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Seismotectonics of Thin- and Thick-Skinned Deformation in the Andean Foreland From Local Network Data: Evidence for a Seismogenic Lower Crust

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Local network data from San Juan, Argentina, provides new information about crustal seismicity in the Andean foreland above a horizontal segment of the subducted Nazca plate. We find two areas of foreland seismicity, one associated with the Sierras Pampeanas basement uplifts, and the other beneath, but not within, the Precordillera foreland fold-thrust belt. The Precordillera seismicity provides direct evidence for basement deformation beneath the sediments of the thrust belt and supports the idea that its eastern part is significantly modified by underlying basement deformation. In both areas, events are concentrated between 15 and 35 km depth and have volumetric, rather than planar, fault-like distributions. The depth distribution is unusually deep for intraplate earthquakes and suggests a brittle-ductile transition near 30-35 km. It also suggests the upper plate may have a geotherm similar to stable cratonic regions, rather than a steeper one that may be expected due to a lithospheric thinning implied by the location of the subducted plate. This can be understood as a transient effect of a Neogene flattening of the dip of the subducted plate, where a rapid thinning of the upper plate was followed with underplating by the cool subducted plate, leaving the upper plate geotherm unchanged while producing a thinned, mechanically weaker lithosphere. Finally, a spatial correlation of upper plate seismicity and structures with subducted plate seismicity raises the possibility that a major lithospheric structure cutting across the strike of the foreland near 31°S may be related to subduction of the Juan Fernandez ridge.

Introduction

Previous studies have shown that major features of both the South America and Nazca plates are segmented along-strike, and that the geography and tectonic development of the various segments are correlated [e.g., *Barazangi and Isacks*, 1976; *Jordan et al.*, 1983]. Figure 1 shows how along-strike variations in upper plate magmatism and tectonics correlate with along-strike changes in the dip of the subducted plate. Above a steeply dipping segment between 15° and 24°S the upper plate includes an active volcanic arc, the Altiplano-Puna plateau, and the Eastern Cordillera and Subandean Zone fold-thrust belts. Above a horizontal segment between 28° and 33°S the upper plate has an extinct volcanic arc, the narrow, thin-skinned Precordillera, and the wide zone of Sierras Pampeanas basement uplifts. The development of flat subduction, onset of both Pampean and Precordilleran deformation, and the shutdown of the volcanic arc are thought to coincide at 10 Ma [*Jordan et al.*, 1983; *Kay et al.*, 1987]. The tectonic relationship of the Pampeanas and Precordillera provinces is thought to be similar to that of the Mesozoic age Sevier and Laramide provinces of North America, which also developed over flat subduction [*Fielding and Jordan*, 1988; *Dickenson and Snyder*, 1978].

Associated with the subduction and mountain building is a high level of seismicity that consists of interplate slip events near the trench, intraplate events in the subducted plate, and shallow intraplate events in the overriding plate. The shallow activity also varies along-strike and is correlated with the along-strike tectonic variations of the upper plate. In the region of steep subduction, crustal seismicity occurs in a narrow band along the eastern margin of the Andes, with little seismicity in the Altiplano-Puna. In contrast, in the region of flat subduction, a broad region of the foreland, with both thin- and thick-skinned tectonic provinces, is highly active seismically (Figures 2 and 3). In this paper, local seismic network data from the flat slab region is used to study a small region of the foreland that straddles the boundary between the basement block uplifts of the Sierras Pampeanas and the thin-skinned Precordillera.

Geologic Background

Precordillera. The Precordillera, composed of Paleozoic clastics [*Baldis et al.*, 1982], is divided into three subprovinces, Western, Central, and Eastern (Figure 4) [*Ortiz and Zambrano*, 1981]. The Western Precordillera will not be discussed as it has little seismicity (Figures 2, 3, and 5). The Central Precordillera is an east verging, thin-skinned, fold-thrust belt [*Ortiz and Zambrano*, 1981; *Baldis et al.*, 1982] where uplift began by 13.5 Ma and may continue today [*Reynolds et al.* 1987; *Fielding and Jordan*, 1988]. The Eastern Precordillera has west verging thrusts south of, and folds north of, 31°S [*Baldis and Chebli*, 1969; *Fielding and Jordan*, 1988]. Shortening began less than 2.3 Ma and continues today [*Johnsson et al.*, 1984; *Whitney and Bastias*, 1984; *Whitney et al.*, 1984; *Bastias et al.*, 1984; *Johnson et al.*, 1986]. Between the Central and Eastern Precordilleras is the fault-bounded Matagusanos Valley (Figure 4), where an apparent 2.5-km downsection cut in the depth of the décollement in the direction of transport occurs [*Baldis and Bordonaro*, 1984].

Sierras Pampeanas. The Sierras Pampeanas are N-S trending basement block uplifts of Precambrian metamorphics [*Cingolani and Varela*, 1975; *Caminos et al.*,

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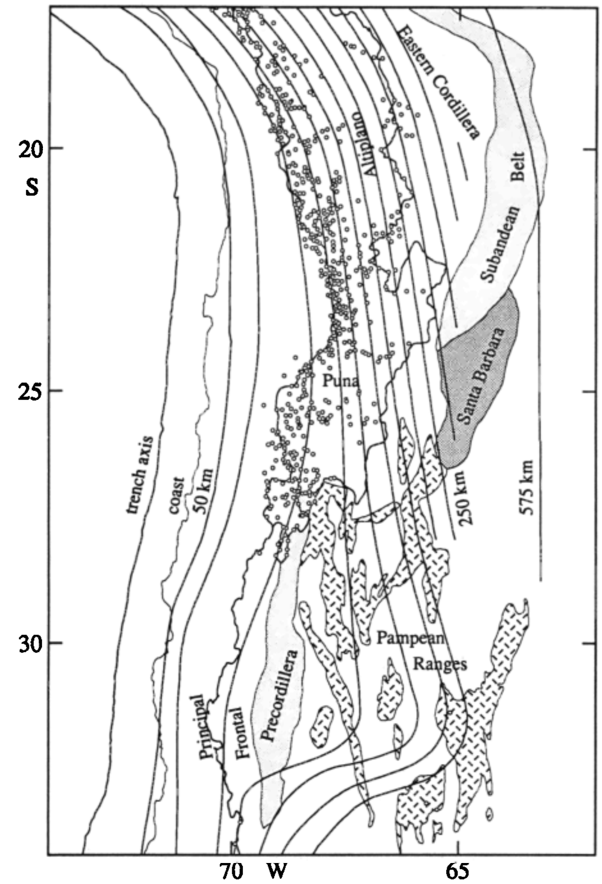

Fig. 1. Map of western South America showing upper plate tectonic features and volcanic arc. Contours of the Wadati-Benioff zone (WBZ) indicate the shape of the subducted plane. Note the smooth transition from steep to shallow between 24° and 28°S and the sharp but continuous change back to steep subduction near 32°S [*Cahill and Isacks*, 1985; *Isacks*, 1988; *Smalley and Isacks*, 1987]. The tectonic provinces of the upper plate [after *Jordan et al.*, 1983] show the corresponding along-strike segmentation of the upper plate. Neogene volcanos (open circles) [*Isacks*, 1988] show the magmatic arc over the steeply dipping segment and the absence of active volcanism over the flat segment. The drainage divide, which defines the Puna/Altiplano and then follows the crest of the Andes between the Principal and Frontal Cordilleras, is also shown.

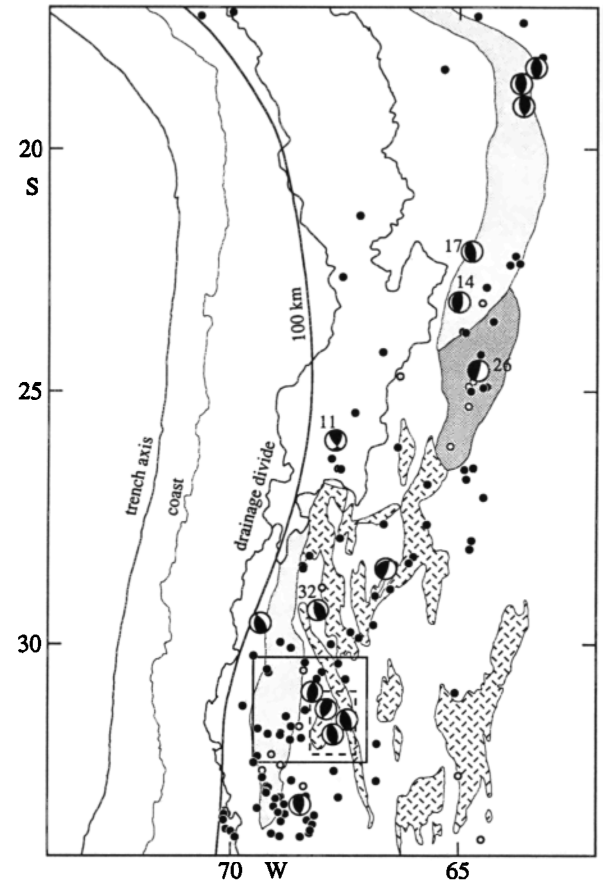

Fig. 2. Map of shallow intraplate seismicity of the Andean back arc (solid circles) and damaging historic earthquakes (open circles) [*Zamarbide and Castano*, 1978]. The epicenters are selected from the PDE for 1963 and 1983-1985, and the ISC for 1964-1982. The selected events were located using 15 or more *P* arrivals, have reported depths less than 50 km and are located east of the 100 km WBZ contour. The last requirement ensures they are contained within the South American lithosphere and are not interplate or WBZ events. Available focal mechanisms from *Stauder* [1973, 1975], *Chinn and Isacks* [1983], *Kadinsky-Cade* [1985], and *Dziewonski et al.* [1987] are also shown. The region of this study is contained in the solid box. The level of seismicity in the dashed box within the study area is very intense, so only earthquakes large enough to have focal mechanisms and accurate source depths from synthetic seismogram modelling are shown. The tectonic provinces of the South American plate are as in Figure 1.

1982] with a distinctive morphology that consists of a steep fault-bounded front side and a gently dipping back side. The fault is generally a moderately dipping thrust or reverse fault with evidence for large amounts of structural relief [*Jordan and Allmendinger*, 1986]. The back side is an erosional basement surface [*Rassmuss*, 1916; *Caminos*, 1979] produced in the late Paleozoic [*Dalla Salda and Varela*, 1984; *Toselli et al.*, 1985]. Wide, shallow basins of unmetamorphosed, relatively undeformed, Carboniferous and younger sediments separate the blocks [*Salfity and Gorustovich*, 1983].

Boundary region. The Bermejo and Tulum valleys separate the Pampean ranges from the Precordillera (Figure 4). The nature of the basement as one crosses from the Sierras Pampeanas to the Precordillera is a major unanswered question [e.g., *Fielding and Jordan*, 1988]. Several outcrops of Pampean basement too small to map (station RTCV is on one) occur in the Tulum Valley within 10 km of the Eastern Precordillera, but there are no known outcrops of basement farther west [*Zambrano*, 1969]. A melange that may be related to a proposed Paleozoic suturing of the two terranes [*Ramos et al.*, 1986] is exposed on the east side of the Eastern Precordillera [*Ortiz and Zambrano*, 1981]. The boundary between Pampean basement and the unknown, and probably allocthonous, basement beneath the Precordillera therefore occurs west of these outcrops.

FORELAND SEISMICITY: TELESEISMIC DATA

Shallow earthquakes in the south central Andean foreland (Figure 2), for which focal mechanisms are available, typically indicate E-W shortening on moderately dipping faults and have source depths of 8 to 40 km [*Chinn and Isacks*, 1983; *Suárez et al.*, 1983; *Kadinsky-Cade et al.*, 1985]. The depths and typical Andean foreland focal mechanisms of most events in the Subandean Zone indicate

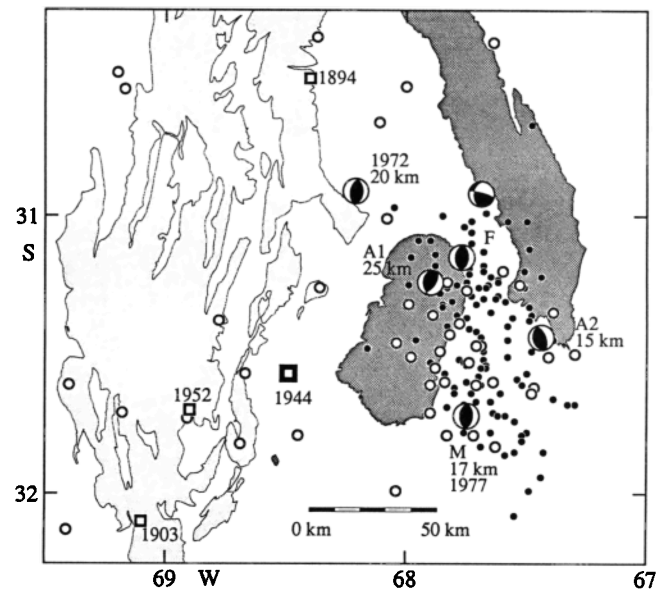

Fig. 3. Map of the San Juan region showing historical seismicity (dated open squares), selected ISC and PDE events (open circles), and aftershocks of the 1977 Caucete earthquake reported by the ISC (solid circles). The selection criteria for the ISC and PDE events are the same as in Figure 2. The aftershock period is November 23, 1977, to December 31, 1978. Focal mechanisms and depths for the 1977 foreshock (F), mainshock (M), two aftershocks (A1 and A2), and two additional events are also shown [*Chinn and Isacks*, 1983; *Kadinsky-Cade*, 1985; *Dziewonski et al.*, 1987]. The mountain ranges are shaded, light for the Precordillera and dark for the Sierras Pampeanas, while the intermountain valleys are in white.

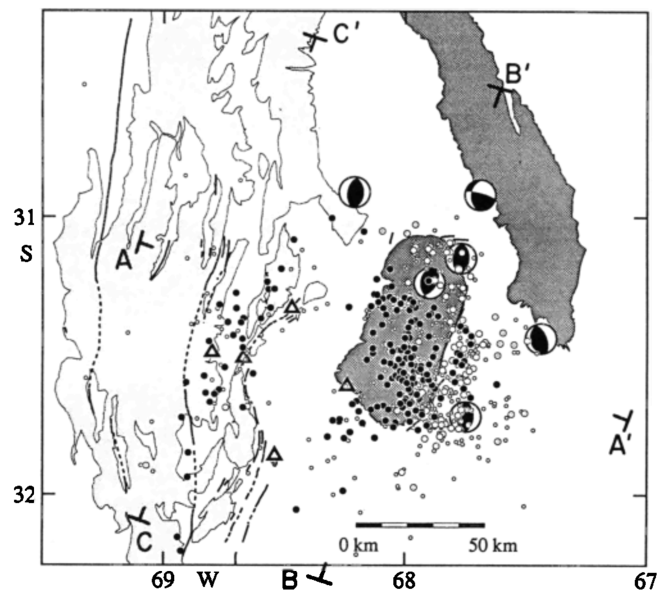

Fig. 5. Map of seismicity recorded by the local networks in 1980 and 1984-1985. Solid circles are class A events, large open circles are class B events, and small open circles are the remaining events. Major Quaternary fault systems associated with the Precordillera and Pie de Palo are also shown [*Bastias and Weidmann*, 1983]. Open triangles mark the stations of the INPRES network, and the focal mechanisms are as in Figure 2. A-A', B-B', and C-C' indicate cross sections shown in Figure 6.

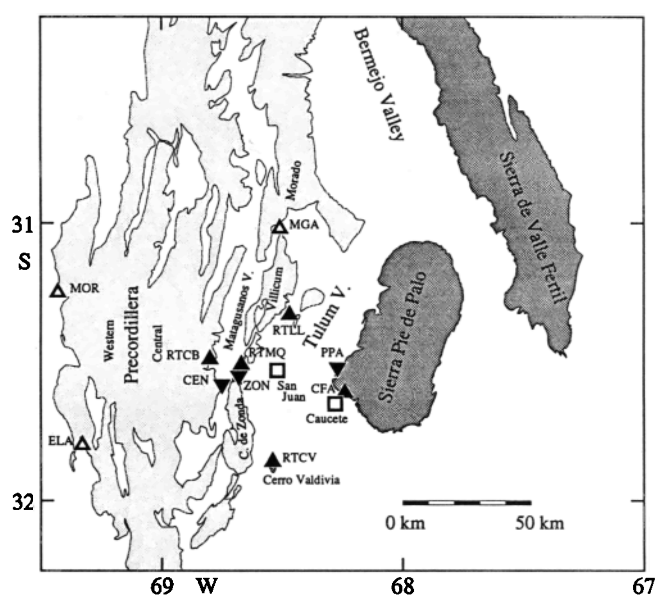

Fig. 4. Map of the San Juan region. The Eastern Precordillera is the chain of mountains along the eastern margin of the Precordillera (Sierras Chica de Zonda, Villicum, and Morado). The Matagusanos Valley separates the east verging Central Precordillera from the west verging Eastern Precordillera. The unlabeled valley to the west of the Precordillera is the Calingasta Valley. Stations and identification codes of the INPRES network (solid upright triangles), the UNSJ network (solid inverted triangles), and a 1-month INPRES experiment in June 1980 (open triangles) are also shown. The open boxes show the locations of the cities of San Juan and Caucete, which were heavily damaged by earthquakes in 1944 and 1977, respectively.

that they are associated with the deformation of Brazilian shield basement underthrust beneath the Subandean Zone [*Chinn and Isacks*, 1983; *Suárez et al.*, 1983], although a few events with atypical focal mechanisms and depths may have occurred on the inferred subhorizontal décollement of the thin-skinned Subandean Zone [*Chinn and Isacks*, 1983; *Jordan et al.*, 1983].

Sierras Pampeanas. Earthquakes in the Sierras Pampeanas are generally correlated with the basement uplifts (Figures 2 and 3) and have the typical Andean foreland focal mechanism. One such event is a 35-km-deep, Ms=6.5, event near the north end of Southern de Valle Fértil. Its east dipping nodal plane projects to the surface near the range-bounding fault along the west side of the range, indicating that it probably occurred on the reverse fault responsible for uplifting the range [*Stauder*, 1973; *Chinn and Isacks*, 1983; *Jordan et al.*, 1983; *Jordan and Allmendinger*, 1986]. Reprocessing of a YPF seismic line across the south end of Southern de Valle Fértil also suggests the range is uplifted on an east dipping reverse fault on its west side [*Snyder et al.*, 1986; *Snyder*, 1988], supporting a thick-skinned model for the Sierras Pampeanas where uplift occurs on major crustal-scale faults similar to that imaged in the Wind River COCORP line [*Brewer et al.*, 1980].

Another typical Pampean event is a 1977, Ms=7.4 earthquake beneath Pie de Palo. *Instituto Nacíonal de Prevención Sísmica* (INPRES) [1977] and *Volponi* [1979] proposed it was a double shock and *Kadinsky-Cade* [1985] showed that it was an Ms=6.8 foreshock and Ms=7.3 mainshock, both at 17 km depth (F and M, Figure 3). Unfortunately, the nodal planes of the foreshock, mainshock, and two aftershocks (A1 and A2, Figure 3) do not project into the surface geology [*Chinn and Isacks*, 1983; *Giardini et al.*, 1985; *Kadinsky-Cade*, 1985], and no surface rupture has been discovered to constrain the earthquake generating fault [*INPRES*, 1977]. Coseismic surface deformation [*Kadinsky-Cade et al.*, 1985; *Reilinger and Kadinsky-Cade*, 1985] and geologic observations [*Zambrano*, 1969; *Bastias and Weidmann*, 1983; *Fielding and Jordan*, 1988], however, suggest that Pie de Palo is uplifted on a west dipping reverse fault along its east side. The lack of correlation between the focal mechanisms and surface geology and the non-uniqueness of leveling inversions therefore preclude an unambiguous determination of the fault plane of the event.

Precordillera. In contrast to the Pampeanas, no focal mechanisms or accurate depths are known for shallow events in the Precordillera. Several large events are known to have occurred there, however, most notably a 1944 ($Ms \approx 7.4$-7.8) earthquake that destroyed the city of San Juan (Figure 3). In spite of a large number of young-looking fault scarps, none can be unequivocally related to the faults responsible for the historical earthquakes. The 1944 event produced 30 cm of coseismic thrust displacement on a small fault in valley fill and alluvium, but the small size of the trace and displacement suggest it is not the main fault, which was probably in the basement [*Castellanos*, 1944]. A 20-km-deep event with the typical foreland focal mechanism also occurred near the Precordillera in the southwestern Bermejo Valley. The depth implies that it is in the basement, and *Fielding and Jordan* [1988] use it to support the proposal that basement deformation beneath the Eastern Precordillera affects its structure, as suggested by *Baldis and Chebli* [1969].

TABLE 1. Crustal Velocity Models

Depth (km)	Velocity (km/s)
Three-layer Over Half-Space Model	
0.0	5.88
10.0	6.20
32.0	7.30
45.0	8.10
Single-Layer Over Half-Space Model	
0.0	6.4
45.0	8.10

The three-layer model was developed by *Bollinger and Langer* [1988] and is based on the the UNSJ velocity model [*Volponi*, 1968] for the bottom two layers. The velocity for the single-layer model is the average crustal velocity of the three-layer model.

Analysis of the Local Network Data

In this paper we study the local seismicity for two periods: (1) the year 1980 using INPRES and Universidad Nacionale de San Juan (UNSJ) data, and (2) the period from January 1984 to September 1985 using INPRES data (see Figure 4 for network map). For approximately half the data, the maximum number of stations reporting is four (typically CFA, RTLL, RTCV, and one of RTCB or RTMQ). We therefore required at least four P and two S arrivals among these five stations to select an event for location, producing a set of 2300 events. The events were located using HYPOINVERSE [*Klein*, 1978] and a measured V_P/V_S ratio of 1.75.

Hypocenters in this area have a strong bimodal distribution, 75% are at intermediate depth within the subducted plate, and the rest are at crustal depths in the overriding plate [*Smalley and Isacks*, 1987]. A pronounced aseismic region from 40 to 95 km depth separates the two zones. To select well-located shallow events, we applied several criteria, described below, producing a subset of 575 events. These events were located with two crustal velocity models, one with a single layer and one with three layers (Table 1). The three-layer model produces better locations in terms of residuals, estimated hypocentral errors, and stability, but we find that the depths of the less well-constrained hypocenters tend to "stick on" boundaries in the velocity model.

The shallow activity is shown in map and cross section in Figures 5 and 6. Two selections were made based on the location quality. The best locations, class A, have well-constrained hypocenters, while good locations, class B, have events with well-constrained epicenters but less reliable depths. The initial criteria, RMS residual between 0.05 and 0.5, estimated epicenter and depth errors of less than 5 and 10 km, respectively, and condition number less than 150, select 280 events. An additional requirement that the ratio of epicentral distance to hypocentral depth be less than 2 selects 190 events as class A events, while the remaining 90 events form class B. Histograms of the depth distribution as a function of class are shown in Figure 7. Note that the depth histogram for events beneath Pie de Palo obtained with the three-layer model has a distinct peak near 10 km for class B events that is absent for class A events (Figure 7*a*). The single-layer velocity model produces similar sets of class A and B events, but the depth histograms do not have peaks near 10 km (Figure 7*b*). The apparent peak in seismicity near 10 km for the three-layer class B events below Pie de Palo is thus probably an artifact of the location process due to events "sticking" on the 10-km boundary of the velocity model.

With respect to the depth distribution, we must consider the relative geometries of the seismicity and network, as most of the events are outside the network where depth resolution degrades. This degradation, however, is a function of the depth of the event. An event whose depth is greater than its epicentral distance typically has the depth determined as well as an event inside the network. In cross section (Figures 6*a* and 6*b*), we see that events beneath Pie de Palo with a distance depth ratio less than 1 are concentrated between 20 and 35 km depth, but there are relatively few of them. In addition, there is no systematic change in the maximum depth with distance. We therefore loosened our criteria to a distance depth ratio of 2, greatly increasing the number of events available for determining the depth distribution while maintaining reasonable depth control. These criteria, however, also affect the shape of the histogram. Applying them to a uniform volumetric distribution, for example, produces a histogram with a quadratic increase with depth. The shapes of the depth histograms indicate this effect is minor for depths greater than about 15 km (Figure 7). Preliminary results from a 40-station, digital network experiment that surrounded the crustal seismicity of both Pie de Palo and the Precordillera

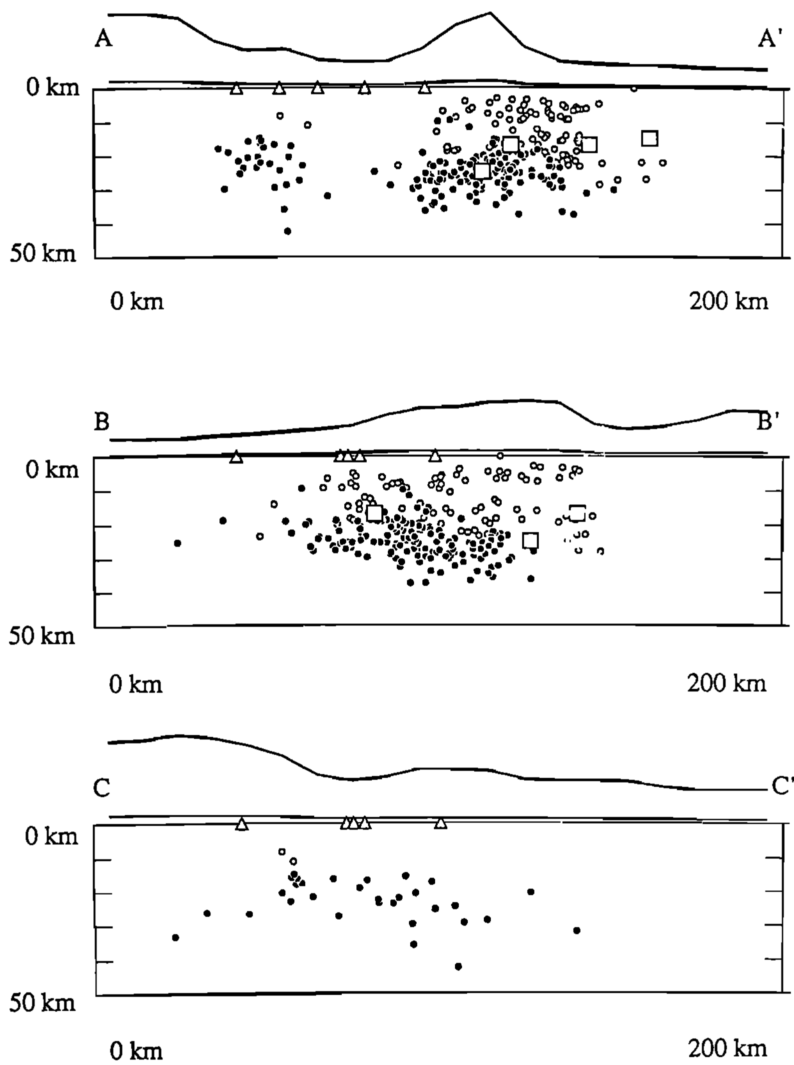

Fig. 6. Cross sections of shallow seismicity from Figure 5. (*a*) North-northeast view of events beneath both Pie de Palo and the Precordillera (A-A'), West-northwest views of events (*b*) beneath Pie de Palo (B-B') and (*c*) beneath the Precordillera (C-C'). Only class A and B events are shown in these sections, using the same symbols as in Figure 5. Projections of station locations are shown by open triangles. A 100-km-wide "swath average" of the topography, computed from digital topography [*Isacks*, 1988], is also projected at scales of 1:1 and 10:1. Projections of events with known focal mechanisms and depths from teleseismic studies shown in Figures 3 and 5 are shown by the open boxes in Figures 6*a* and 6*b*.

confirm both the map and depth distributions found in this study [*Regnier et al.*, 1988; *Smalley et al.*, 1988].

FORELAND SEISMICITY: LOCAL NETWORK DATA

Seismicity of Sierra Pie de Palo. The patterns of seismicity associated with Pie de Palo from local and teleseismic data are similar, although their "centers of mass" differ by about 20 km (Figures 3 and 5). This is likely due to a real difference in the locations of the events rather than systematic differences in location obtained from the two data sets [*Kadinsky-Cade*, 1985]. In general, the seismicity is concentrated in a small region beneath the eastern half of Pie de Palo and the valley to its east and does not define simple planar faults, and the majority of the activity does not correlate with surface faults. Cross sections of the network data (Figures 6*a* and 6*b*) illustrate the difficulty of associating the seismicity with specific faults or fault planes in this area.

Seismicity of the Precordillera. The local network data provide several new observations with respect to the seismicity of the Precordillera. First, the majority of events are located on its east side south of about 31.1°S, beneath the fault-bounded Matagusanos Valley, a pattern not at all evident from the sparse activity reported in the Preliminary Determination of Epicenters (PDE) and International Seismological Centre (ISC) (Figures 2 and 3). Second, the Western and Central Precordillera subprovinces are nearly aseismic. Third, the activity has a volumetric distribution (Figures 6*a* and 6*c*). Fourth, the hypocentral depths range from 10 to almost 40 km and are concentrated between 15 and 30 km (Figure 7*c*). This depth range is similar to that reported throughout the Andean foreland from teleseismic data, and beneath Pie de Palo from aftershock studies [*Kadinsky-Cade*, 1985] and our local network data.

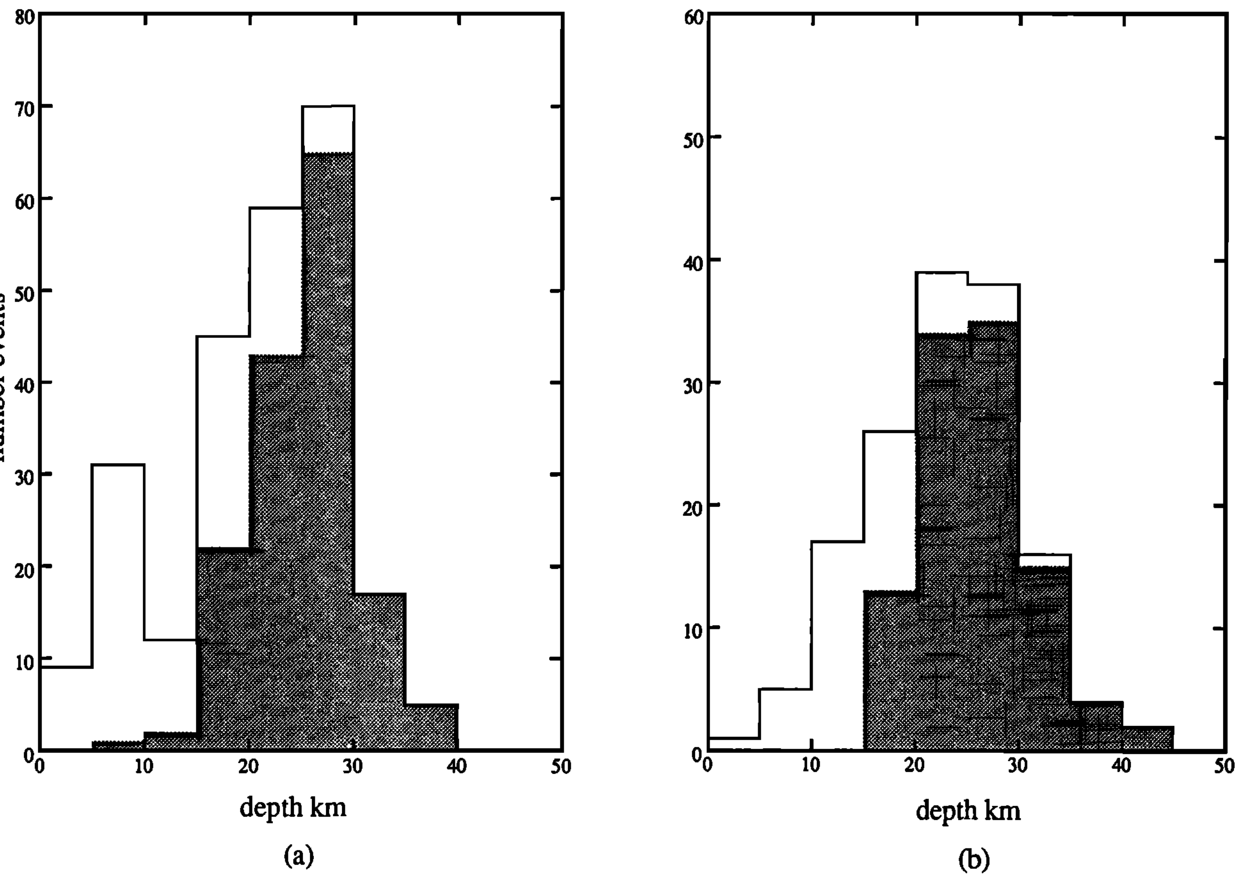

Fig. 7. Depth histograms for local network data. Class A data are shown by the gray bars and combined class A and B data by the solid lines. (*a*) Data from Pie de Palo region obtained using three-layer crustal velocity model. Hypocentral activity beneath Pie de Palo is concentrated between 20 and 30 km, with the maximum depth between 35 and 40 km. Note that the large peak between 5 and 10 km in the class B histogram is absent in the class A histogram. (*b*) Data from Pie de Palo region using a single-layer crustal velocity model. Note that a peak near 10 km is not present. (*c*) Data from Precordillera obtained using three-layer crustal velocity model. Hypocentral activity beneath the Precordillera is concentrated between 15 and 30 km with the maximum depth between 35 and 40 km.

The depth distribution provides important clues to understanding the anomalous structures of the Eastern Precordillera: the reversal of thrust vergence and the apparent downsection cut of the décollement. In the allocthonous sedimentary package of a thin-skinned thrust belt, the major thrusts verge and young in the direction of transport [*Dahlstrom*, 1970]. The Western and Central Precordillera follow this pattern, but the vergence reverses in the Eastern Precordillera. Such a change, while unusual, is observed along the leading edge of several thin-skinned fold-thrust belts, such as the Canadian Rockies of Alberta [e.g., *Price*, 1981]. Another characteristic of thin-skinned thrusting is that the stratigraphic level of the décollement cuts upsection in the transport direction. Coincident with the change in vergence, however, the décollement appears to cut more than 2.5 km downsection in the transport direction [*Baldis and Bordonaro*, 1984]. This presents a serious problem for an interpretation of the Eastern Precordillera as part of a standard thin-skinned thrust belt. Several models, including a triangle zone, an "out of the basin" thrusted anticline, and buried thick-skinned deformation have been proposed to explain the observations inconsistent with a thin-skinned deformation model [*Baldis and Chebli*, 1969; *Fielding and Jordan*, 1988].

The depth of the basal décollement in the Central Precordillera is estimated to be 3-4 km in the east and 5-8 km in the west [*Baldis and Chebli*, 1969], much less than the shallowest well-located earthquakes, and about 8 km in the Eastern Precordillera and Matagusanos Valley [*Fielding and Jordan*, 1988], close to the depth where the seismicity starts. The seismicity recorded by the local network is therefore constrained to the basement, beneath the décollement, and provides direct evidence that basement deformation is occurring there. The geologic observations inconsistent with traditional thin-skinned deformation can therefore be interpreted as a result of this basement deformation. The seismicity could represent (1) deformation in Pampean basement underthrust beneath the Eastern and Central Precordillera, similar to the basement underthrusting occurring beneath the Subandean Belt, (2) deformation in the unknown basement beneath the Precordillera, or (3) reactivation of a suture between the basement beneath the Precordillera and the Pampean basement. In each model, the melange along the east side of the Eastern Precordillera could represent the surface exposure of the suture, although the melange is probably displaced from the boundary at depth.

Transverse lithospheric structure. The seismicity of Pie de Palo terminates in a sharply defined, E-W striking linear

(c)

Fig. 7. (continued)

boundary near its north end at about 31.1°S (Figures 3 and 5), coincident with a major offset in the depth to basement along the southern margin of the Bermejo Valley. Geologic and geophysical data indicate that basement reaches depths of up to 13 km in the southern Bermejo Valley [*Jordan and Allmendinger*, 1986], while it reaches 3 km elevation on Pie de Palo, for a total structural relief of more than 16 km both across and along the strike of the N-S structures of the Sierras Pampeanas. In the valleys on either side of Pie de Palo, the basement is shallow, typically 1-2 km [*Gray de Cerdan*, 1969; *Snyder*, 1988].

Simple models of Laramide-style block faulting are typically two-dimensional, with the deformation creating a series of long, linear mountain blocks oriented perpendicular to the shortening direction with no along-strike structure developed. This approximation is valid on the large scale but is an over simplification when applied to individual blocks. The sharpness of the cross-strike discontinuities in the basement topography and seismic activity here suggests that the basement blocks of the Pampeanas may be broken along-strike by major fault systems.

The north end of Pie de Palo is bounded by E-W striking, steeply dipping faults with the north sides downthrown, several of which show evidence of Quaternary activity [*Zambrano*, 1969; *Bastias and Weidmann*, 1983; *Fielding and Jordan*, 1988]. As Pie de Palo is thrust eastward, these faults may accommodate the movement through oblique slip [*Fielding and Jordan*, 1988]. Inferred and mapped continuations of these faults are found eastward towards Southern de Valle Fértil and westward into the Eastern Precordillera [*Zambrano*, 1969]. The along-strike change in the structure of the Eastern Precordillera, from thrust sheets in the south to folds in the north, and the bifurcation of the south end of Southern de Morado (Figure 4) occur where these faults enter the Eastern Precordillera. Westward projection of the cross-strike structure may also include several unusual closed basins within the Central Precordillera and a major structural high in the Calingasta Valley west of the Precordillera. To the east, the strike of Southern de Valle Fértil exhibits a change where it intersects the cross-strike feature. The seismicity may therefore indicate a structure with dimensions of lithospheric, rather than upper crustal only, scale.

Almost directly beneath the shallow upper plate, seismicity is an active nest of intermediate-depth seismicity (Figures 8 and 9) in a 12-km-thick zone centered at 106 km depth [*Smalley and Isacks*, 1987]. This nest lies in an E-W trending concentration of Wadati-Benioff zone seismicity coinciding with an along-strike continuation of the Juan Fernandez ridge of the oceanic Nazca plate [*Smalley and Isacks*, 1987] and may represent the continuation of the ridge in the subducted plate. The nest may be due to a concentration of seismicity in a structure such as a seamount, where a change in the strength of the subducted plate may occur or friction between the upper and lower plates may be greater. Although resolution is poor and identification of specific features speculative, it is reasonable to expect the seismicity of a subducted plate to be affected or possibly controlled by major structures, such as aseismic ridges or seamount chains, in the subducting plate. In map view, the northern boundary of the intermediate depth nest also defines a remarkably sharp east-west striking line at about 31.1°S coincident with and parallel to the northern boundary of the shallow seismicity and the transverse feature of the upper plate basement described above. This coincidence raises the possibility that the upper plate seismicity and basement structure may be related to the interaction of structures in the lower plate with the upper plate.

Earthquake Depths and Crustal Rheology

The depth distribution of the crustal activity in San Juan is unusual in terms of continental seismicity. Local network studies in California [*Marks and Lindh*, 1978; *Wong et al.*, 1977; *Wong and Savage*, 1983], the Colorado plateau [*Wong et al.*, 1983], Switzerland [*Deichmann*, 1987], and the Shillong Plateau in India [*Kharshiing et al.*, 1986; *Kayal*, 1987] have found sparse seismicity to depths of 30-40 km, although the activity is concentrated at depths shallower than 20-25 km. *Chen and Molnar* [1983] found that the majority of shallow seismicity not associated with recent subduction is shallower than 20 km, although they found isolated, anomalous events at depths of 30-100 km. Another area where lower crustal earthquakes are found is the East African rift system. The seismicity there is concentrated between 5 and 20 km depth, but events are found as deep as 30 km, implying that thermal weakening of the crust is localized to the immediate region of the rift [*Shudofsky*, 1985; *Shudofsky et al.*, 1987]. In the Zagros thin-skinned fold and thrust belt, seismicity is observed in the basement below the basal décollement of a thin-skinned thrust belt, but it is limited to depths of 6-13 km [*Ni and Barazangi*, 1986]. While these studies show that mid to lower crustal seismicity is not unique to the San Juan area, we know of no other regions of intracontinental seismic activity where

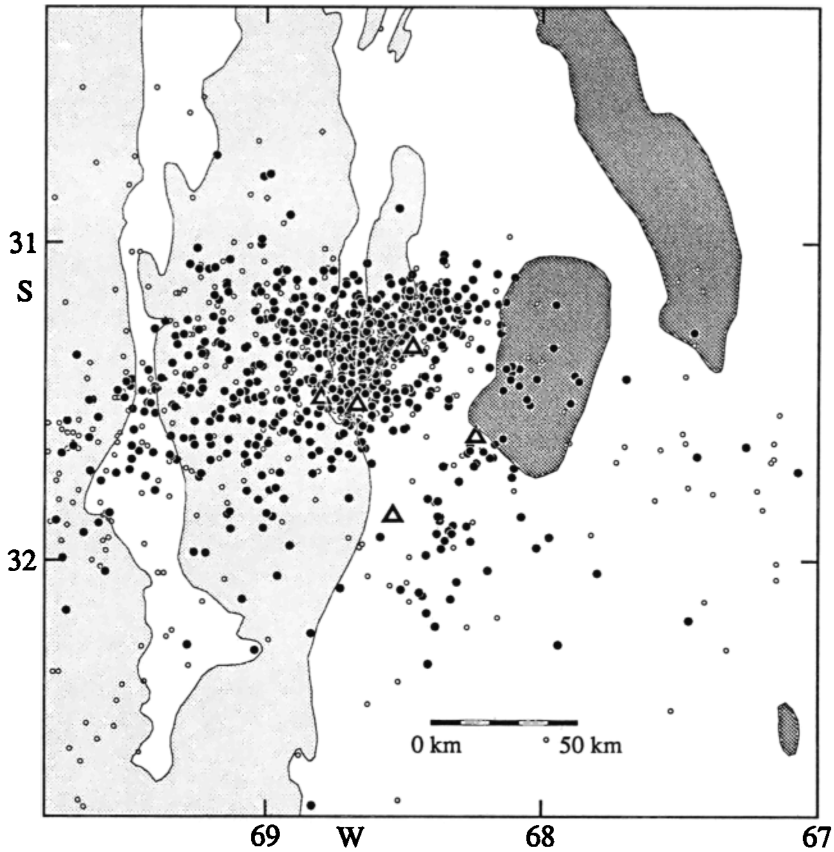

Fig. 8. Map of approximately 800 intermediate depth events recorded by the local networks illustrating the very active nest of intermediate depth seismicity almost directly below the network. Selection criteria and symbols are the same as in Figure 5.

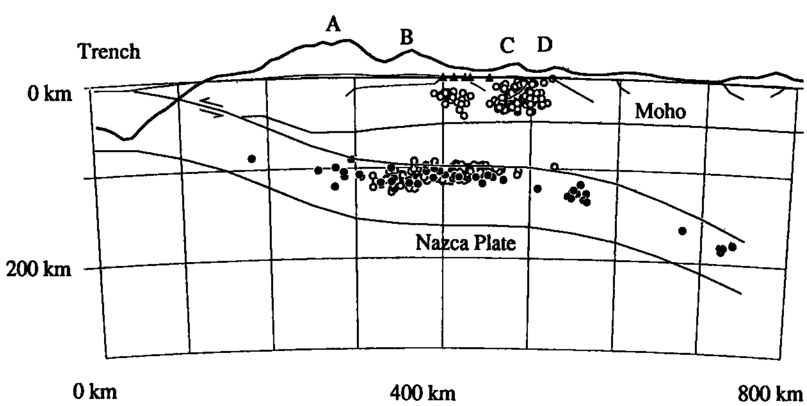

Fig. 9. East-west trending cross section at 31.25°S showing projections of the class A and B crustal (A, solid circles; B, open circles) and intermediate-depth events (both classes, open circles), major inferred crustal-scale faults, inferred location of Moho, and a 100-km-wide swath average of topography at 1:1 and 10:1. The crustal faults and the inferred position of the Moho are taken from work by *Isacks* [1988], *Kadinsky-Cade* [1985], and *Ortiz and Zambrano* [1981]. The widths of the cross sections are 70 km for the local network data and 160 km for the ISC data (solid circles, intermediate depth only). Projections of the INPRES network stations are shown by triangles.

earthquake depths are concentrated at depths between 20 and 30 km beneath a relatively aseismic upper crust.

We propose that the unusually deep depth distribution of crustal earthquakes in San Juan is a consequence of the late Cenozoic development of the subhorizontal subduction geometry there. The Wadati-Benioff zone, at about 105 km, depth limits the maximum thickness of the South American lithosphere to approximately 90 km, assuming that the Wadati-Benioff zone events are about 15 km beneath the top of the subducted slab [*Smalley and Isacks*, 1987]. While the thickness of the lithosphere before the development of the present subduction geometry is unknown, the peneplain-like basement surfaces exposed throughout the 400 km width of the Sierras Pampeanas and the absence of significant metamorphism or deformation in the sediments covering the province suggest that it was a

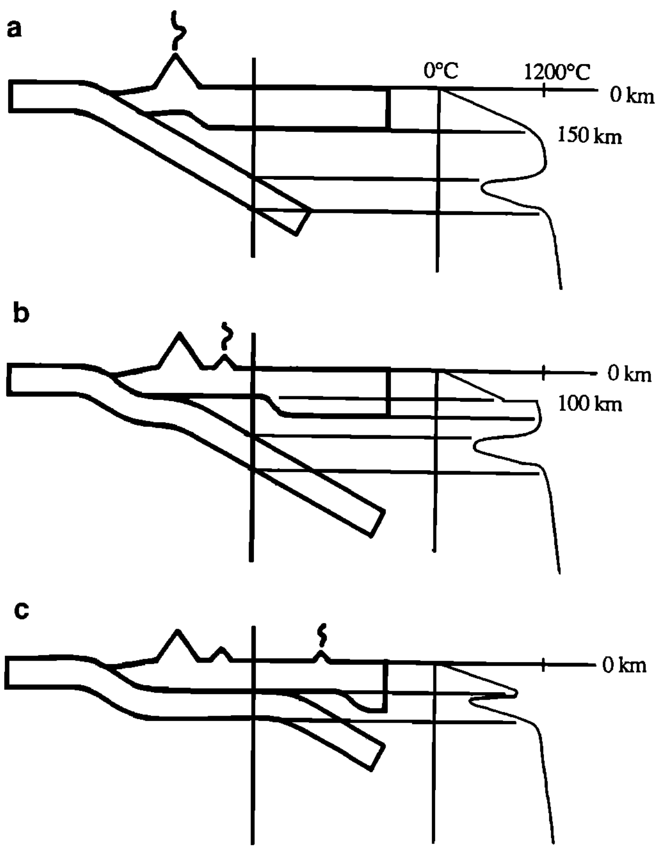

Fig. 10. Schematic cross sections at 31°S illustrating model of Neogene evolution of the South American crust and lithosphere, the volcanic arc and the Nazca plate subduction geometry at approximately (*a*) 20 Ma, (*b*) 10 Ma, (*c*) the present. Geotherms representing model of thermal evolution of the South American crust and lithosphere at each time period are shown on the right. Based on work of *Isacks* [1988], *Kay et al.* [1987], *Smalley and Isacks* [1987], and *Cahill and Isacks* [1985]. Initial geotherms for continental plate 8°C/km, oceanic plate 12°C/km and, the asthenosphere 0.3°C/km. See text for discussion.

low-lying, stable cratonic region during a large portion of the Mesozoic and Cenozoic. It may therefore have had a lithospheric thickness and geotherm typical of continental shields. If this were so, a large amount of thinning took place during the development of the flat subduction geometry, produced by processes in the asthenospheric wedge between the upper plate and the inclined subducted plate [*Isacks*, 1988].

Thinning the lithosphere typically produces high plateaus, such as the Colorado Plateau, due to thermal effects [*Thompson and Zoback*, 1979; *Keller et al.*, 1979]. The Sierras Pampeanas are relatively low-lying, however, with an average elevation less than 1 km. Thinning the lithosphere, promptly followed by emplacement of the cool, horizontally subducting slab, can produce a thin upper plate without changing its geotherm if the process occurs rapidly enough that significant heat transfer into the upper plate cannot occur. This is schematically illustrated in Figure 10. The double lithospheric thickness and the absence of thermal expansion prevent the uplift normally associated with thin lithosphere [*Smalley and Isacks*, 1987].

Examination of the strength of various crustal rock types using the models of *Meissner and Strehlau* [1982] and *Sibson* [1982] for a range of typical heat flows and strain rates constrains the conditions necessary to produce the observed earthquake depth distribution (Figure 11). These models, based on a moderate strain rate of 10^{-16-1} from estimates of total shortening across the Sierras Pampeanas [*Jordan and Allmendinger*, 1986], a heat flow near the average value for continental crust, and a mafic lower crustal rheology produce a crust and upper mantle with a maximum strength near 35 km.

The crust of the Sierras Pampeanas is therefore failing under compression because a substantial fraction of the mantle part of the continental lithosphere, which in stable continental areas mainly supports the lithospheric stresses, is here largely missing. The nearly horizontal subducted plate does not provide this strength because it is decoupled from the upper plate by an intervening zone of low shear

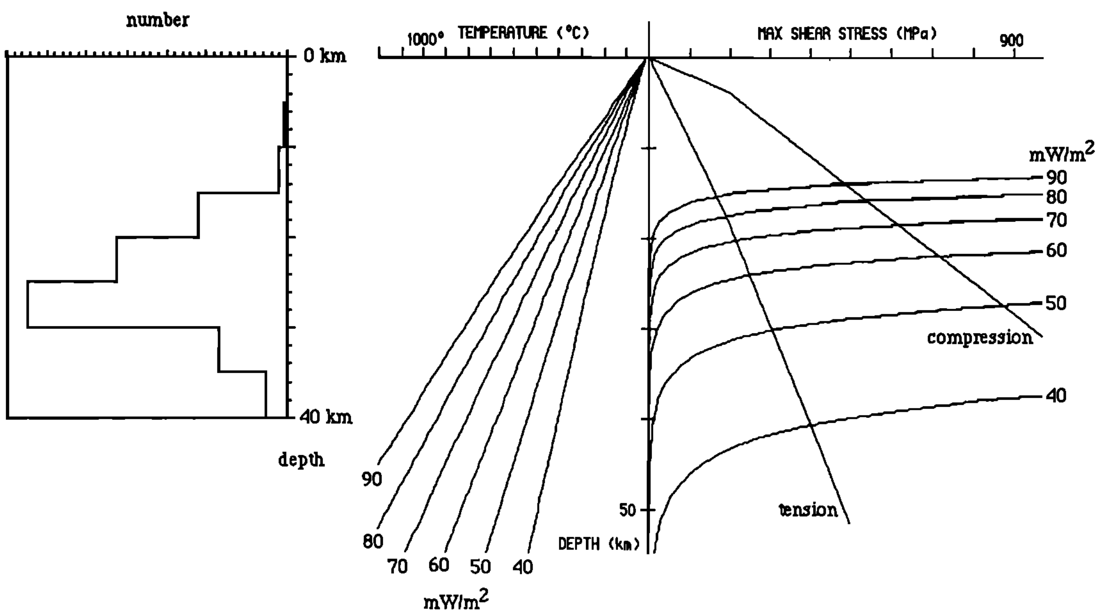

Fig. 11. Crustal rheology model based on models of *Meissner and Strehlau* [1982] and *Sibson* [1982]. Depth histogram from Figure 7*a*, set of geotherms for surface heat flows of 40-90 mW/M^2, and brittle and shear resistance versus depth for diabase at a strain rate of 10^{-16-1} illustrate how the combination of cool crustal geotherm and relatively mafic crustal composition can produce a maximum crustal strength at depths of 30-35 km.

strength. The stresses in the foreland crust are therefore increased relative to those in the South American craton because of the thinning of the foreland lithosphere and the consequent concentration of the horizontal lithospheric stresses. The stresses would be approximately increased by the ratio of the cratonic lithospheric thickness to the Pampeanas lithospheric thickness.

CONCLUSIONS

New data from local networks in a seismically active region of the Andean foreland shows mid to lower crustal basement seismic activity associated with the basement uplifts of the Sierras Pampeanas and the eastern portion of the thin-skinned Precordillera fold-thrust belt. In both regions there is little seismicity shallower than 10 km and the maximum depth is near 40 km. The seismicity associated with Pie de Palo, concentrated between 15 and 30 km depth, supports the interpretation that the Sierras Pampeanas are uplifted on faults that extend through much of the crust. Although few events were located in the Western and Central Precordilleran, significant activity at 10-40 km depth was detected beneath the Matagusanos Valley between the Central and Eastern Precordillera. This seismicity is also concentrated between 20 and 30 km depth. While more subsurface information, focal mechanisms, and other data are needed to fully constrain models for the Precordillera, the seismicity indicates that basement deformation plays an important and possibly dominant role in determining the structure of the Eastern Precordillera. The unusually deep concentration of the crustal seismicity in both areas of the foreland can be understood as a consequence of transient thermal effects of the late Cenozoic change in dip of the subducted plate.

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