[AMERICAN JOURNAL OF SCIENCE, VOL. 286, DECEMBER, 1986, P. 737-764]

American Journal of Science

DECEMBER 1986

THE SIERRAS PAMPEANAS OF ARGENTINA: A MODERN ANALOGUE OF ROCKY MOUNTAIN FORELAND DEFORMATION

TERESA E. JORDAN and RICHARD W. ALLMENDINGER
Institute for the Study of the Continents and Department of Geological Sciences, Cornell University, Ithaca, New York 14853

ABSTRACT. The Sierras Pampeanas province of Argentina is a region of large mountain ranges of crystalline basement and intervening broad valleys. It is located on the eastern side of a thin-skinned thrust belt of the Andes Mountains, coincident with a region where the subducted Nazca plate is sub-horizontal. Its morphology and tectonic setting are similar to those of the Rocky Mountain foreland province of the North American Cordillera. Because the deformation is of Late Cenozoic age, it provides new insight into regions with ancient foreland basement deformation.

The mountain ranges are uplifted by reverse faulting and local folding. The faults dip at moderate angles at the surface, and compressional earthquakes indicate similar dips at mid and lower crustal depths. Throws of 2 to 8 km are estimated across the major structures. Net shortening is estimated to be about 2 percent. Tilting of the upper surface of the crystalline basement suggests that many of the faults are listric in the subsurface and flatten at mid to lower crustal depths.

The age constraints on the deformation are incomplete. It is clear that many of the faults have been active during the Quaternary. We estimate that faulting began during the past 10 Ma.

INTRODUCTION

Since Chamberlin (1919) first characterized the Colorado Rockies as "thick-shelled" and the Appalachians as "thin-shelled" mountain ranges, geologists have distinguished two fundamental structural styles of foreland deformation. Thin-skinned fold and thrust belts such as the Cordilleran thrust belt or the Appalachian Valley and Ridge province parallel and are continuous along the length of most continental orogenic belts and therefore can be related in a general way to plate interactions. Much more enigmatic are thick-skinned deformational provinces, which occur within discrete segments of an orogenic foreland and exhibit a structural style that can be characterized only in three-dimensions.

The most widely recognized thick-skinned province is the basement uplifts province of the Rocky Mountain foreland ("Laramide province") of New Mexico, Colorado, Wyoming, and southern Montana (fig. 1). This province has been extensively described in the literature, and its subsurface structure is relatively well known. However, because the deformation occurred during the late Mesozoic and early Cenozoic, its plate tectonic setting is much less well understood, and the mechanical nature of the crust during deformation remains in the realm of speculation. A less familiar thick-skinned province is the Sierras Pampeanas of western Argentina (fig. 1). Although the Sierras Pampeanas has been less explored than its Rocky Mountain counterpart, it constitutes an active province with dynamic morphology, crustal seismicity, and known plate setting. Thus, a comparison of both provinces can define the salient characteristics of thick-skinned deformation and place it in a well defined plate tectonic setting. Because the Sierras Pampeanas are little known outside the southern cone of South America, the majority of this paper describes their geology and tectonic setting. With the assumption that the readers are already familiar with the Rocky Mountain foreland, we then draw comparisons and contrasts to the structural geology and tectonics of that region.

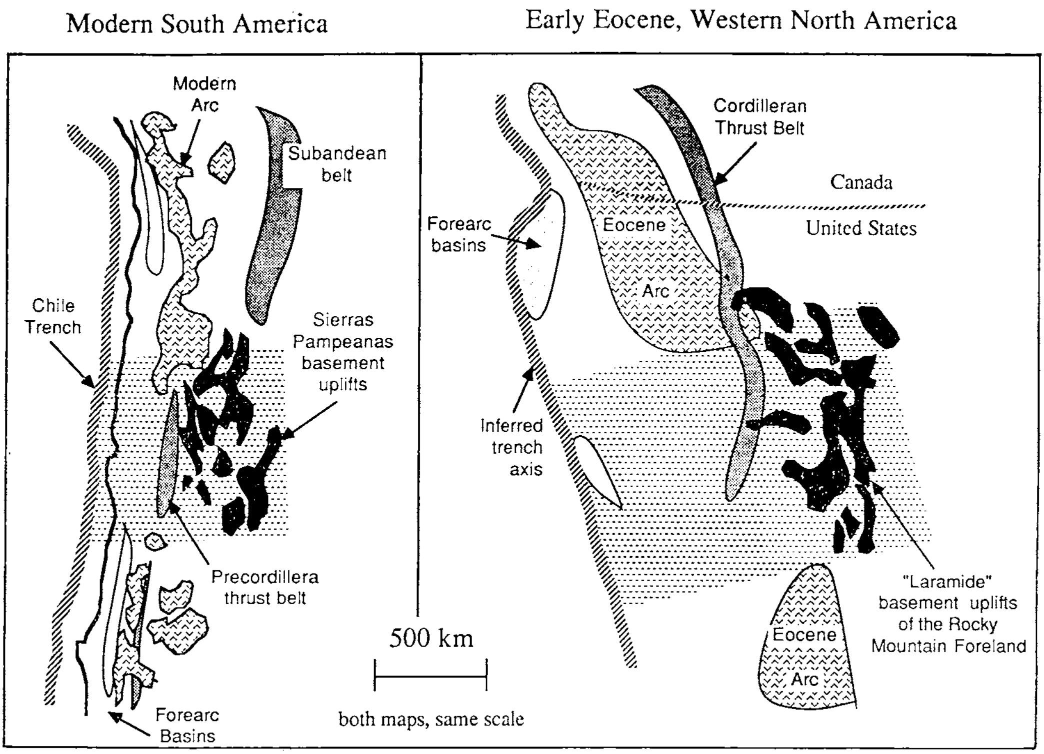


Fig. 1. Location of Sierras Pampeanas (Rocky Mountain Foreland Ranges) and their setting in a map of active (early Cenozoic) tectonic features of Central Andes (North American Cordillera). The region of flat subduction is shown by dashed pattern; areas to north and south have a steeper dipping subducted Nazca (Farallon) plate beneath them. (Names in parentheses refer to right hand side of diagram.) After Jordan and others (1983a).

The Sierras Pampeanas are a series of mountains and basins in west central Argentina which constitute a distinctive morphotectonic province between 26°00′ and 33°15′S, 63°30′ and 68°30′W. There is a coeval thrust belt (but no active volcanic arc) along most of the western border of the Sierras Pampeanas (fig. 1). It is above a segment of the subducted Nazca plate that is nearly horizontal at 120 to 180 km depth. Although the geographic definition of the Andes Mountains refers only to the ranges to the west, the Andean orogenic zone includes the 450 km wide by 800 km long Sierras Pampeanas. About 12 major mountain blocks, composed of crystalline basement rock, were uplifted on reverse faults during the Late Tertiary and Quaternary. The mountains are generally asymmetric, with a set of north-trending faults bounding either the east or west side and gentle dip slopes on the other flank. The valleys are flat and broad and have acted as depositional centers with either axial or closed drainage systems (fig. 2).

Parallel questions remain unanswered in the Sierras Pampeanas and in the Rocky Mountain foreland. In the Rocky Mountains, much interest has focused on the subsurface geometry of the faults and folds that bound the basement uplifts. Seismic reflection profiling and drilling have demonstrated that many of the basement blocks were uplifted along 20°- to 30°-dipping thrust faults which formed by compression (Smithson and others, 1979; Gries, 1983a, b). Similarly, the Sierras Pampeanas have traditionally been recognized as fault blocks formed by compression (Gonzalez Bonorino, 1950; Cuerda, 1973), but the geometries of the faults at depth are not well known. In both regions, pre-existing structures and poorly understood crustal rheology interact as controls on the young deformation. Also in both regions, the fault chronology is not well known, and hence spatial trends in deformation through time are poorly determined (see Gries, 1983a). A particularly interesting question is how the crust of these areas is coupled to adjacent areas and to the subducting sub-horizontal plate.

PLATE TECTONIC SETTING

The Nazca plate subducting beneath western South America is divided into five large segments: beneath two segments it is nearly horizontal below the continent, whereas in alternating segments it dips about 30° eastward (Barazangi and Isacks, 1976). The Sierras Pampeanas mostly overlie one region of sub-horizontal subduction of the Nazca plate (fig. 1). Between 28° and 33°15′S the sub-horizontal Benioff zone is well-defined. To the north, the region between 24° and 28°S is interpreted as a gradual southward shoaling of the Benioff zone (Bevis and Isacks, 1984; Cahill and Isacks, 1985). This transition in subduction geometry is matched by a transition in foreland structural geometry. The main part of the Sierras Pampeanas occurs south of 28°S, and the thin-skinned Subandean belt occurs north of 24°S; in between lie the northern Sierras Pampeanas and the Santa Barbara System, which have characteristics of both thick- and thin-skinned provinces. The eastern side of the Sierras

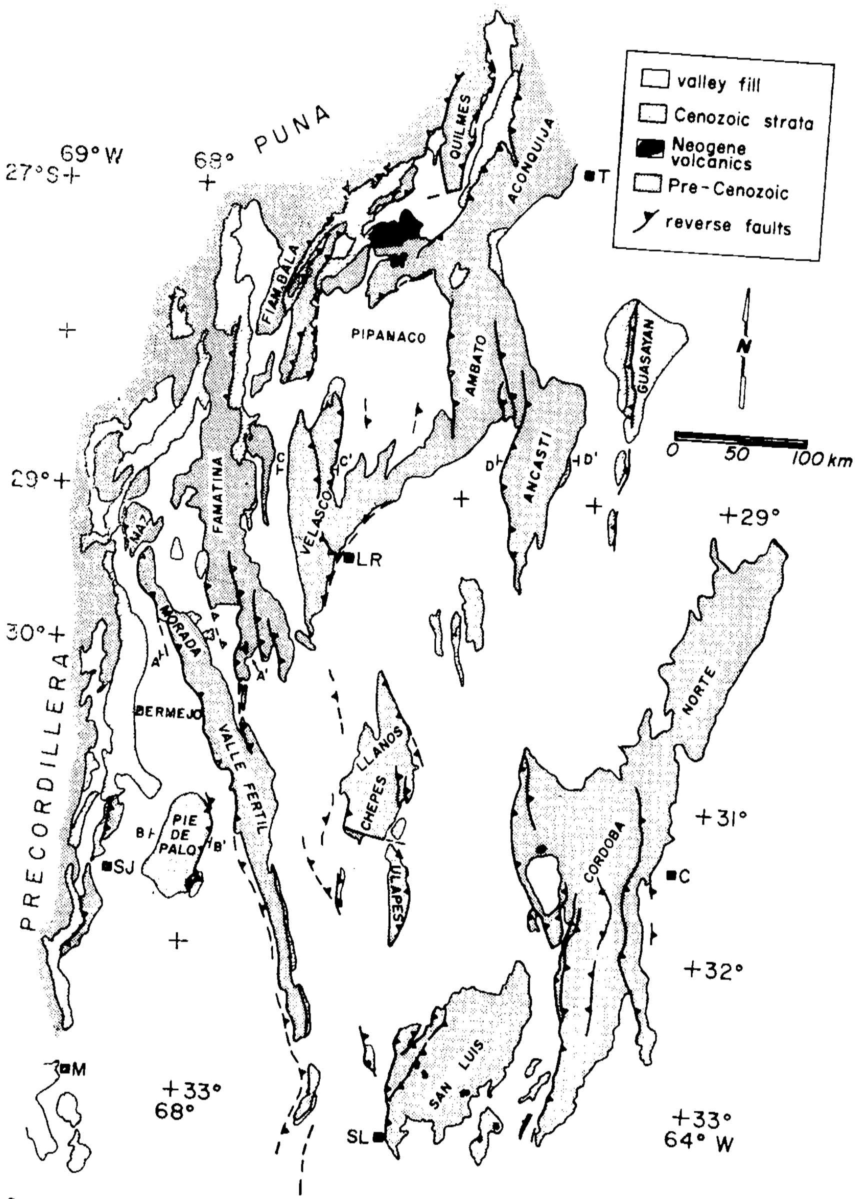


Fig. 2. Distribution of uplifts, basins, and principal structural features of the Sierras Pampeanas and easternmost parts of Precordillera thrust belt along the western boundary. Cities noted at filled boxes are: SJ = San Juan; M = Mendoza; SL = San Luis; C = Cordoba; LR = La Rioja; T = Tucuman. Locations of cross sections in figure 4 are shown.

Pampeanas is roughly coincident with the easternmost (most cratonward) Benioff zone seismicity at 180 km depth, which is 650 km east of the trench (Jordan and others. 1983a), and the western border lies above the 120 km depth contour of the Benioff zone (Bevis and Isacks, 1984).

Theories of the genesis of flat-subduction commonly relate it to one of four characteristics of the subducted oceanic plate or plate motions: (1) increased convergence rate, (2) subduction of young crust, (3) subduction of bathymetric features such as aseismic ridges or seamount chains, or (4) absolute trenchward overriding of the upper plate (Pilger, 1981). Examination of the Nazca-South American plate boundary can rule out at least two of these as explanations of flat-subduction, as discussed by Pilger (1981). First, the present day convergence rate at the Nazca-South America plate boundary is relatively rapid [approx 10 cm/yr at the latitude of the Sierras Pampeanas (Minster and Jordan, 1978)] and nearly perpendicular to the plate margin, and there is no significant difference in rate between regions with flat-subduction and those with steep-subduction. Even if the convergence rate had increased through time (Pilger, 1984, shows that the reverse occurred since about 20 Ma), that would not explain why flat-subduction is confined to discrete segments (Henderson, Gordon, and Engebretson, 1984). Likewise, the age of oceanic crust can be ruled out as a control: the youngest crust subducted along the margin occurs at the triple junction with the Chile Ridge at 49°S, the age increases steadily northward from there, and changes in age do not correlate with the flat-subduction segment boundaries (Corvalan, 1981).

The third hypothesis is not as easy to rule out. Pilger (1981) suggested that the flat-subduction beneath central Chile and Argentina is caused by subduction of the Juan Fernandez island-seamount chain, and that the flat-subduction beneath Peru is due to subduction of another aseismic ridge, the Nazca Ridge; those bathymetric highs are presumably expressions of anomalously warm and consequently buoyant lithosphere. A careful examination of bathymetry, Benioff zone geometry, and convergence direction shows that the Nazca Ridge is being subducted several hundred kilometers north of the flat-to-steep transition in the subducted plate (Bevis and Isacks, 1984, Cahill and Isacks, 1985). The east-trending Juan Fernandez Chain now coincides closely with the southern boundary of the flat-subduction segment. Pilger (1981, 1984) hypothesized that its sub-South America extension trends sharply northeastward, such that at about 13 Ma it first intersected the trench zone farther north than at present and steadily swept southward to its present location. This model predicts a north to south progression in the upper plate responses to low-angle subduction, such as the extinction of the magmatic arc and onset of Sierras Pampeanas foreland deformation (Pilger, 1981). A more complete data set than that used by Pilger (1981, 1984) shows that cessation of volcanism in the main Andean cordillera (discussed below) coincides well with his predicted time of flattening of the angle of subduction (Jordan and Gardeweg, in press). However, there is no significant difference in the youngest measured ages of arc volcanism between the northern and southern parts of the extinct arc (11 ± 0.2 Ma near 29°20′S compared to 12.3 ± 0.4 Ma near 32°20′S; Maksaev and others, 1984; Munizaga and Vicente, 1982, respectively). Thus, the southward sweep of flattening of the subduction angle is not supported by current information about the arc history, or else the flattening progressed more rapidly than the magma-generating processes could respond.

The fourth hypothesis, of increased absolute overriding of the subducted plate by the South American lithosphere, is also attractive, if there is a mechanism by which it can vary along the margin even though the convergence rate remains nearly constant. Isacks (1985) has proposed a model that would result in such a variation along the margin, if shortening of the western South American plate differs between the 30°-subducting segments and the flat-subduction segments to the north and south of it. Finally, we have noted elsewhere (Allmendinger and others, 1983a) that pre-existing structure of the continental lithosphere, perhaps in combination with some of the mechanisms described above, may play an important role in determining the location of flat segments.

The Rocky Mountain foreland basement uplifts formed in western North America between about 80 (locally 90) and 40 Ma and were located 1000 to 1500 km from an active subducting plate margin (fig. 1). Patterns of volcanism suggest that the basement deformation in Wyoming and Colorado may have occurred above a nearly flat subducted plate (Coney, 1976; Dickinson and Snyder, 1978). Plate reconstructions have suggested that the time of formation coincided with a time of unusually high convergence rates (10-15 cm/yr) and absolute motion of the North American plate toward the trench (Engebretson, Cox, and Thompson, 1984). The latitudinal limits of the flat subduction, and thus the Rocky Mountain foreland deformation, may have been caused by subduction of buoyant aseismic ridges (Henderson, Gordon, and Engebretson, 1984). It is not clear that the space-time trends in deformation synthesized by Gries (1983a), with shortening direction swinging from east to northeast to north trends through time, bear any direct relationship to the southward migration of the flat slab geometry that was hypothesized by Henderson, Gordon, and Engebretson (1984).

ASSOCIATED TECTONIC PROVINCES

The Sierras Pampeanas are an integral part of the Andean deformation (fig. 1), even though they are not considered part of the morphological Andes. To the north they pass along strike into the high plateau of the Argentine Puna, the late Miocene thrust belt of the Eastern Cordillera, and the southern extension of the thin-skinned Pliocene to Quaternary Subandean belt (Jordan and others, 1983a; Allmendinger and others, 1983a). To the east and south, the Sierras Pampeanas abut undeformed cratonic rocks. To the west, the Sierras Pampeanas parallel and locally impinge upon the Precordillera thin-skinned thrust belt, which was deformed during the late Miocene to Recent (Jordan and others, 1985). The Precordillera forms the morphological "foothills" of the main Andes Mountains.

West of the Sierras Pampeanas between 28° and 33°S, the main Andes are a high, narrow chain comprised of an inactive thrust belt and an extinct volcanic arc. A period of faulting on high-angle reverse faults occurred between 19 and 14 Ma, and thrusting occurred locally about 10 Ma (Munizaga and Vicente, 1982; Maksaev and others, 1984; Ramos and others, 1985). Between 29° and 33°S, ignimbrites, mafic andesites, andesites, and rhyolites with dates between 27 and 12 Ma crop out principally in a 30 km wide zone on the western flank of the mountains, about 200 km east of the trench (Munizaga and Vicente, 1982; Maksaev and others, 1984). Similar aged subvolcanic bodies (Lencinas, 1976) in the Precordillera indicate that the early and middle Miocene arc was up to 150 km wide. Younger volcanic activity is indicated only by volumetrically minor ignimbrites, small subvolcanic dacites, and dikes in the vicinity of the old arc (Munizaga and Vicente, 1982; Maksaev and others, 1984; Jordan and Gardeweg, in press). Thus, we infer that, in the area where the Nazca plate is now known to be subhorizontal, the angle of descent was significantly steeper before about 10 Ma (Jordan and others, 1983b; Kay and others, in press).

Farther east, localized subvolcanic bodies and ashes in the eastern and southeastern Sierras Pampeanas have ages of 5 to 8 Ma (Gordillo and Linares, 1981). This latest Miocene calcalkaline volcanic activity suggests that the subducted Nazca plate had reached the depth in the mantle at which magmas feeding overlying arcs are generated (120-170 km). This was about 650 km east of the trench (Kay and Gordillo, 1985), suggesting a 10° to 15° dip of the subducted plate at that time. The space-time patterns of magmatism and deformation and constraints on the Nazca plate geometry are discussed further in Jordan and Gardeweg (in press) and Kay and others (in press).

The northern Sierras Pampeanas (26°-28°S) lie east of an active segment of the arc. A major 11 to 7 Ma volcanic center occurs within the Sierras Pampeanas (fig. 2) (Caelles and others, 1971; Acenolaza and others, 1982). Available information on the ages of volcanism between 28° to 26°S (Gonzalez-Ferran, Baker, and Rex, 1985; Jordan and Gardeweg, in press) is not sufficient to show if the south end of the active arc may have migrated northward or southward since 10 Ma. Gonzalez-Ferran, Baker, and Rex (1985) noted a realignment of spatial patterns of volcanism near 27°S at about 5 to 6 Ma.

BASEMENT GEOLOGY OF THE SIERRAS PAMPEANAS

The Sierras Pampeanas consist largely of metamorphic rocks and extensive granitoid plutons. The metamorphic basement is mostly composed of low- to medium-grade schists, amphibolites, and gneisses, although granulites are important in the eastern ranges. Metamorphic assemblages indicate that the exposed crust in the western region was buried to depths of 20 to 35 km (Dalla Salda and Varela, 1984), and mylonite fabrics and mineral assemblages in the central region indicate that exposed rocks were buried to at least 10 to 15 km depth during the Paleozoic (Toselli, Toselli, and Acenolaza, 1985). The metamorphism has been dated between

600 and 900 Ma (Caminos and others, 1982; Spinelli, 1983). The present character of the Sierras Pampeanas basement was strongly determined by Paleozoic plutonism. The mapping and dating of these intrusive rocks is not yet complete, but two separate phases and types of intrusions emerge from the data.

During the Ordovician, granodioritic and tonalitic batholiths were emplaced in a north-trending zone in the central Sierras Pampeanas, accompanied by metamorphism throughout the region (Rapela, Mheaman, and McNutt, 1982). Northward, the plutonic belt can be traced into regions where andesitic volcanics are intercalated with Ordovician sedimentary rocks. The Ordovician plutons and metamorphism are interpreted as a magmatic arc related to subduction (Caelles, ms; McNutt and others, 1975); during the Ordovician, subduction occurred along the western margin of the South American continent, which was along the western flank of the Precordillera (Ramos and others, 1984). The batholiths grade northward into volcanic rocks indicating that deeper crustal levels are exposed in the Sierras Pampeanas than are exposed farther to the north (Allmendinger and others, 1983a; Toselli and others, 1984).

A second period of plutonism occurred during the Carboniferous. These plutons are smaller and less abundant than the older batholiths and occur principally in the eastern Sierras Pampeanas. They are true granites and are isotopically distinct from the older arc rocks (Rapela, Mheaman, and McNutt, 1982; Caminos and others, 1982).

During the late Paleozoic, the roof of the plutonic terrane was regionally eroded, bevelling off an unknown thickness (perhaps 10 to 14 km) of the upper crust (Mirré, 1971; Dalla Salda and Varela, 1984). This extensive erosional surface is currently exposed on many of the mountain blocks, and it serves as an important reference horizon in structural reconstructions.

Synchronous with and following the erosion, a broad series of nonmarine to marginal marine sedimentary basins formed. The clastic strata, known collectively as the Paganzo Group, are typically 1000 to 2000 m thick in a band across the northern and western Sierras Pampeanas. Local depocenters accumulated over 4000 m of sediments, but the strata are thin or absent in other areas (Salfity and Gorustovich, 1983). Alkaline basalt lavas and dikes are associated with Paganzo Group strata and locally yield ages of 266 to 295 Ma (Late Carboniferous-Permian) (Azcuy, 1975; Mirré, 1976). Triassic strata are also thick in localized depocenters, mostly in the western Sierras Pampeanas. Reactivation during the Cenozoic of basement faults and faults that bounded the late Paleozoic and Mesozoic basins is commonly cited in the literature (Caminos, 1979; Toselli, Toselli, and Acenolaza, 1985).

Little is known about the late Mesozoic-early Cenozoic history of the region. Mid-Cretaceous basaltic dikes occur locally in the western and eastern Sierras Pampeanas (Lencinas, 1971; Cuerda and others, 1981), but their regional importance and continuity are not known.

CENOZOIC STRUCTURE OF THE SIERRAS PAMPEANAS

The Cenozoic structures of the region are defined by the range-bounding faults, the attitude of the tilted upper Paleozoic erosional surface, and folds in upper Paleozoic strata. Although many late Cenozoic structures may occur in the valleys rather than in the mountains, the central parts of the broader basins show only minor faulting and folding of Cenozoic strata and unconsolidated Quaternary deposits.

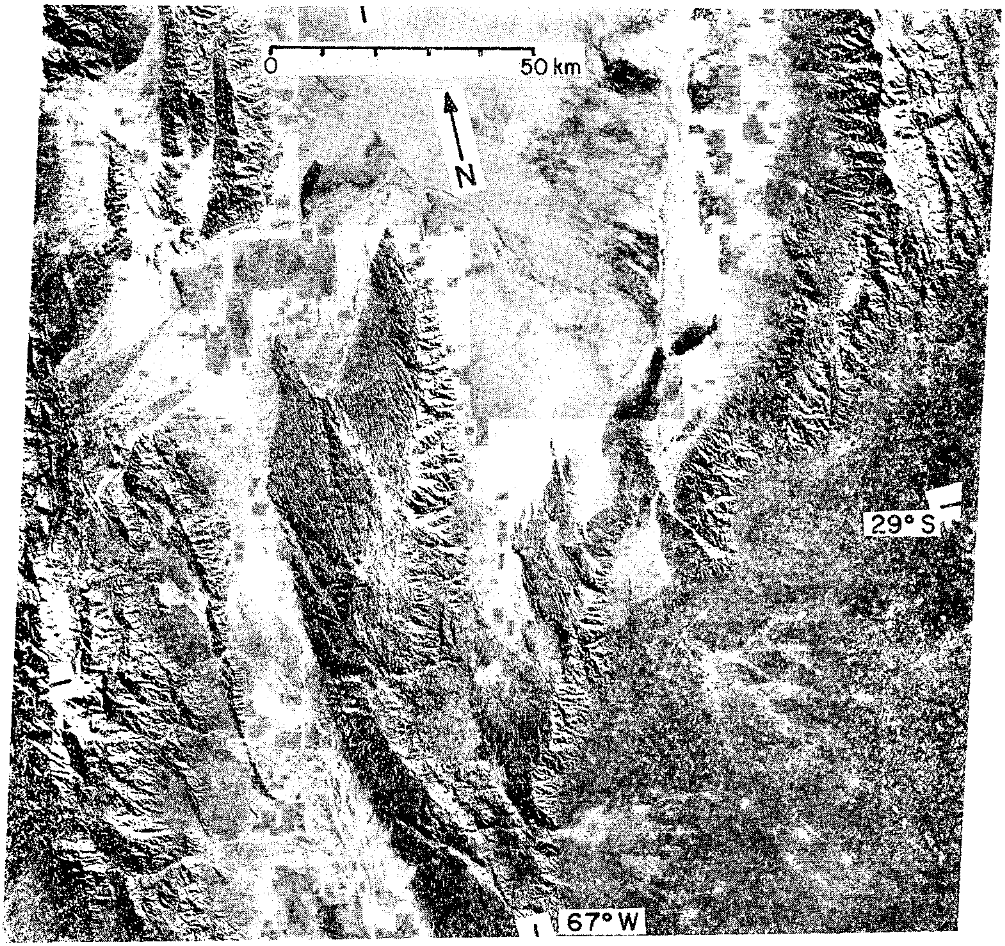
The mountain blocks trend northward, or, rarely, north-northwest or north-northeast, and are bounded on either the east or west flanks by a reverse fault (fig. 2). The upper Paleozoic erosional surface and overlying strata dip gently under the basin on the opposing side, in most cases creating a marked structural asymmetry or vergence (pl. 1). Some of the broader ranges are subdivided by major subparallel fault sets, spaced 10 to 20 km apart, which dip in either the same direction as or opposite to the range-bounding faults. These are particularly important in the Sierras de Cordoba, Velasco (pl. 1), and Ambato (fig. 2).

Some ranges are bounded or bisected along strike by east-trending reverse or strike-slip faults. In some instances, such as the transverse zones within Pie de Palo, between Llanos and Chepes, and at the south end of Sierra de Chepes (pl. 2) there has been little or no movement since the time of deposition of the upper Paleozoic-Mesozoic strata. In other cases, such as the south end of Sierra de Quilmes and north end of Pie de Palo (pl. 3), the geometry of the structures suggests that oblique slip or scissoring accommodates east-west shortening. But in nearly all cases the top of the basement dips gently to the north and south from the center of the range, and throw on the range-bounding reverse fault diminishes gradually toward the ends (Cuerda, 1973).

A notable exception to the general structural pattern is the Sierra de Famatina (fig. 2) which, at 6200 m, is the highest of the ranges in the Sierras Pampeanas. On the basis of its basement geology, which includes metasedimentary rocks of Ordovician depositional age associated with a large Paleozoic batholith and minor Cenozoic volcanic rocks, it is not considered part of the Sierras Pampeanas structural province by Argentine geologists. However, in the Cenozoic, it has been deformed at the same time as and on the same scale as the other ranges in the Sierras Pampeanas. It is uplifted along a complex set of reverse faults within the range, rather than along a system of faults at the range boundary (de Alba, 1979).

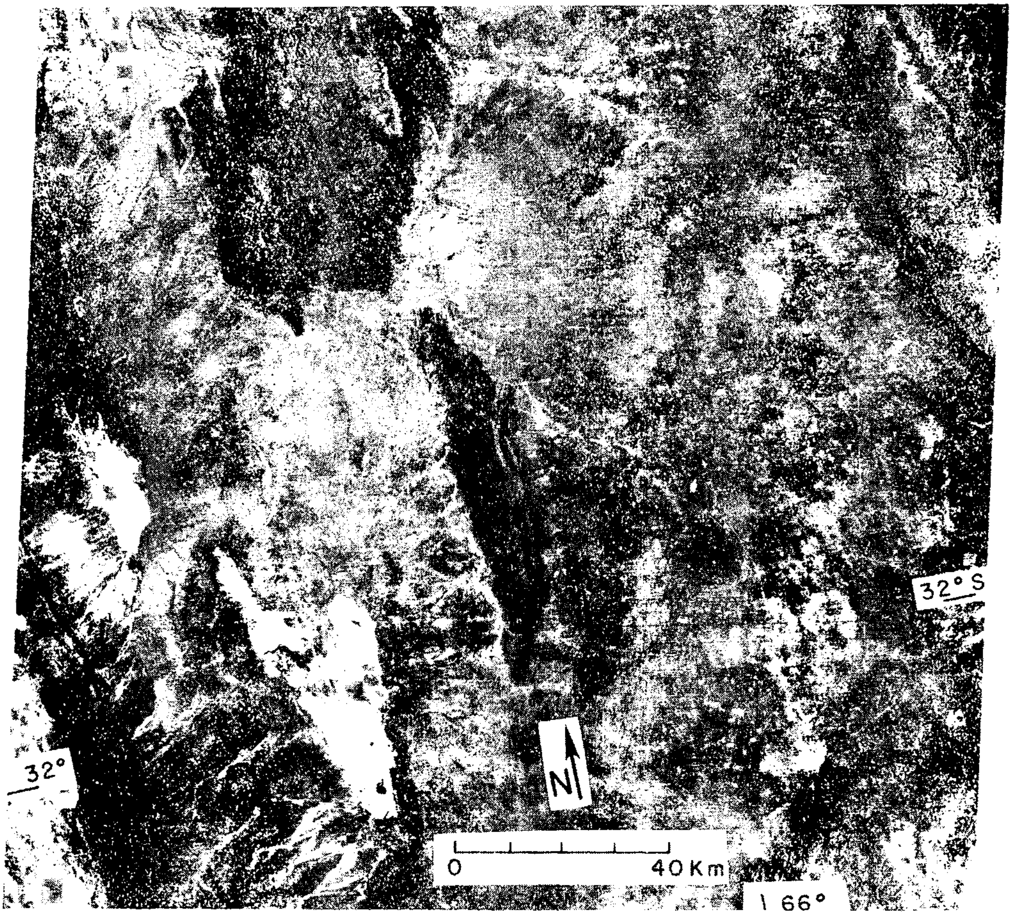
The surface morphology of the province is inclined gently southward, with both crestlines and valley elevations higher in the northern than in the southern Sierras Pampeanas. The principal mountain ranges vary in maximum elevation from about 1000 m (Sierra de Ulapes) to about 3000 m for the larger ranges south of 30°S (southern part of Sierra de Cordoba, Valle Fértil, and Pie de Palo), to 4200, 5500, and 6000 m, for the largest ranges north of 30°S (Sierras de Velasco, Aconquija, and Famatina, respectively). Subsurface data on the depth of the Paleozoic erosion surface on the downthrown sides of the range-bounding faults are not available in most cases, but elevation differences to the topographic

PLATE 1



Landsat MSS image of Sierra de Velasco and neighboring ranges and basins. Velasco is bound along its eastern margin by a major set of west-dipping reverse faults, and a second set of west-dipping reverse faults border the two northwest projections of the range. The exhumed late Paleozoic erosion surface forms a broad smooth area in the northern projection of Sierra de Velasco. Compare to figures 2 and 4C.

PLATE 2



Landsat MSS image of central Sierras Pampeanas, illustrating structural corner at southeast margin of Sierra de Chepes and minor rotated block to south (Sierra de Ulapes). Compare to figure 2. Neither Chepes nor Ulapes appears to be substantially uplifted (topographic relief of a few 100 m), and the distribution of upper Paleozoic-Mesozoic strata suggests that the young faults reactivate older structures. The structural corner occurs at the intersection of a north-trending, east-dipping reverse fault and an east-trending strike-slip (?) fault. In the southwest corner is the southern part of Sierra de Valle Fértil, composed of an anticline in Mesozoic-Cenozoic strata. In the northeast corner is the Sierra de Pocho fault scarp in the Sierra de Cordoba. The white areas in the basins are salt pans.

PLATE 3

Landsat MSS image of Sierra Pie de Palo and basins surrounding it. Note the arched profile of Sierra Pie de Palo demonstrated by the outwardly inclined exposures of the Gondwana erosion surface (indicated by arrows). Compare to figure 2. City of Caucete ("C") on southwest flank of range was heavily damaged in 1977 earthquake. Part of the faulted western flank of Sierra de Valle Fértil visible in northeast corner of scene.

axes of the basins show a minimum of 2000 to 4000 m structural relief across some of the major range-bounding faults.

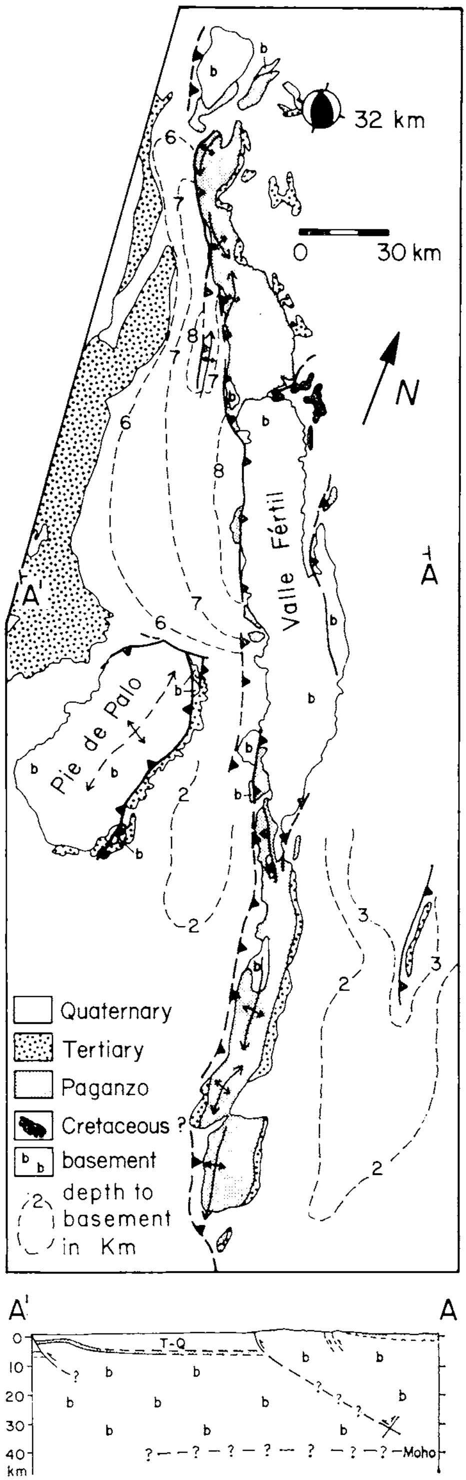
Refraction, seismic reflection, stratigraphic, and drilling data from the Bermejo-Tulum Valley system west of the Sierra de Valle Fértil (fig. 3) suggest that there are 2 to 8 km of Cenozoic and Upper Paleozoic strata overlying the basement surface in the lower plate of the 400 km long Sierra de Valle Fértil fault. Over the same distance, a variety of units, ranging from basement metamorphic rocks to units stratigraphically 4 to 5 km above the basement contact are exposed in the upper plate of the Sierra de Valle Fértil fault. Because the upper Paleozoic and Mesozoic strata did not cover the basement uniformly (A. Ortiz, personal commun., 1984), a similar crustal depth may be exposed along the whole length of the fault system. There has been an estimated maximum of 8 km of throw on the Sierra de Valle Fértil fault system, but it is not known whether all of it is of late Cenozoic age.

Geometry of range bounding faults.—In the field, range-bounding faults are commonly reverse faults with dips from 35° to 70° (Allmendinger and others, 1983b). Drilling through a 40 m thick overhang in one of the basement slices indicated a dip of about 55° (Lencinas, 1971).

A major unknown is the deep geometry of these faults. Earthquakes indicate that reverse motion occurs on faults at 15 to 32 km depth that dip 30° to 60° (Stauder, 1973; Chinn and Isacks, 1983; Kadinsky-Cade, ms). The middle and lower crustal earthquakes suggest that brittle or frictional deformation occurs on faults whose orientations are similar to those of the surface, but there are very little data relating those earthquakes to surface deformation. An earthquake and its aftershocks beneath Sierra Pie de Palo have been interpreted to indicate motion on a 35° dipping master fault between 17 and 25 km depth, but that did not continue to shallower levels (fig. 4B) (Kadinsky-Cade, ms; Kadinsky-Cade, Reilinger, and Isacks, 1985).

The commonly observed tilt of the upper Paleozoic erosional surface suggests rotation of crustal slivers rather than translation along planar thrust faults. Gonzalez Bonorino (1950) suggested that the reverse faults are listric at depth, and Erslev (1986) illustrated a similar geometry in constructing balanced cross sections of basement blocks in the Rocky Mountain foreland. If all the deformation occurs by rotation along the primary fault set rather than by folding or slip on small, closely spaced faults, the fault bounding Sierra de Ancasti would be horizontal at 10 to 16 km depth, and the fault bounding Sierra de Velasco would flatten at 15 to 20 km depth (fig. 4C, D). There are no focal mechanisms for crustal earthquakes located at appropriate sites to test this model (fig. 4C and D).

The 400 km long Sierra de Valle Fértil fault system may be a candidate for a surface fault that can be traced into the deep crust (fig. 3, 4A). The surface fault is defined by a set of north-northwest trending structures, of which the Sierra de Valle Fértil is the central morphological unit. In the middle 200 km long segment of the range, metamorphic basement is exposed across the 30 km width of the block, dipping about


32 km
0 30 km
N
A'
A
Valle Fértil
Pie de Palo
Quaternary
Tertiary
Paganzo
Cretaceous ?
basement
depth to basement in Km
T-Q
Moho
km

← **Fig. 3. Structure map of the Sierra de Valle Fértil mountain system and Sierra Pie de Palo, which form the western boundary of the southern and central Sierras Pampeanas. The exposed Tertiary strata west of Sierra de Valle Fértil crop out in the Eastern Precordillera, where they have been studied and dated by Johnson and others (1984) and Johnson and others (in press). Contours suggest depth to basement based on Ortiz and Zambrano (1981), Criado Roque, Mombru, and Ramos (1981), Kadinsky-Cade (ms), and Langer and Bollinger (in press). Earthquake focal mechanism from Stauder (1973) and hypocentral depth from Chinn and Isacks (1983). Surface mapping by Gentili (1970), Flores (1970), and Bossi (1976). In the cross section, the wedge of strata includes Ordovician limestone in its western part that is not exposed at the eastern end.**

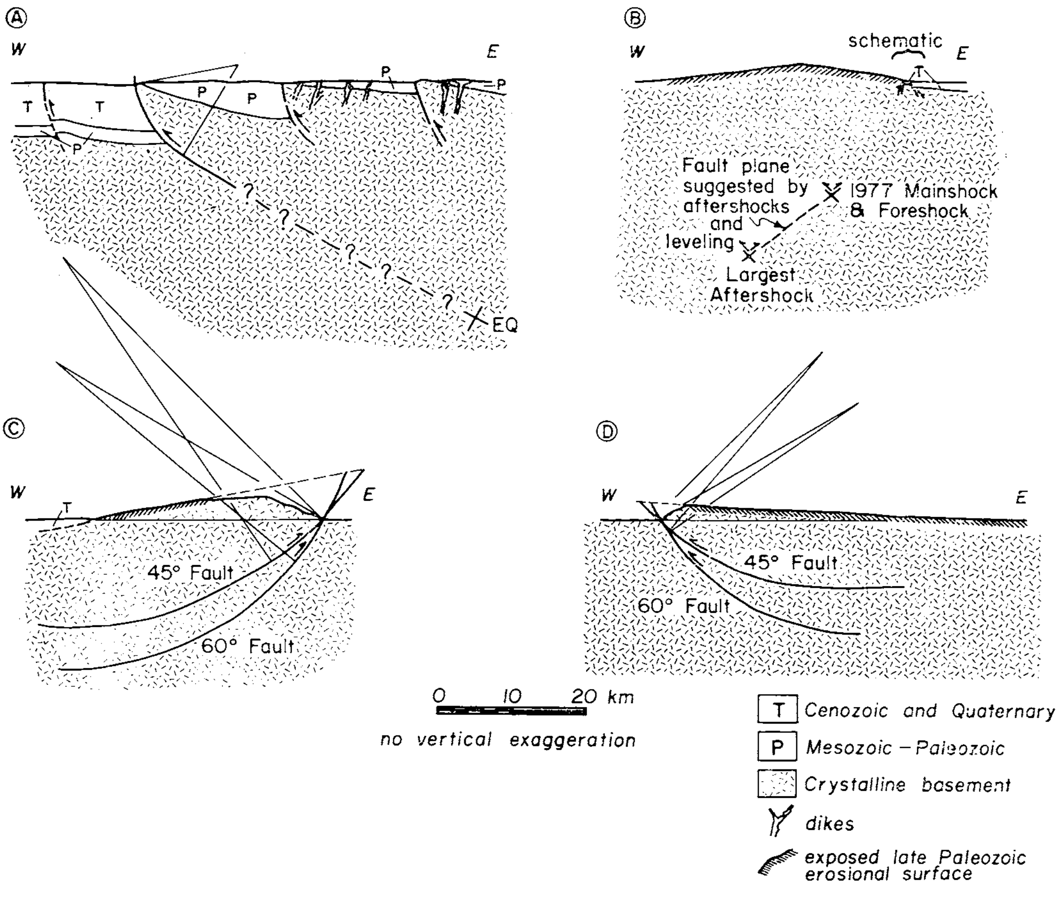
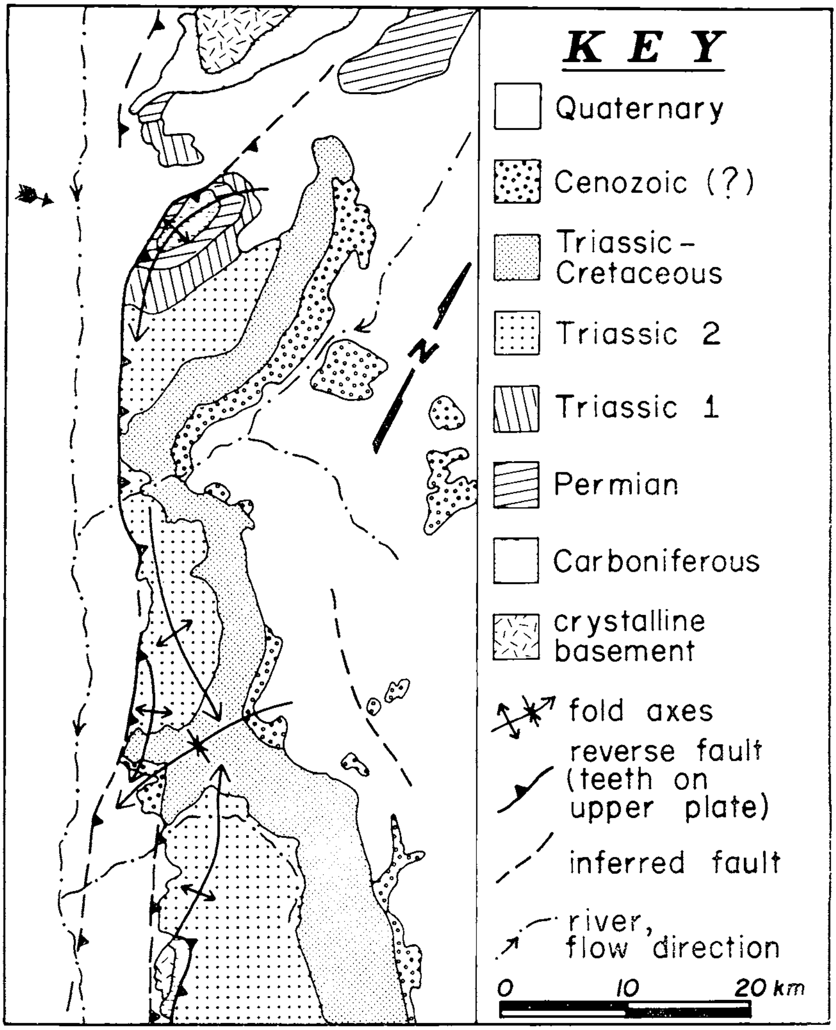


Fig. 4. Structural cross sections of major ranges in the Sierras Pampeanas, located on figure 2. (A) Sierra de Valle Fértil system along line of section of Ortiz and Zambrano (1981) and including earthquake data of Stauder (1973) and Chinn and Isacks (1983). Fault geometry reconstructed from rotational models which conserve volume. (B) Sierra Pie de Palo, after Kadinsky-Cade (ms). (C) Sierra de Velasco with subsurface geometry based on projecting intersection of fault plane and tilted erosion surface which caps the basement rocks, and restoring the tilted surface to an original assumed horizontal orientation while conserving volume (Erslev, 1986). Two possible faults are shown, depending on dip of fault at surface: 45° fault at surface would imply that listric fault is horizontal at depth of 15 km; 60° dip yields fault that flattens at 20 km depth. Depth to basement in footwall block not known, so constructions show a minimum amount of throw. (D) Construction of fault geometry beneath Sierra de Ancasti using same method as (C).

Fig. 5



A. Geologic map of drape folds in northern part of Sierra de Valle Fértil system (Sierra de Morada), after Gentili (1972).

5° eastward. Higher crustal levels (upper Paleozoic and Mesozoic strata) crop out at a few scattered sites along the western, faulted, margin (Ortiz and Zambrano, 1981). Both to the north and south, basement is replaced by a series of doubly plunging anticlines which expose the upper Paleozoic-Mesozoic section (fig. 5A) (Gentili, 1972; Criado Roque, Mombru, and Ramos, 1981; Servicio Geol. Nac., 1982). A 32 km deep earthquake located 45 km northeast of the surface trace of the Sierra de Valle Fértil fault system has nodal planes oriented parallel to the Sierra de Valle Fértil trend (Stauder, 1973; Chinn and Isacks, 1983) (fig. 3). First motion data

Fig. 5

B. Photo of western face of drape fold of Sierra de Morada. Direction of view of photo indicated by arrow in northwest corner of (A).

and S-wave polarizations define the nodal planes within a narrow range (Stauder, 1973). If that earthquake (m_b 5.9) were associated with the Valle Fértil fault system, it would indicate a 35° east-northeast dipping fault plane (in close accord with one of the two nodal planes) cutting through at least the upper 32 km of the crust. This structure would be similar in scale and continuity to the Wind River Fault (Smithson and others, 1979; Lynn, Quam, and Thompson, 1983). Modeling based on a gravity survey that crosses the fault and projects about 2 km into the range suggests a minimum 80° dip for the bounding fault at the surface (D. Snyder, personal commun., 1984). This surface dip and the tilt of the strata and exhumed basement surface suggest that the fault is curved in the shallow crust. To match the depth to basement on the downthrown side of the fault to the rotated basement contact on the upthrown side requires a combination of rotation and translation. If the deformation is inferred to occur only along a single fault, the geometry shown in figure 4A creates a minimum of space problems along the fault. It indicates that about 4.5 km of vertical motion is due to rotation on a curved fault, and 3.5 km is due to non-rotational translation. The fault may change in dip along strike, and there may be splays or other complexities to the fault geometry; a great deal more subsurface data are needed in order to clarify its geometry. A strikingly similar geometry, combining planar splays and curved splays, has been interpreted for the Wind River fault (Lynn, Quam, and Thompson, 1983; Sharry and others, 1986).

In detail, the range-bounding structures tend to be sets of parallel faults. At the surface, the splays are separated by gently rotated or broadly folded remnants of the upper Paleozoic erosion surface, forming a set of steps or platforms rising into the mountains (Acenolaza and Bortolotti, 1981; T. Jordan, P. Zeitler, and E. Fielding, unpub. field data).

Without better definition of the subsurface geometry of the reverse faults bounding the mountains, it is difficult to estimate the total throw on the faults. Given that most observations indicate moderate angles of dip (30°-60°), a first approximation suggests that horizontal throw is about equal to vertical throw. Thus each principal fault system represents a minimum of about 2 to 8 km of horizontal shortening of the continental crust. In the case of the Sierra de Valle Fértil fault system (figs. 3 and 4A), 8 km of horizontal shortening is estimated. These estimates suggest that total shortening of the Sierras Pampeanas is roughly 2 percent or between 10 and 20 km. In comparison, there has been roughly 7 percent shortening of the Rocky Mountain foreland province in Wyoming and Colorado (Bird, 1984).

Folds.—Two types and scales of flexures are recognized in the mountain blocks of the Sierras Pampeanas. The first involves layered sedimentary strata and is analogous in form and distribution to the "drape folds" of Wyoming. The second involves the crystalline basement, forming broad arches, perhaps analogous to the "Sherman Arch" east of the Laramie Range, the Rock Springs Uplift, or the San Rafael swell on the Colorado plateau.

Folds in strata are not common in most of the Sierras Pampeanas, primarily because there was little sedimentary cover over basement when the Cenozoic deformation began. Upper Paleozoic-Cenozoic strata and large-scale folds are most common in the western and northern Sierras Pampeanas.

In the northwestern Sierras Pampeanas, the folded strata occur along strike of basement faults, where the faults gradually lose throw, and the folds have wavelengths of 10 to 20 km (Allmendinger, 1986). In the Sierra de Valle Fértil trend, the anticlines have half-wavelengths of 30 to 40 km, amplitudes up to 4000 m, and are strongly asymmetric (figs. 3, 5). The hinge lines tend to be folded, creating series of plunging folds.

The arches in crystalline basement are best displayed on Sierra Pie de Palo (pl. 3 and fig. 3) and in the northern projection of Sierra los Llanos (fig. 2). The broad arch form is revealed by tracing apparent remnants of the upper Paleozoic erosion surface, using Landsat images. The Sierra Pie de Palo arch is approx 35 km wide and continues 80 km along strike (the total dimensions of the range) with a minimum of 3.0 km of relief; it is somewhat asymmetric to the east (fig. 3). The geometry and amount of offset of range-bounding faults paralleling the arch axis are still poorly known (Whitney and Bastias, 1984; Kadinsky-Cade, Reilinger, and Isacks, 1985). It is likely that the arch represents a broad fold above a blind reverse fault (Kadinsky-Cade, ms) (fig. 4B). The basement arch in Sierra los Llanos is exposed for about 70 km along strike, plunging gently northward to create a conical shape which pinches out northward from about 25 km width at the south end. The minimum relief at the southern end is 1000 m. The arch is asymmetric eastward and in part bounded on the east side by a reverse fault (fig. 2). These examples may represent early stages of Berg's (1962) fold-thrust model of basement faulting in the Laramide province.

Little is known about the mechanism by which the basement is arched. Much of the Sierra Pie de Palo is composed of strongly layered low to medium grade metasedimentary rocks (Dalla Salda and Varela, 1984), suggesting the possibility that layer parallel slip has accommodated the folding. The Sierra los Llanos, however, has a large proportion of tonalite and granodiorite plutons (Caminos, 1979; Servicio Geol. Nac., 1982).

Deformation in the basins.—By analogy with the early Cenozoic Rocky Mountain foreland deformation of Wyoming and Colorado, one would seek highly deformed syntectonic deposits in at least some of the basins of the Sierras Pampeanas. However, for the most part, deformation in the broader basins is not present or has not been recognized. One exception is found in the basin on the east side of the Sierra de Valle Fértil (part of this area is included in pl. 2 and fig. 3). This basin is reported to have subsided more than other basins in the central and eastern Sierras Pampeanas (Criado Roque, Mombru, and Ramos, 1981). Short wavelength folds (3000-4000 m) and faults can be traced up to 60 km in length in various parts of the basin (J. Beer, 1985, personal commun.). Elsewhere, faults are also reported to cut Quaternary sediments 20 to 30 km basinward from the range-bounding faults in the southern half of the Pipanaco basin (fig. 2) (Castano, 1976; Sosic, 1973).

DATING CONSTRAINTS

The younger limit of deformation in the Sierras Pampeanas is the Holocene, but it is difficult to determine the time at which deformation began. Faults cutting the alluvial fans and Quaternary sediments, or the development of multiple pediment levels, seem to be common to all the mountain block, such as examples at the margins of the Sierras de Aconquija (Strecker and others, 1984), Ambato (Gonzalez Diaz, 1974), Velasco (Castano, 1976), Hualfin (Allmendinger, 1984), Valle Fértil and Pie de Palo (Whitney and Bastias, 1984)

In most cases, the youngest strata deformed coherently with the Cenozoic section are not dated. Two exceptions occur in the northwestern Sierras Pampeanas. Marshall and others (1979) dated an ash near the top of a thick section at Corral Quemado as 3.52 Ma. Allmendinger (1984 and 1986) showed that this section, plus about 750 m of strata overlying that ash, are folded around a plunging basement block. Similarly, a 3.5 Ma ash is found near the top of a Cenozoic section folded during the uplift of Sierra Aconquija (Strecker and others, 1984). Because both these uplifts are located in the extreme northern part of the Sierras Pampeanas, a region that is transitional into other tectonic provinces, it is not known whether their chronology can be extrapolated to other parts of the Sierras Pampeanas.

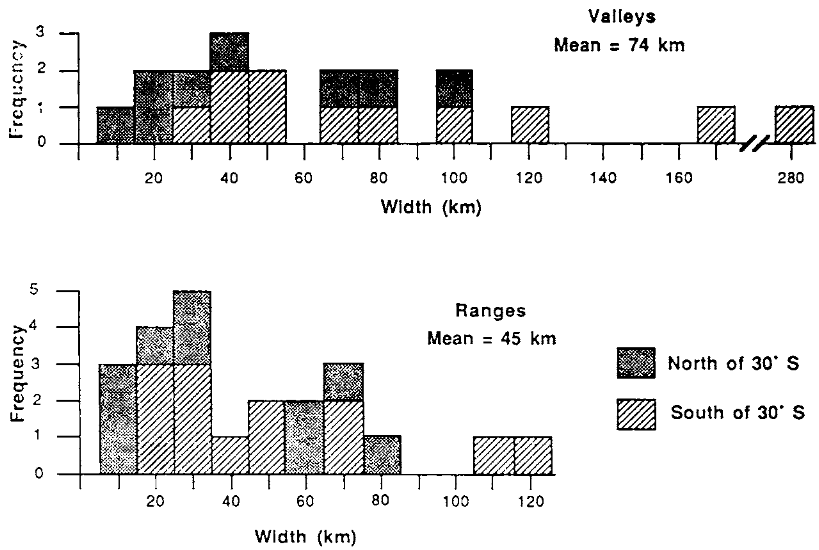
Elsewhere, Cenozoic strata are cut by the range-bounding fault systems or are folded, as at the Sierras de Velasco (Sosic, 1973; Castano, 1976), Cordoba (Lencinas, 1971), Famatina (Turner, 1971), Pie de Palo (Cuerda, Varela, and Iniguez, 1983), and Valle Fértil (Gentili, 1972; Flores, 1970 and 1979). This is consistent with the chronology noted in the north, with deformation restricted to the Pliocene to Recent, but the stratigraphic correlations are not tightly constrained. At the southern end of the Sierra de Valle Fértil structure, mammal fossils indicate an age of 9 to 11 Ma for deformed strata in the upper 150 m of the 500 m thick, folded, Cenozoic section (figs. 2, 3) (Flores, 1979; Pascual and Bondesio, 1981). Ongoing chronological studies in these strata promise to clarify the age at which deformation began (Jordan and others, 1985; Johnson and others, in press).

Another potential source of chronological control in the Sierras Pampeanas are Miocene-Pliocene volcanic centers located in the Sierras de Cordoba and San Luis (fig. 2). These calcalkaline trachyandesites, andesites, quartz latiandesites, and quartz latites form resistant volcanic necks associated with tuffaceous sediments (Llambias and Brogioni, 1981; Gordillo and Linares, 1981; Kay and Gordillo, 1985). The bodies in the Sierra de Cordoba vary in age from 4.7 ± 0.3 to 7.9 ± 0.6 Ma (Gordillo and Linares, 1981). Mammal fossils in the tuffaceous sediments show that the volcanic rocks in Sierra de San Luis are correlative with those in the Sierra de Cordoba (Pascual and Bondesio, 1981). The volcanic edifices in the Sierra de Cordoba are faulted, and yet their distribution suggests that their emplacement postdates some of the faults (Kay and others, in review). In Sierra de San Luis mapping has not yet shown whether the main uplift of the ranges predates or postdates the volcanic activity (Gonzalez Diaz, 1981).

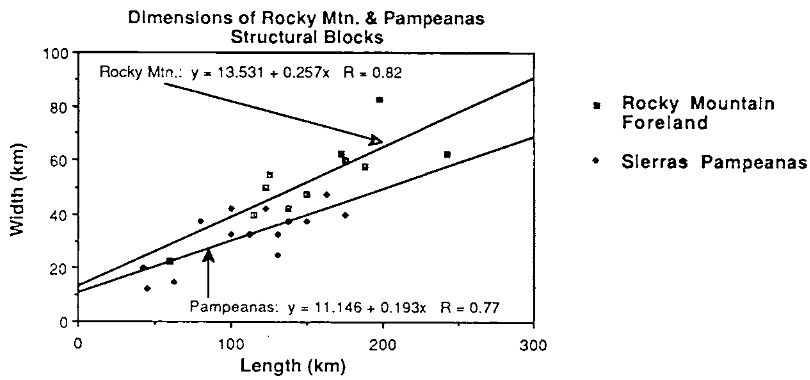
In summary, the deformation of the Sierras Pampeanas has spanned the Quaternary. A reasonable but unproven maximum age at which this deformation began is about 10 Ma, given the character of the sedimentary rocks in the few places around the periphery of the province where they are dated (Butler and others, 1984; Jordan and Alonso, in press; Johnson and others, in press). A reasonable but unproven minimum age at which the deformation began is about 3 Ma. In either case, the duration of deformation (3 to 10 Ma) is considerably shorter than the 35 to 40 Ma of deformation in the Rocky Mountain foreland province (Love, 1960).

CRUSTAL RHEOLOGY AND LONG WAVELENGTH DEFORMATION

There are no refraction data that constrain the Moho depth in the Sierras Pampeanas, but elevations and earthquake data suggest normal continental crustal thicknesses. Average elevations in the broader basins are approx 300 to 400 m, reaching 700 m in the northern part of the province. For an Airy isostatic model with mantle density of 3.3 gm/cm^3, crust of 2.8 gm/cm^3, and 33 km thick crust at sealevel, this suggests that the crust in the Sierras Pampeanas is about 35 km thick. This computation does not take into account any possible buoyancy effects due to the underlying flat Nazca slab, such as regional subsidence predicted by Cross and Pilger (1978). However, elevations several hundred kilometers to the east of the Sierras Pampeanas (and of any buoyancy problems related to the flat slab) average about 100 m, implying a crustal thickness of approx 33 km. Therefore the predicted crustal thickness in the Sierras Pampeanas is similar to that in the adjacent, undeformed craton. Gravity data (Introcaso, 1980) suggest that crustal thickness increases toward the west.



A.



B.

Fig. 6. (A) Summary of widths of valleys and widths of ranges, measured across 7 east-west transects spaced at intervals of 1° of latitude, in the Sierras Pampeanas between 27° and 33°S. (B) Length-to-width ratios of the structural blocks of the Sierras Pampeanas compared to those of the Rocky Mountain foreland. Note that the ranges measured in (A) may be composed on more than one of the structural blocks measured in (B).

The principal ranges of the Sierras Pampeanas vary in east-west width from 10 to 120 km, with a mean width of about 45 km (fig. 6A). The broad, flat valleys vary in width from 10 to 280 km, with a mean of about 74 km (fig. 6A). About 40 percent of the area is mountain, and 60 percent is valley, perhaps suggesting that even though the basement deformation is continuing, the positive structures have been partly buried by synorogenic deposits. In sum, the mean east-west wavelength of the deformation, parallel to the shortening direction, is about 120 km; the wavelength seems to increase from west to east and from north to south (fig. 2).

A comparison of the length-to-width ratios of the Sierras Pampeanas uplifts shows that they have similar ratios but are generally smaller than the Rocky Mountain foreland uplifts (fig. 6B). The difference in size may well be due to the younger and less advanced nature of Sierras Pampeanas deformation, even though the similar ratios indicate that the mechanical processes of thick-skinned deformation in the two regions may be the same.

Fletcher (1984) derived an analytical model that related wavelength of a deformation (S) to the thickness of a brittle layer (H) and showed that $S/H \approx 4\text{-}6$. Schmidt and others (1985) and Kulik and Schmidt (in press) used that relation and spacing of the major uplifts in the basement-uplift province of Montana and Wyoming as a measure of the subsurface structure and the mechanical properties of the crust. They reason that the basement-uplifts are detached from the lower crust across a rheological boundary, the brittle-ductile transition. To the extent that the brittle-ductile transition is thermally controlled, regions of differing geothermal gradients should have different spacing between ranges. Schmidt and others (1985) and Kulik and Schmidt (in press) estimate that the depth to the brittle-ductile transition in the Rocky Mountains ranges from a minimum of 11 to 16 km for the western Montana zone, to 17 to 25 km for the western Wyoming zone, to 25 to 38 km for the average Wyoming area. Fletcher's (1984) relation applied to the Sierras Pampeanas would suggest that the average depth to the brittle-ductile transition there is presently ~ 20 to 30 km, with shallower depths in the western and northern parts of the region, probably due to elevated geotherms.

Crustal earthquakes provide a more direct and less assumption-dependent measure of the depth to the brittle-ductile transition, at least for rapid strain rate conditions. The western Sierras Pampeanas, where the deformation wavelength would suggest a depth of 11 to 16 km for the brittle-ductile transition, has abundant seismicity with well determined hypocenters for major events ($Ms \geqslant 7$) between 15 to 32 km and with hypocenters for lesser events as deep as 40 km (Chinn and Isacks, 1983; Kadinsky-Cade, ms; Triep and Cardinali, 1984; Kadinsky-Cade, Reilinger, and Isacks, 1985; R. Smalley, unpub. data, 1986). Thus, much of the crust in the western Sierras Pampeanas appears to be deforming in a brittle-like manner. The analytical relation appears to be a poor predictor for the Sierras Pampeanas, and, if the analogy between the Sierras Pampeanas and the Rocky Mountain foreland is correct, it may inadequately describe the Rocky Mountain deformation as well. Although earthquakes occur elsewhere in the Sierras Pampeanas, their depths are not well known because of the lack of local seismic networks.

IMPLICATIONS TO MODELS OF ORIGIN OF THE ROCKY MOUNTAIN FORELAND PROVINCE

Several authors have suggested that the Central Andes Mountains are a modern analogue of the Sierra Nevada-Rocky Mountain orogenic belt during the Late Cretaceous and early Cenozoic (fig. 1) (Sales, 1968; Coney, 1976; Dickinson and Snyder, 1978). This comparison has been elaborated upon by Jordan and others (1983a), who emphasized the similarity between the Sierras Pampeanas and the "Laramide" basement uplifts of the Wyoming, southern Montana, and Colorado Rocky Mountains.

The manner in which the crust of foreland basement-uplift areas is coupled to adjacent areas and to the subducting plate is not fully understood. For the Rocky Mountain foreland, Sales (1968), Hamilton (1981), Chapin and Cather (1981), and Gries (1983a) have proposed models that explain the complex directions of shortening in terms of relative motions of the Wyoming-Colorado foreland province against neighboring crustal provinces. In the Sierras Pampeanas it is clear that shortening direction is parallel to the plate convergence direction, and through-going faults do not seem to bound its north, south, or eastern limits, so coupling to adjacent parts of South America is not an issue.

The other dimension of coupling is between the two converging plates. Cross and Pilger (1982) discussed the consequences of efficient stress transmission from the convergent plate boundary to the upper plate in cases of very low-angle subduction. They pointed out that it might be due to either mechanical coupling due to shear at an interface or to hydrodynamic coupling with the asthenosphere. Bird (1984) recently presented a mechanical analysis of the early Cenozoic shear between the flat-subducting Farallon plate and North American lithosphere. He inferred that the lower crust is a mechanically weak zone, capable of significant ductile flow that in the long term can be thickened and/or thinned by shear transport. For the Rocky Mountain foreland, it was thickened from an initial cratonic thickness of about 35 km to its present thickness of about 45 km. If the structure of the North American craton in the Cretaceous were similar to that of the Sierras Pampeanas today, it would seem that the lower crust must not be weak until a depth greater than 35 ± 5 km.

It has also been suggested that the entire region underlain by the Farallon flat slab in the Late Cretaceous-early Cenozoic underwent subsidence, as an isostatic response to the doubled thickness of lithosphere. This has been proposed as an explanation for anomalous subsidence of the regional basin in which the Lance Formation accumulated (Cross and Pilger, 1978). Given the lack of direct measurements of the thickness of the crust and lithosphere in the Sierras Pampeanas, it is impossible to determine directly what its average elevation should be in the absence of such postulated buoyancy effects. However, there is no record of widespread marine incursion, and the only thick accumulations of Cenozoic strata occur in the localized basins. Thus, for lack of evidence, it seems that any regional subsidence due to the added thickness of underlying Nazca lithosphere has been minor (a few 100 m) or non-existent.

CONCLUSION

The Sierras Pampeanas, a broad region of basement deformation in the eastern half of the Andean orogenic belt, serve as an example of foreland basement deformation. Reverse faults that segregated the cratonic basement into tilted range blocks and broad basins have been active during the late Cenozoic, apparently for not longer than the last 10 Ma. Their development appears to be coincidental with the time during which the subducting Nazca plate developed its present sub-horizontal trajectory. The continental basement of the Sierras Pampeanas has been shortened in an east-west sense, parallel to the plate convergence direction, by an estimated 2 percent during that time.

Because the data base for the Sierras Pampeanas is quite different than that of the older Rocky Mountain foreland province, we are able to glean new information about the dynamics of the well-described deformation in the Rocky Mountain province. The analogy between Sierras Pampeanas and Rocky Mountain structures supports the interpretation of other workers that the Rocky Mountain province of the United States was probably also deformed by horizontal compression, rather than by differential vertical stresses.

One of the most interesting observations is that earthquake focal mechanisms in the Sierras Pampeanas indicate moderate-angle reverse faulting at mid to lower crustal depths. This suggests that faults such as the Wind River Fault system, which have been traced to 25 to 35 km depth by reflection seismology, probably were brittle or frictional to those depths. Thus the brittle/ductile transition for relatively cold cratonic basement is probably deeper than 30 km.

ACKNOWLEDGMENTS

We wish to thank Drs. V. A. Ramos, B. L. Isacks, A. Ortiz, M. V. Pensa, and A. L. Bloom for numerous interesting discussions of the Sierras Pampeanas. This paper is an outgrowth of a cooperative program between Cornell University and the Servicio Geologico Nacional of Argentina, supported by the donors to the Petroleum Research Fund of the American Chemical Society (grant #13297-AC2), COPSTOC (Cornell Program for the Study of the Continents), and NSF grants EAR-8206787, EAR-8218067, and EAR-8206178. INSTOC Contribution No. 32.

References

Acenolaza, F. G., and Bortolotti, P., 1981, Estratigrafía y evolución tectónica de la quebrada de La Rioja: Acta Geol. Lilloana, v. 15, no. 3, p. 31-39.

Acenolaza, F. G., Toselli, A. J., Durand, F. R., and Taddei, R. D., 1982, Geología y estructura de la region norte de Andalgalá, Provincia de Catamarca: Acta Geol. Lilloana, v. 16, p. 121-139.

Allmendinger, R. W., 1984, Estructuras transversales de la zona de transición entre 26° y 27° S, Provincias de Tucumán y Catamarca, Argentina: resultados preliminares: Argentina Geol. Cong., 9th, v. 2, p. 31-47.

——— 1986, Kinematic and tectonic history of the southern margin of the Puna Plateau, northwest Argentina: Geol. Soc. America Bull., v. 97, p. 1070-1082.

Allmendinger, R., Jordan, T., Ramos, V., and Strecker, M., 1983b, Foreland structures associated with flat subduction, Argentine Andes: Geol. Soc. America, Abs. with Programs, v. 15, p. 514.

Allmendinger, R., Ramos, V. A., Jordan, T. E., Palma, M., and Isacks, B., 1983a, Paleogeography and Andean structural geometry, northwest Argentina: Tectonics, v. 2, p. 1-16.

Azcuy, C. L., 1975, Palinología estratigráfica de la cuenca Paganzo: Asoc. Geol. Argentina Rev., v. 30, p. 104-109.

Barazangi, M., and Isacks, B. L., 1976, Spatial distribution of earthquakes and subduction of the Nazca plate beneath South America: Geology, v. 4, p. 686-692.

Berg, R. R., 1962, Mountain flank thrusting in Rocky Mountain foreland, Wyoming and Colorado: Am. Assoc. Petroleum Geologists, v. 46, p. 2019-2032.

Bevis, M., and Isacks, B., 1984, Hypocentral trend surface analysis: Probing the geometry of Benioff zones: Jour. Geophys. Research, v. 89, no. B7, p. 6153-6170.

Bird, P., 1984, Laramide crustal thickening event in the Rocky Mountain foreland and Great Plains: Tectonics, v. 3, p. 741-758.

Bossi, G., 1976, Geología de la cuenca de Marayes-El Carrizal (provincia de San Juan, República Argentina): Argentina Geol. Cong., 6th, v. 1, p. 23-38.

Butler, R. F., Marshall, L. G., Drake, R. E., and Curtis, G. H., 1984, Magnetic polarity stratigraphy and ^{40}K-^{40}Ar dating of Late Miocene and Early Pliocene continental deposits, Catamarca province NW Argentina: Jour. Geology, v. 92, p. 623-636.

Caelles, J. C., ms, 1979, The geological evolution of the Sierras Pampeanas massif, La Rioja and Catamarca Provinces, Argentina: Ph.D. dissert., Queens Univ., Ontario, Canada, 493 p.

Caelles, J. C., Clark, A. H., Farrar, E., McBride, S. L., and Quirt, S., 1971, Potassium-argon ages of porphyry copper deposits and associated rocks in the Farallón Negro-Capillitas district, Catamarca, Argentina: Econ. Geology, v. 66, p. 961-964.

Cahill, T., and Isacks, B., 1985, Shape of the subducted Nazca plate: EOS (AGU Trans.), v. 66, p. 299.

Caminos, R., 1979, Sierras Pampeanas noroccidentales, Salta, Tucumán, Catamarca, La Rioja y San Juan, *in* Segundo Simposio de Geología Regional Argentina: Córdoba, Acad. Nac. Cienc., v. 1, p. 225-292.

Caminos, R., Cingolani, C. A., Herve, F., and Linares, E., 1982, Geochronology of the pre Andean metamorphism and magmatism in the Andean Cordillera between latitudes 30° and 36°S: Earth Sci. Rev., v. 18, p. 333-352.

Castano, O. M., 1976, Aspectos geológicos y hidrogeológicos de la Zona de la Costa, Departamentos Castro Barros y Arauco, La Rioja: Asoc. Geol. Argentino Rev., v. 3, p. 11-22.

Chamberlin, R. T., 1919, The building of the Colorado Rockies: Jour. Geology, v. 27, p. 145-251.

Chapin, C. E., and Cather, S. M., 1981, Eocene tectonics and sedimentation in the Colorado Plateau-Rocky Mountain area, *in* Dickinson, W. R., and Payne, W. D., eds., Relations of Tectonics to Ore Deposits in the Southern Cordillera: Arizona Geol. Soc. Digest, v. 14, p. 173-198.

Chinn, D., and Isacks, B. L., 1983, Accurate focal depths of shallow earthquakes near the New Hebrides and South American subduction zones: Tectonics, v. 2, p. 529-563.

Coney, P. J., 1976, Plate tectonics and the Laramide orogeny: New Mexico Geol. Soc. Spec. Pub. 6, p. 5-10.

Corvalan, J., chairman, 1981, Plate-tectonic map of the Circum-Pacific region, Southeast quadrant (scale 1:10,000,000): Tulsa, Okla., Am. Assoc. Petroleum Geologists.

Criado Roque, P., Mombru, C. A., and Ramos, V. A., 1981, Estructura e interpretación tectónica, *in* Relatorio, Geología y recursos naturales de la provincia de San Luis: Argentino Geol. Cong., 8th, p. 155-192.

Cross, T. A., and Pilger, R. H., Jr., 1978, Tectonic controls of Late Cretaceous sedimentation: Nature, v. 274, p. 653-657.

——— 1982, Controls of subduction geometry, location of magmatic arcs, and tectonics of arc and back-arc regions: Geol. Soc. America Bull., v. 93, p. 545-562.

Cuerda, A. J., 1973, Sierras Pampeanas, una nueva interpretación de su estructura: Asoc. Geol. Argentina Rev., v. 28, p. 293-303.

Cuerda, A. J., Cingolani, C. A., Varela, R., and Schauer, O. C., 1981, Geología de la Sierra de Mogna, Provincia de San Juan: Argentina Geol. Cong., 8th, v. 3, p. 139-158.

Cuerda, A., Varela, R., and Iniguez, A. M., 1983, Secuencia salifera en el Terciario de Nikizanga, Sierra Pie de Palo, Provincia de San Juan: Asoc. Geol. Argentina Rev., v. 38, p. 412-420.

Dalla Salda, L. H., and Varela, R., 1984, El metamorfismo en el tercio sur de la Sierra Pie de Palo, San Juan: Asoc. Geol. Argentina Rev., v. 34, p. 68-93.

de Alba, C., 1979, Sistema del Famatina, *in* Leanza, A. F., ed., Geología Regional Argentina: Córdoba, Acad. Nac. Cienc., v. I, p. 349-396.

Dickinson, W. R., and Snyder, W. S., 1978, Plate tectonics of the Laramide orogeny, *in* Matthews, V., III., ed., Laramide Folding Associated with Basement Block Faulting in the Western United States: Geol. Soc. America Mem. 151, p. 355-366.

Engebretson, D., Cox, A., and Thompson, G. A., 1984, Correlations of plate motion with continental tectonics: Laramide to Basin and Range: Tectonics, v. 3, p. 115-120.

Erslev, E. A., 1986, Basement balancing of Rocky Mountain foreland uplifts: Geology, v. 14, p. 259-262.

Fletcher, R. C., 1984, Instability of lithosphere undergoing shortening: a model for Laramide foreland structures [abs.]: Geol. Soc. America, Abs. with Programs, v. 16, p. 83.

Flores, M., 1970, Cuenca de San Luis, *in* Leanza, A. F., ed., Geología Regional Argentina: Córdoba, Acad. Nac. Cienc., p. 567-579.

Flores, M. A., 1979, Cuenca de San Luis, *in* Segundo Simposio de Geología Regional Argentina: Córdoba, Acad. Nac. Cienc., v. 1, p. 745-770.

Gentili, C., 1972, Descripción geológica de la Hoja 17c, Cerro Rajado: Serv. Nac. Min. Geol., Bol. 131, 62 p.

González Bonorino, F., 1950, Algunos problemas geológicos de las Sierras Pampeanas: Asoc. Geol. Argentina Rev., v. 5, no. 3, p. 81-110.

González Diaz, E. F., 1974, Superficies de erosión (abanicos rocosos) exhumadas en el flanco occidental de la Sierra de Ambato, al sur de la Quebrada de la Cebila (La Rioja): Asoc. Geol. Argentina Rev., v. 29, p. 5-22.

——— 1981, Geomorfología: *in* Geología de la Provincia de San Luis, Relatorio: Argentina Geol. Cong., 8th, p. 193-236.

Gonzalez-Ferran, O., Baker, P. E., and Rex, D. C., 1985, Tectonic-volcanic discontinuity at latitude 27° South Andean Range, Associated with Nazca plate subduction: Tectonophysics, v. 112, p. 423-441.

Gordillo, C. E., and Linares, E., 1981, Geocronología y petrografía de las vulcanitas terciarias del departamento Pocho, provincia de Córdoba: Asoc. Geol. Argentina Rev., v. 36, p. 380-388.

Gries, R., 1983a, Oil and gas prospecting beneath Precambrian of foreland thrust plates in Rocky Mountains: Am. Assoc. Petroleum Geologists Bull., v. 67, no. 1, p. 1-28.

——— 1983b, North-south compression of Rocky Mountain foreland structures, *in* Lowell, J. D., ed., Rocky Mountain Foreland Basins and Uplifts: Rocky Mountain Assoc. Geologists, p. 9-32.

Hamilton, W., 1981, Plate-tectonic mechanism of Laramide deformation: Contr. Geology, v. 19, p. 87-92.

Henderson, L. J., Gordon, R. G., and Engebretson, D. C., 1984, Mesozoic aseismic ridges on the Farallon plate and southward migration of shallow subduction during the Laramide orogeny: Tectonics, v. 3, p. 121-132.

Introcaso, A., 1980, Resultados gravimétricos en la banda latitudinal de Argentina central y paises vecinos: Inst. Panamericano de Geog. e Historia, Rev. Geofis., no. 12, p. 5-25.

Isacks, B. L., 1985, Topography and tectonics of the central Andes: EOS (AGU Trans.), v. 66, p. 375.

Johnson, N. M., Jordan, T. E., Naeser, C. W., and Johnsson, P. A., in press, Magnetic polarity stratigraphy, age, and tectonic setting of sediments in the eastern Andean foreland basin, San Juan Province, Argentina, *in* Homewood, P., Allen, P., and Williams, G., eds., Foreland Basins: Internat. Assoc. Sedimentologists.

Johnsson, P. A., Johnson, N. M., Jordan, T. E., and Naeser, C. W., 1984, Magnetic polarity stratigraphy and age of the Quebrada del Cura, Rio Jáchal, and Mogna Formations near Huaco, San Juan Province, Argentina: Argentina Geol. Cong., 9th, v. 3, p. 81-96.

Jordan, T. E., and Alonso, R., 1987, Cenozoic stratigraphy and basin tectonics of the Andes mountains, 20-28° South Latitude: Am. Assoc. Petroleum Geologists Bull., v. 71, no. 1.

Jordan, T. E., and Gardeweg, M., in press, Tectonic evolution of the late Cenozoic Central Andes, *in* Ben-Avraham, Z., ed., Mesozoic and Cenozoic Evolution of the Pacific Margins: Oxford Univ. Press.

Jordan, T. E., Isacks, B. L., Allmendinger, R. W., Brewer, J. A., Ramos, V. A., and Ando, C. J., 1983a, Andean tectonics related to geometry of subducted Nazca plate: Geol. Soc. America Bull., v. 94, p. 341-361.

Jordan, T. E., Isacks, B. L., Ramos, V. A., and Allmendinger, R. W., 1983b, Mountain building in the Central Andes: EPISODES, v. 1983, no. 3, p. 20-26.

Jordan, T. E., Naeser, C. W., Johnson, N. M., Johnsson, P. A., Johnson, A., Reynolds, J., Reynolds, S. A., and Fielding, E. J., 1985, Foreland basin evolution in the Central Andes, Bermejo basin, San Juan province, Argentina: Geol. Soc. America, Abs. with Programs, v. 17, p. 621.

Kadinsky-Cade, K., ms, 1984, Seismotectonics of the Chile margin and the 1977 Caucete earthquake of western Argentina: Ph.D. dissert., Cornell Univ.

Kadinsky-Cade, K., Reilinger, R., and Isacks, B., 1985, Surface deformation associated with the November 23, 1977, Caucete, Argentina, earthquake sequence: Jour. Geophys. Research, v. 90, p. 12691-12700.

Kay, S. M., and Gordillo, C. E., 1985, Expiration of volcanism over the Andean flat-slab: The 5-7 my Pocho volcanic field central Argentina: Geol. Soc. America, Abs. with Programs, v. 17, p. 624.

Kay, S. M., Maksaev, V., Moscoso, R., Mpodozis, C., Nasi, C., and Gordillo, C., in press, Tertiary Andean magmatism in Argentina and Chile between 28°-33°S: Correlation of magmatic chemistry with a changing Benioff Zone: Jour. South Am. Geology, v. 1.

Kulik, D., and Schmidt, C., in press, Region of overlap and styles of interaction of Cordilleran thrust belt and Rocky Mountain foreland, *in* Schmidt, C., and Perry, W. L., eds., Interaction of the Rocky Mountain foreland and the Cordilleran thrust belt: Geol. Soc. America Spec. Paper.

Langer, C. J., and Bollinger, G. A., in press, The western Argentina earthquake of November 23, 1977 — Spatial distribution of the aftershock sequence: Seismolog. Soc. America Bull.

Lencinas, A. N., 1971, Geología del Valle de Punilla entre Bialet Masse y la Cumbre, provincia de Córdoba: Asoc. Geol. Córdoba Bol., v. 1, no. 2, p. 61-70.

Lencinas, M. A., 1976, Edad de intrusivos cenozoicos en la Precordillera de San Juan y su implicancia estratigráfica: Asoc. Geol. Argentina Rev., v. 31, p. 53-58.

Llambías, E. J., and Brogioni, N., 1981, Magmatismo Mesozoico y Cenozoico, *in* Geología de la Provincia de San Luis Relatorio: Argentina Geol. Cong., 8th, p. 101-115.

Love, J. D., 1960, Cenozoic sedimentation and crustal movement in Wyoming: Am. Jour. Sci., v. 258-A, p. 204-214.

Lynn, H. B., Quam, S., and Thompson, G. A., 1983, Depth migration and interpretation of the COCORP Wind River, Wyoming, Seismic reflection data: Geology, v. 11, p. 462-469.

Maksaev, V., Moscosco, R., Mpodozis, C., and Nasi, C., 1984, Las unidades volcanicas y plutonicas del Cenozoico superior en la Alta Cordillera del Norte Chico (29°-31°S): Geología, alteración hidrotermal y mineralización: Geol. Chile Rev., v. 21, p. 11-51.

Marshall, L. G., Butler, R. F., Drake, R. E., Curtis, G. H., and Tedford, R. H., 1979, Calibration of the Great American Interchange: Science, v. 204, p. 272-279.

McNutt, R. H., Crocket, J. H., Clark, A. H., Caelles, J. C., Farrar, E. H., Haynes, S. J., and Zentilli, M., 1975, Initial $^{87}Sr/^{86}Sr$ ratios of plutonic and volcanic rocks of the central Andes between latitudes 26° and 29°S: Earth and Planetary Sci. Letters, v. 27, p. 305-313.

Minster, J. B., and Jordan, T. H., 1978, Present-day plate motion: Jour. Geophys. Research, v. 83, p. 5331-5354.

Mirré, J. C., 1971, Caracterización de una comarca de metamorfismo regional epizonal de alto grado: la sierra de Valle Fértil, provincia de San Juan, República Argentina: Asoc. Geol. Argentina Rev., v 26, no. 1, p. 113-127.

——— 1976, Descripción geológica de la hoja 19e, Valle Fértil: Serv. Nac. Minería Geología, Bol. 147, 71 p.

Munizaga, F., and Vicente, J. C., 1982, Acerca de la zonación plutónica y del volcanismo miocenico en los Andes de Aconcagua (lat. 32-33 S): datos radiométricos K-Ar: Geol. Chile Rev., no. 16, p. 3-21.

Ortiz, A., and Zambrano, J. J., 1981, La provincia geológica Precordillerana oriental: Argentina Geol. Cong., 8th, v. 3, p. 59-74.

Pascual, R., and Bondesio, P., 1981, Sedimentitas Cenozoicas, *in* Geología de la provincia de San Luis, Relatorio: Argentina Geol. Cong., 8th, p. 117-154.

Pilger, R. H., Jr., 1981, Plate reconstructions, aseismic ridges and low-angle subduction beneath the Andes: Geol. Soc. America Bull., P. I, v. 92, p. 448-456.

Pilger, R. H., 1984, Cenozoic plate kinematics, subduction and magmatism: South American Andes: Geol. Soc. London Jour., v. 141, p. 793-802.

Ramos, V. A., Jordan, T. E., Allmendinger, R. W., Kay, S. M., Cortes, J. M., Palma, M. A., 1984, Chilenia: un terreno alóctono en la evolución Paleozoica de los Andes centrales: Argentina Geol. Cong., 9th, v. 2, p. 84-106.

Ramos. V. A., Mahlburg Kay, S., Cingolani, C., and Kawashita, K., 1985, The volcanic rocks of Cerro Aconcagua — Cordillera Principal (32°S), Argentina: Univ. Chile, Comun., Spec. v. 35, p. 191-194.

Rapela, C. W., Mheaman, L., and McNutt, R. J., 1982, Rb/Sr geochronology of granitoid rocks from the Pampean Ranges, Argentina: Jour. Geology, v. 90, p. 574-582.

Sales, J. K., 1968, Crustal mechanics of Cordilleran foreland deformation: a regional and scale model approach: Am. Assoc. Petroleum Geologists Bull., v. 52, p. 2016-2044.

Salfity, J. A., and Gorustovich, S., 1983, Paleogeografía de la cuenca del Grupo Paganzo (Paleozoico superior): Asoc. Geol. Argentina Rev., v. 38, p. 437-453.

Schmidt, C. J., Evans, J. P., Fletcher, R. C., and Spang, J. H., 1985, Spacing of Rocky Mountain foreland arches and Laramide magmatic activity [abs.]: Geol. Soc. America, Abs. with Programs, v. 17, p. 710.

Servicio Geológico Nacional, 1982, Mapa Geológico de la República Argentina.

Sharry, J., Langan, R. T., Jovanovich, D. B., Jones, G. M., Hill, N. R., Guidish, T. M., 1986, Enhanced imaging of the COCORP seismic line, Wind River Mountains, *in* Barazangi, M., and Brown, L., eds., Reflection Seismology: A Global Perspective: Am. Geophys. Union, Geodynamics ser., v. 13, p. 223-236.

Smithson, S. B., Brewer, J. A., Kaufman, S., Oliver, J. E., and Hurich, C. A., 1979, Structure of the Laramide Wind River uplift, Wyoming, from COCORP deep reflection data and from gravity data: Jour. Geophys. Research, v. 84, p. 5955-5972.

Sosic, M. V. J., 1973, Descripción geológica de la hoja 14e, Salar de Pipanaco, provincias de Catamarca y La Rioja: Serv. Nac. Minería Geología, Bol. 137.

Spinelli, R. O., 1983, Edades potasio-argon de algunas rocas de la Sierra de Valle Fértil, provincia San Juan: Asoc. Geol. Argentina Rev., v. 38, p. 405-411.

Stauder, W. M., 1973, Mechanism and spatial distribution of Chilean earthquakes with relation to subduction of the oceanic plate: Jour. Geophys. Research, v. 78, p. 5033-5061.

Strecker, M. R., Bloom, A. L., Carrion, M., Villanueva, A., and Naeser, C., 1984, Piedmont terraces in the Valle de Santa Maria and in front of southwestern Sierra Aconquija, Prov. Catamarca, Tucumán and Salta, northwestern Argentina: Argentina Geol. Cong., 9th, v. 2, p. 448-465.

Toselli, A. J., Durand, F. R., Medina, M. E., Ramirez Vergaray, R., 1984, Caracteres petrológicos y geoquímicos del stock de las canadas, Provincia de Catamarca: Argentina Geol. Cong., 9th, v. 3, p. 185-204.

Toselli, A. J., Toselli, J. R. de, and Acenolaza, F., 1985, Milonitas de bajo grado de la megafractura de Sierras Pampeanas en la Quebrada de La Rioja, Sierra de Velasco, Argentina: Chileno Geol. Cong., 4th, v. 1, 2-159—2-171.

Triep, E. G., and Cardinali, C. B., 1984, Mecanismos de sismos en las Sierras Pampeanas occidentales: Argentina Geol. Cong., 9th, v. 3, p. 61-80.

Turner, J. C. M., 1971, Descripción geológica de la hoja 15d, "Famatina": Buenos Aires, Dirección Nac. Geología y Minería Bol., v. 126, 96 p.

Whitney, R. A., and Bastias, H., 1984, The Precordilleran active overthrust belt, San Juan Province, Argentina: Geol. Soc. America 1984, 19th Ann. Mtg., Reno, Nev., Western Geol. Excursions (Field Trip 17), p. 354-386.