonite stained pebbles, typically 6 cm in diameter. The gravel accumulated during the latest phase of the Matacaballo erosion stage. Similar gravel deposits occur up to 100 m thick between altitudes of 200–450 m on the Minas Pampa surface on both sides of the Rio Huarmey valley, east of Huarmey, between altitudes of 700–800 m south of the Cerros Yana Orcco, and on terraces in the Rio Fortaleza valley between altitudes of 700–1400 m (Fig. 2 and Fig. 4 sections 7, 8 and 9). The terraces in the Rio Fortaleza valley represent remnants of the broader valley floor at the time of the Matacaballo erosion stage. The gravel south of the Cerros Yana Orcco is mainly composed of gabbro pebbles and a smaller number of pebbles of volcanic rocks. The gabbro pebbles were derived from gabbro which mostly occurred in the uppermost part of the Coastal Batholith and was exposed and eroded for the first time during the Minas Pampa erosion stage.

2f. Canyon erosion stage

The Canyon erosion stage is marked by major incision of former valleys and the widespread destruction of older erosion surfaces. Individual valleys are progressively more incised towards the northeast, and this suggests that uplift was greatest in the northeast and the old landscape was again tilted to the southwest.

The Canyon erosion stage is the youngest major episode of erosion and continues to the present time, although erosion was formerly more vigorous than at present. This can be seen in the Rio Fortaleza valley which is now only cutting through bedrock above an altitude of 1700 m. Below this altitude the valley has been partly filled in by gravel, sand and silt and has a U-shaped rather than a V-shaped profile. This aggradation represents a phase of deposition during the Canyon erosion stage. It may have been caused by a decrease in the rate of uplift, by an independent rise of sea level, by decrease in the volume and flow rate of water, or by combinations of these processes. Garner (1959) described how similar aggradation in the coastal desert of Peru was caused by increasing aridity.

The Canyon erosion stage has now entered a third phase in which these alluvial deposits are being incised during infrequent flow of water in the valleys and quebradas. The numerous minor terraces of alluvial deposits in the lower parts of most valleys

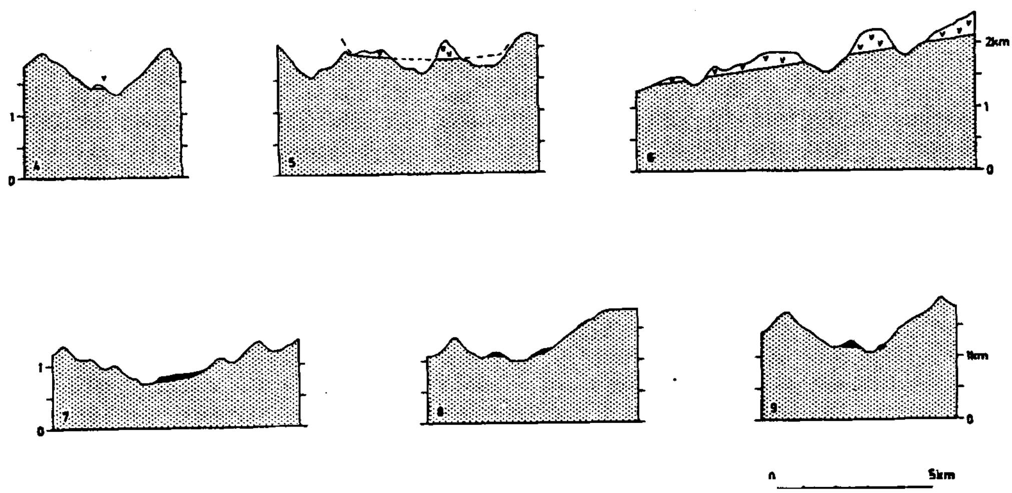


Fig. 4. Topographic profiles along lines 4–9 shown on Fig. 2. Profiles 4 and 5 are across the Rio Fortaleza and adjacent valleys and show the distribution of the Rio Fortaleza Ignimbrite on the older valley floor now incised by Canyon stage erosion. Profile 6 shows the down-valley slope of the the ignimbrite. Profiles 7–9 show gravel deposits of the Matacaballo erosion stage on the older valley floor now incised by Canyon stage erosion.

were formed during this phase, and most may represent recent pulses of uplift or independent sea level changes (cf. Paskoff 1964; Cooke 1964), climatic changes (cf. Garner 1959), or combinations of these processes.

3. Rio Fortaleza Ignimbrite

The Rio Fortaleza Ignimbrite was erupted down the Rio Fortaleza valley during or just after the latest phase of the Matacaballo erosion stage, from a source near Conococha (Fig. 5). It lies on the same old valley floor surface (Fig. 4 sections 4, 5 and 6) as the gravel deposits lower down the valley between Chasquitambo and Chaucayan (Fig. 2 and Fig. 4 sections 7, 8 and 9), and is thus broadly contemporary with these gravel deposits.

Most of the ignimbrite is white, friable and non-welded, and is composed of fragmented pumice and phenocrysts of quartz, altered plagioclase and potash feldspar. It contains partly rounded fragments of white quartzite from the site of eruption near Conococha, and in the lower part below Cajacay (Fig. 5), it also contains locally derived fragments of granite from the Coastal Batholith. The white quartzite fragments are present in all but the lowermost part of the flow near Chaucayan, and are irregularly but uniformly distributed throughout the main flow unit. Most of the ignimbrite is about 750 m thick and has well developed columnar jointing.

In some places the main flow unit overlies finely bedded tuff up to a few metres thick. The tuff shows size-graded bedding and was probably associated with the ignimbrite eruption. It may be the result of either minor eruptions which preceded the major eruption of the ignimbrite and scattered a thin blanket of tuff over a wide region, but which was only protected from erosion beneath the ignimbrite, or it may have formed

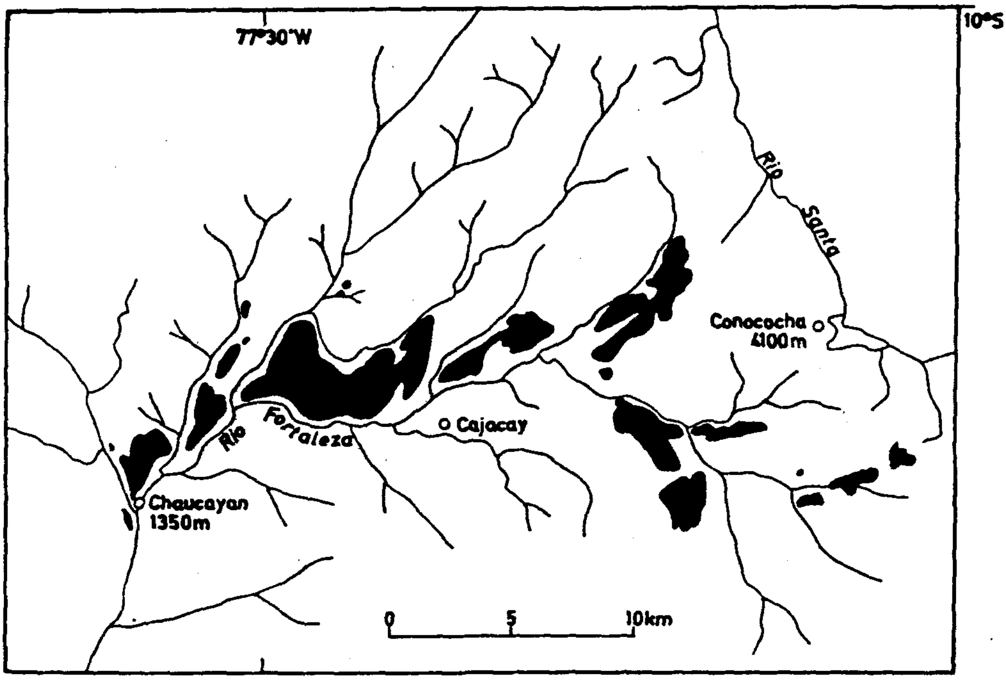


Fig. 5. Map showing the extent of the Rio Fortaleza Ignimbrite Formation (black) in the upper part of the valley of the Rio Fortaleza.

from crystal and pumice fragments which dropped out of the bottom of the flow during its movement down the valley, and settled through a cushion of air over which the ignimbrite may have ridden. The ignimbrite locally oversteps ridges adjacent to the main Rio Fortaleza valley and occurs for short distances up tributary streams. The flow motion may have been similar in many respects to the recent earthquake-triggered mud flow from the mountain of Huascaran which buried the town of Yungay in the Rio Santa valley (Plafker *et al.* 1971).

The ignimbrite rests on an irregular surface of thick grey pebbly soil in some places. The topmost metre of soil is red and was oxidised by the ignimbrite. No traces of charred vegetation were seen in the soil and therefore the climate at the time of eruption may have been arid, similar to that of the present time. If the region was desert and the cold Humboldt current existed off the coast of Peru as far north as at the present time, then the height of the Andes must already have been sufficient to prevent the westward spread of rain-laden winds from the Amazon region, as it does today.

The ignimbrite was deeply incised during the Canyon stage of erosion, and the river bed now lies between 150–500 m below the base of the ignimbrite (Fig. 6). Biotite phenocrysts from the ignimbrite have given a K-Ar isotopic age of $5{\cdot}84 \pm 0{\cdot}2$ m.y. (P. A. Wilson personal communication) which is probably close to the age of eruption. This age is very late Upper Miocene on the time scale proposed by Berggren (1969), it indicates that major uplift and erosion of the Andes took place before 6 m.y. ago, and that the Canyon stage extends back through the Pleistocene and Pliocene.

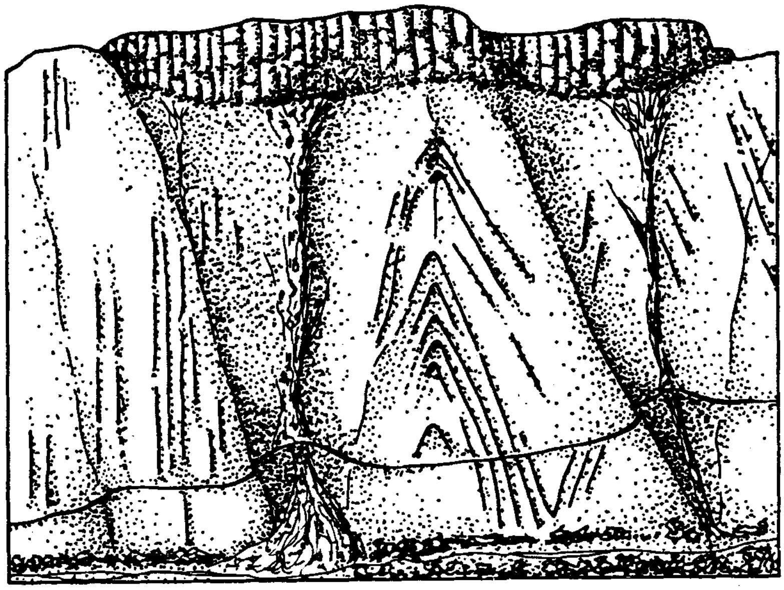


Fig. 6. Canyon stage erosion of the Rio Fortaleza Ignimbrite which rests unconformably on folded Albian volcanic rocks. Height of valley section is 500 m; location is north-east of Chaucayan; (after Myers 1974).

4. Significance of the erosion surfaces

The dissected remnants of pediments which occur at various altitudes on the western flank of the Andes signify the alternation of major episodes of laterally spreading, planation-aggradation, with major episodes of sharp vertical incision of older subplanar surfaces. The great range of altitude of the pediments and the fact that older pediments occur at higher elevations than younger ones suggest that the various altitudes of the pediments resulted from intermittent pulses of tectonic uplift, as there is no worldwide evidence for contemporary intermittent lowering of sea level in this order of magnitude. On the other hand, Garner (1959) has demonstrated how alternations of arid and humid climatic conditions could produce alternations of planation-aggradation and vertical incision of these planed surfaces. The question then arises whether the erosion surfaces reflect tectonic uplift, climatic variations or combinations of both processes.

There is abundant evidence of tectonic elevations and subsidences of the Andes in the geological record. The distribution of Cretaceous and early Tertiary volcanic and sedimentary rock types in northern Peru, their facies variations and deformation structures, suggest that these features were controlled by oscillatory vertical movements of ribbon-like blocks bounded by steep major shear zones parallel with the continental margin during Cretaceous and early Tertiary time (Myers 1974, 1975b). During this time, already established lines of weakness were reactivated during successive elevations and subsidences of the crustal blocks. These events occurred intermittently from at least 100 to 50 m.y. ago, and therefore the elevation of the modern Andes during the last fifty million years was probably the result of continuing similar tectonic processes.

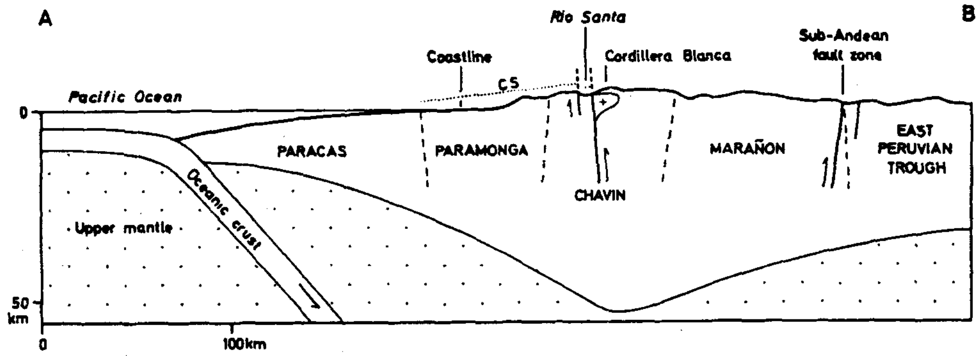


Fig. 7. Section along the line A–B of Fig. 1 showing the movements (arrows) on major tectonic dislocations during Tertiary and younger elevation of the Andes. The dotted line CS marks the slope of the Cochapampa surface (slightly elevated for clarity). Major blocks of continental crust are named in capitals and previously active junctions between them are shown by broken lines. The Miocene Cordillera Blanca Batholith is marked by a cross. Thickness of oceanic and continental crust after James (1971).

During late Tertiary and younger tectonic uplift, the main vertical movements appear to have occurred on steep faults in the Sub-Andean fault zone in the eastern foothills of the Cordillera Oriental (Cobbing 1972), located near the boundary between the East Peruvian Trough and Marañon crustal blocks (Myers 1975b) (Fig. 7). The region of northern Peru described in this account lies on the eastern part of the Paramonga block which probably extends westwards to about 15–20 km southwest of the present coastline (Myers 1975b). The southwestward projection of the approximate slope of the Cochapampa surface intersects the continental shelf at about the same distance from the present coastline. Thus the southwestern edge of the Paramonga block may have acted as a hinge about which the Cochapampa surface was tilted upwards to the northeast.

To the northeast, the Paramonga, Chavin and Marañon blocks were raised together by vertical movements along faults in the Sub-Andean fault zone (Fig. 7). In addition, the Chavin block was split longitudinally by a complex of steep faults just to the west of the present continental divide, in the region of the Rio Santa valley. The eastern part of the Chavin block and the Marañon block were raised above the general level of the Puna surface, and the eastern part of the Chavin block now forms the Cordillera Blanca, the highest part of the Andes in Peru, with peaks of over 6000 m in altitude (Fig. 1). The elevation of this region continues to the present time. The Rio Santa valley itself, which is the only major subsequent stream on the western side of the Andes in northern Peru (Fig. 1), lies in a graben which was probably initiated before or during the first stages of elevation of the Cochapampa-Puna surface, and since then has been intermittently active (Fig. 7).

Whereas tectonic uplift appears to have controlled the elevation of the pediments, the morphology of the erosion surfaces appears to reflect climatic variations and alternations of weathering processes such as those described by Garner (1959). The five major erosion surfaces each show the same sequence of events, although the events are of various magnitudes. In each case active erosion and pedimentation was followed by the accumulation of coarse clastic deposits on the pediments, and then by sharp incision of these sub-planar surfaces. Garner (1959) showed how this sequence of events could result from climatic changes with semi-arid followed by arid and then by humid conditions. This sequence is most clearly seen in the younger erosion surfaces of northern Peru.

The Matacaballo surface represents an episode of arid pedimentation, scarp retreat and aggradation, and is mostly covered by a veneer of coarse clastic deposits, mainly formed by mechanical erosion. The succeeding Canyon erosion stage indicates vigorous incision of this surface during more humid conditions. This incision could have resulted from increased precipitation alone, but the magnitude of the incision suggests that tectonic uplift occurred as well.

Above an altitude of 1800 m there is a widespread soil cover up to 2 m thick on most slopes. This indicates chemical and mechanical weathering in more humid conditions than those of the present time and than conditions which prevailed during the formation of the Matacaballo surface. In many places the lower termination of this soil layer is abrupt, and the edge of the soil layer forms a low scarp below which all soil has been stripped and the bedrock laid bare. The main erosion of this soil cover must have occurred before the accumulation of coarse clastic deposits on the Matacaballo erosion surface because these deposits, such as the Huarmey, Cerros Yana Orcco and Chasquitambo gravels, abut against and partly overlie rock slopes from which the soil was previously removed, as thick soil occurs higher up the same hillslopes. This soil cover indicates that more humid conditions occurred before the formation of the Matacaballo erosion surface.

There is thus evidence of fluctuating climatic conditions both during and between the formation of the erosion surfaces, but the great range of altitude of the erosion surfaces and the widespread evidence of major tectonic elevation and subsidence of the Andes in the geological record, suggest that intermittent tectonic uplift may have been the major process which elevated the erosion surfaces, and the Andes.

Quaternary changes of sea level which accompanied worldwide climatic changes and intensification of polar glaciations can only have influenced deposits which accumulated near the coast during the latter part of the Canyon erosion stage, as the other erosion surfaces are older than the Upper Miocene. Such modifications of the Canyon stage morphology are difficult to distinguish in the region described here because of younger earthquake movements of much of the cliff-bound coastline, but they are more prominent to the south where marine terraces have been described by Cobbing (1973) in central Peru, by Dollfus, Gabert and Laharie (1970) in southern Peru, and by Cooke (1964), Segerstrom (1963, 1964) and Paskoff (1964, 1970) in northern Chile. Thus in the region of Peru described here, changes of climate and sea level may only have been relatively minor modifiers of a dominantly tectonic process of uplift.

5. Correlations in Peru and Chile

The exact correlation of erosion surfaces which are geographically isolated can only be achieved when precise information is available about the age of the surfaces, because the altitude and morphology of contemporary surfaces may differ from place to place as a result of regional differences in tectonic uplift and climate. In southern Peru and northern Chile, extensive sheets of ignimbrite were interbedded with alluvial deposits during Tertiary and Quaternary time, and the relationships between the ignimbrites, alluvial deposits and erosion surfaces can be observed. Determinations of the isotopic ages of the ignimbrites enables fairly precise age constraints to be placed on the erosion surfaces and alluvial deposits. The alluvial deposits are generally unfossiliferous and before isotopic work on the ignimbrites, the ages of the erosion surfaces were based on their relationship with glacial features at high altitudes, and with fossiliferous deposits associated with marine terraces at low altitudes.

In Peru, McLaughlin (1924), Dresch (1958), Dollfus (1965), Jaén (1965), Mendívil (1965), Wilson *et al.* (1967), and Cobbing (1973) all considered that Andean relief was generated by uplift along faults and flexures, and that uplift occurred in the Plio-Pleistocene, but these ages were postulated without isotopic evidence. Noble *et al.* (1974) and McKee *et al.* (1975) provide isotopic data from ignimbrites in southern Peru and Noble *et al.* (1974) concluded that a major episode of folding, uplift and erosion occurred between 14 and 10·5 m.y. ago, but they did not relate the ignimbrites and isotopic work to the evolution of Andean landforms. The isotopic age of the Rio Fortaleza Ignimbrite determined by P. A. Wilson (personal communication) described in this paper provides the first isotopic constraint on the age of the erosion surfaces of Peru.

The erosion surfaces of the desert region of northern Chile have been studied in more detail than those of Peru. Brüggen (1950), Rutland *et al.* (1965), Galli-Olivier (1967), Hollingworth and Rutland (1968) and Guest (1969) all considered that the Andes were elevated by tectonic uplift along faults or flexures. They concluded that the main uplift occurred in the Upper Miocene (Galli-Olivier 1967), or Pliocene (Brüggen 1950), or both (Guest 1969), or began in the Upper Miocene and continued with major pulses in mid-Pliocene and late Pleistocene time (Rutland *et al.* 1965; Hollingworth and Rutland 1968). The latter authors derived the ages of uplift from the isotopic ages of ignimbrites which were erupted intermittently during elevation of the Andes in the northern Atacama desert region (Guest 1969). A similar study of the isotopic ages of ignimbrites associated with erosion surfaces in the southern Atacama desert region by Clark *et al.* (1967) and Mortimer (1973) has shown that much of the Andean relief of that region developed before the Upper Miocene (before 9 m.y. ago), and that the oldest elements of the present topography already existed in the Lower Eocene (Mortimer 1973).

Table 1 tentatively correlates the erosion surfaces described in this paper with surfaces in other parts of northern and central Peru and with the sequence in northern Chile described by Mortimer (1973). The Cochapampa surface is continuous with the gently undulating top of the Cordillera Negra which is probably part of the Puna surface described by McLaughlin (1924), Dresch (1958), Dollfus (1965), Jaén (1965), Mendívil (1965) and Wilson *et al.* (1967). It is not known how the Puna surface relates with the oldest erosion surfaces of northern Chile described by Mortimer (1973), but as the oldest of these surfaces in Chile already existed in the Lower Eocene, the Puna surface of Peru may be younger as it developed on Eocene volcanics. It may be possible to delineate the age of the Puna surface by dating volcanic rocks in southern Peru such as the Toquepala Group and Moquegua Formation which are older and younger than the Puna surface respectively (Jaén 1965). The Puna surface may represent the last of a series of late Cretaceous and early Tertiary erosion surfaces which followed uplift of the region above sea level in mid-Cretaceous time and developed during pauses between eruptions of lava and pyroclastic flows.

McLaughlin (1924) recognised two major erosion surfaces below the Puna erosion surface in central Peru, and similar surfaces were described by Jaén (1965) and Mendívil (1965) in southern Peru. Wilson *et al.* (1967) and Cobbing (1973) similarly recognised two major episodes of erosion which incised the Puna surface in northern Peru which they called the Valley and Canyon stages, equivalent to the Chacra and Canyon stages of McLaughlin (1924), (Table 1). The Matacaballo erosion surface is tentatively correlated with the Chacra stage of McLaughlin (1924) and Valley stages of

Table 1. Correlation of erosion surfaces in Peru and Chile

N. Peru, western Andes this paper	*N. Peru, central Andes Wilson* et al. *1967*	*Central Peru McLaughlin 1924*	*N. Chile Mortimer 1973*
Max. uplift of preceding erosion surface Canyon 750 m	Canyon	Canyon	Phase 4
Rio Fortaleza Ignimbrite 6 m.y. Matacaballo 750 m	Valley	Chacra	Ignimbrite 9 m.y. Ignimbrite 11·5 m.y.
Minas Pampa 1500 m Shanan B 750 m Shanan A 500 m		Junin	Phase 3 Atacama pediplain
Cochapampa	Puna	Puna	Phase 2 Phase 1 pre-55 m.y.

Wilson *et al.* (1967) and Cobbing (1973) because of their similar features and position in the sequence of erosion surfaces. An extensive and generally similar surface was described from the lower reaches of the Rio Elqui and Rio Huasco in the southern Atacama desert region of Chile by Cooke (1964) and Paskoff (1964, 1970), but because of its great distance from the Matacaballo surface and the lack of precise isotopic evidence of the ages of the two surfaces, they are not correlated in Table 1.

The Canyon erosion stage described in this paper probably correlates with the Canyon stage described by McLaughlin (1924), Wilson *et al.* (1967) and Cobbing (1973). It is similar in morphology and relative age to the stage of canyon formation (phase 4) described by Mortimer (1973 p. 515) in the southern Atacama desert region (Table 1). In both Peru and Chile these are the youngest major stages of erosion and they continue to the present time. Isotopic evidence indicates that these stages are broadly equivalent in age, although the canyon stage of erosion in Chile appears to have begun 9 m.y. ago (Mortimer 1973), three million years before the Canyon stage described here from northern Peru.

The sequence of incised erosion surfaces described here are all older than the 6 m.y. old Rio Fortaleza Ignimbrite. This evidence contrasts with that of Rutland *et al.* (1965) and Hollingworth and Rutland (1968) from Chile, who concluded that the northern Atacama desert region was of low relief until late Miocene time, and that a large portion of Andean uplift occurred there in the late Pliocene and Pleistocene. On the other hand, the evidence in northern Peru is similar to that of Clark *et al.* (1967) and Mortimer (1973) who found that most Andean relief in the southern Atacama desert formed before the late Miocene, and that the history of the present landforms extends far back into the Tertiary.

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References

BERGGREN, W. A. 1969. Cenozoic chronostratigraphy, planktonic foraminiferal zonation and the radiometric time scale. *Nature, Lond.* **224,** 1072.

BRÜGGEN, J. 1950. *Fundamentos de la geologia de Chile.* Instituto Geográfico Militar, Santiago, Chile.

CLARK, A. H., MAYER, A. E. S., MORTIMER, C., SILLITOE, R. H., COOKE, R. U. and SNELLING, N. J. 1967. Implications of the isotopic ages of ignimbrite flows, southern Atacama desert, Chile. *Nature, Lond.* **215,** 723.

COBBING, E. J. 1972. Tectonic elements of Peru and the evolution of the Andes. *24th Int. Geol. Congr. Montreal,* section 3, 306.

—— 1973. Geologia de los cuadrángulos de Barranca, Ambar, Oyon, Huacho, Huaral y Canta. *Boln. Serv. geol. min. Peru* **26.**

COOKE, R. U. 1964. Les niveaux marins des baies de la Serena et de l'Huasco. *Bull. Ass. Geogr. fr.* **320,** 19.

DOLLFUS, O. 1965. Les Andes centrales du Pérou et leurs piémonts (entre Lima et le Péréné), étude géomorphologique. *Trav. Inst. fr. Etud. Andines* **10.**

——, GABERT, P. and LAHARIE, R. 1970. Les problèmes morphologiques du piémont désertique des Andes péruviennes méridionales. *Revue de geogr. Alpine, Grenoble* **58,** 265.

DRESCH, J. 1958. Problèmes morphologiques des Andes centrales. *Ann. Geogr.* **67,** 130.

GALLI-OLIVIER, C. 1967. Pediplain in northern Chile and the Andean uplift. *Science* **158,** 653.

GARNER, H. F. 1959. Stratigraphic-sedimentary significance of contemporary climate and relief in four regions of the Andes mountains. *Geol. Soc. Am. Bull.* **70,** 1327.

GUEST, J. E. 1969. Upper Tertiary ignimbrites in the Andean cordillera of part of the Antofagasta province, northern Chile. *Geol. Soc. Am. Bull.* **80,** 337.

HOLLINGWORTH, S. E. and RUTLAND, R. W. R. 1968. Studies of Andean uplift, part 1—post-Cretaceous evolution of the San Bartolo area, north Chile. *Geol. J.* **6,** 49.

JAEN, H. 1965. Geologia del cuadrángulo de Tarata. *Boln. Comn. Carta nac. Peru* **11.**

JAMES, D. E. 1971. Plate tectonic model for the evolution of the Central Andes. *Geol. Soc. Am. Bull.* **82,** 3325.

McKEE, E. H., NOBLE, D. C., PETERSEN, U., ARENAS, M. F. and BENAVIDES, A. Q. 1975. Chronology of late Tertiary volcanism and mineralization, Huachocolpa district, central Peru. *Econ. Geol.* **70,** 388.

McLAUGHLIN, D. H. 1924. Geology and physiography of the Peruvian cordillera, Departments of Junin and Lima. *Geol. Soc. Am. Bull.* **35,** 591.

MENDIVIL, S. 1965. Geologia de los cuadrángulos de Maure y Antajave. *Boln. Comn. Carta nac. Peru* **10.**

MORTIMER, C. 1973. The Cenozoic history of the southern Atacama desert, Chile. *J. geol. Soc. Lond.* **129,** 505.

MYERS, J. S. 1974. Cretaceous stratigraphy and structure, western Andes of Peru between latitudes 10°—10°30′. *Bull. Am. Ass. Petrol. Geol.* **58,** 474.

—— 1975a. Cauldron subsidence and fluidization: mechanisms of intrusion of the Coastal Batholith of Peru into its own volcanic ejecta. *Geol. Soc. Am. Bull.* **86,** 1209.

—— 1975b. Vertical crustal movements of the Andes in Peru. *Nature, Lond.* **254,** 672.

—— in press. Geologia de los cuadrángulos de Huarmey y Huayllapampa. *Boln. Serv. geol. min. Peru.*

NOBLE, D. C., McKEE, E. H., FARRAR, E. and PETERSEN, U. 1974. Episodic Cenozoic volcanism and tectonism in the Andes of Peru. *Earth Planet. Sci. lett.* **21,** 213.

PASKOFF, R. 1964. Remarques sur des niveaux marins et fluviaux autour de la baie de Coquimbo (Chili). *Bull. Ass. Geogr. fr.* **320,** 2.

—— 1970. *Rechérches géomorphologiques dans le Chili semi-aride.* Bordeaux (Biscaye Frères).

PLAFKER, G., ERICKSEN, G. E. and CONCHA, J. F. 1971. Geological aspects of the May 31, 1970, Peru earthquake. *Seismol. Soc. Am. Bull.* **61**, 543.
RUTLAND, R. W. R., GUEST, J. E. and GRASTY, R. L. 1965. Isotopic ages and Andean uplift. *Nature, Lond.* **208**, 677.
SEGERSTROM, K. 1963. High marine terraces in the Caldera region of northern Chile. *Geol. Soc. Am.* Spec. Pap. **73**, 237.
—— 1964. Quaternary geology of Chile: brief outline. *Geol. Soc. Am. Bull.* **75**, 157.
WILSON, J., REYES, L. and GARAYAR, J. 1967. Geología de los cuadrángulos de Mollebamba, Tayabamba, Huaylas, Pomabamba, Carhuaz y Huari. *Boln. Serv. geol. min. Peru* **16**.

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