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SEISMICITY, FAULT PLANE SOLUTIONS, DEPTH OF FAULTING, AND ACTIVE TECTONICS OF THE ANDES OF PERU, ECUADOR, AND SOUTHERN COLOMBIA

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Abstract. The intracontinental seismicity of the Andes of Peru, Ecuador, and southern Colombia is concentrated along the easternmost flank of the Cordillera beneath the western margin of the sub-Andes. The focal depths and fault plane solutions of the largest events were constrained by comparing the observed long-period P waves with synthetic waveforms. In general, the fault plane solutions show reverse faulting on steeply dipping nodal planes striking northwest-southeast and reflect crustal shortening perpendicular to the range, probably in response to the subduction of the Nazca plate to the west. Earthquakes in the sub-Andes occur at depths of between 8 and 38 km, indicating that much of the crust deforms in a brittle manner. The seismicity seems to reflect antithetic underthrusting of the Brazilian shield beneath the eastern margin of the Andes. However, the earthquakes are too deep and their nodal planes are too steep to be associated with the thin-skinned decollement of the sedimentary cover. Instead, they appear to reflect the tectonic deformation of the underthrust basement west of the zone of active deformation at the surface in the sub-Andes. Hence the earthquakes on the east side of the Andes may reflect the style of deformation in the hinterland of fold and thrust belts, such as in the Canadian Rockies, while much of the decollement farther east occurs aseismically. In contrast, normal faulting occurs in the high Andes on planes parallel to the strike of the mountain belt. Thus, there is a delicate balance between the compressive stress applied to the Andes in the direction of subduction, which causes thrust faulting in the sub-Andes, and the gravitational body forces acting on the topographically higher parts and the crustal root of the Andes, which may cause normal faulting at high elevation.

Introduction

The Andes are important as a contemporary example of a mountain belt formed as a result of subduction of oceanic lithosphere beneath a continental plate. They are frequently used as the type example of a tectonic environment in which many ancient orogenic belts formed [e.g., Dewey and Bird, 1970]. "Andean margins" are characterized by a voluminous magmatic arc bounded on one side by a trench and on the other side by a fold and thrust belt. They are inferred to have been present along some continental margins during the closing of oceanic terranes that ultimately led to continental collisons.

Western North America is interpreted to have been an "Andean margin" during the late Mesozoic and early Tertiary, when the Farallon plate or other oceanic plates were subducted beneath it [Burchfiel and Davis, 1972, 1975; Hamilton, 1969], much as the Nazca plate presently plunges beneath western South America. The variability of structural development along strike in both North and South America, however, allows the generic term "Andean margin" to embrace a broad spectrum of tectonic styles. Specifically, differences between the tectonic development of western Canada, with its classic thin-skinned fold and thrust belt [Bally et al., 1966; Price and Mountjoy, 1970], and the broad zone of Late Cretaceous-early Tertiary Laramide deformation, with faults extending through the crystalline crust of the western United States in Colorado and Wyoming [e.g., Burchfiel and Davis, 1972, 1975; Sales, 1968; Stearns, 1978], are as great as the differences, for instance, in the tectonic style of the Alps, a collisional orogen, and the High Atlas, an intracontinental zone of deformation. Yet the Late Cretaceous and early Tertiary structures in both Canada and the United States formed in a setting typically classified as an "Andean margin". Moreover, the structure and history of the Andes shows similar and equally large variations in tectonic style along strike. Even though knowledge of Andean geology is still limited, it is clear that the Pampean ranges in central Argentina represent a broad zone of deformation characterized by faults that extend into crystalline basement. They are clearly different from the wide belt of folds and thrust faults that appear to affect only the Paleozoic sedimentary cover in Bolivia in what can be interpreted as a thin-skinned tectonic style [Jordan et al., 1983]. Accordingly, before the concept of an "Andean margin" can contribute more to the understanding of older belts than simply to suggest cartoonlike analogies, a more comprehensive understanding of the tectonics of different parts of the Andes is required.

The purpose of this paper is to contribute some new data bearing on the active tectonics of the Andes of Peru, Ecuador, and southern Colombia and to summarize these data in light of plausible mechanisms of how the Andes might have formed and evolved. We are concerned primarily with evidence of recent faulting, particularly reflected by intracontinental earthquakes and geologic evidence of Quaternary faulting.

In the Central Andes, local seismic station coverage is scanty, and routine determinations of hypocentral depths are frequently grossly in error [e.g., James et al., 1969]. Using

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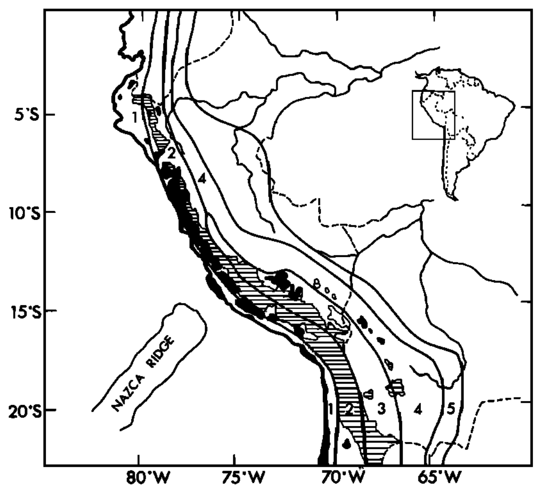


Fig. 1. Physiographic provinces of the Andes of Peru and northern Bolivia: (1) coastal plains, (2) Cordillera Occidental, (3) Altiplano, (4) Cordillera Oriental, (5) sub-Andean zone. Shaded areas indicate plutonic bodies and hatchured areas the volcanic rock cover [after Dalmayrac et al., 1980].

synthetic seismograms of long-period body waves, we revise published fault plane solutions [Pennington, 1981; Stauder, 1975; Wagner, 1972] and determine both focal depths and fault plane solutions for others. At epicentral distances of between 30° and 80°, the shape of long-period body waves depends upon both the depth of the hypocenter and the orientation of the nodal planes and is not very sensitive to changes in the local structure of the upper mantle. Thus focal depths and fault plane solutions can be constrained by a visual comparison between observed long-period P waves and synthetic waveforms on a trial-and-error basis.

The Peruvian Andes: An Overview

The main morphological units in the Peruvian Andes are, from the trench eastward, the coastal plains, the Cordillera Occidental, the Altiplano, the Cordillera Oriental, and the sub-Andes (Figure 1). A brief description of the geology and tectonic evolution of each region is taken from recent reviews of central Andean geology [Audebaud et al., 1973; Dalmayrac et al., 1980; Gansser, 1973; Megard, 1978; Zeil, 1979].

The Coastal Plains

In northern Peru, the coastal plain is a narrow, arid strip of land not wider than 40 km. Limited on the west by the coastline and to the east by the Cordilleran batholith, it consists mainly of gently folded volcanic and sedimentary rocks of Mesozoic age. In southern Peru, the coastal plains widen, and strongly folded crystalline basement rocks crop out. The Precambrian coastal massif in southern Peru, generally called the Arequipa block, contains rocks as old as 1.8-2.0 b.y. [Cobbing et al., 1977; Dalmayrac et al., 1977] and was subjected to deformation in Precambrian, Paleozoic, and Mesozoic times [Shackleton et al., 1979].

The Cordillera Occidental

The western cordillera consists mainly of volcanic and plutonic rocks of Mesozoic and Cenozoic age and shallow water marine deposits of Mesozoic age. It forms a continuous and impressive structural entity parallel to the coast and is the locus of most of the late Cenozoic volcanism. The immense Coastal Batholith extends from about 6° to 16° S parallel to the coast (Figure 2). Plutons ranging in composition from gabbro to syenogranite [Cobbing and Pitcher, 1972; Cobbing et al., 1977] were emplaced episodically from 100 to 30 m.y. ago [Pitcher, 1975]. The plutons become younger and show an increase in silicic content eastward [Bussel et al., 1976; Stewart et al., 1974]. Cenozoic volcanic rocks are widespread in the western cordillera, and massive ignimbrites of Neogene age are present [Dalmayrac et al., 1980]. In central Peru, volcanic activity began to wane about 11 m.y. ago and stopped rather abruptly 5 m.y. ago [Noble and McKee, 1977]. This cessation of Quaternary volcanism has been interpreted as marking the beginning of shallow plate subduction under Peru [Barazangi and Isacks, 1976, 1979; Megard and Philip, 1976].

East and north of the Cordillera Blanca, structures developed in the Mesozoic rocks are similar to those of a foreland fold and thrust belt. The folds and thrusts in this region are generally of flexural slip type with both broad-rounded hinges or chevron geometry. The Mesozoic limestones were thrust eastward on gentle west dipping faults, while the underlying Jurassic shales formed tight flexural slip folds [Dalmayrac, 1978; Dalmayrac et al., 1980; Wilson et al., 1967]. The tectonic style implies detachment of the Mesozoic cover from an older substratum and extensive east-west shortening [Coney, 1971]. In the western part of the belt, volcanic rocks of Oligocene-Miocene age unconformably overlie deformed Mesozoic rocks and limit the age of deformation to Late Cretaceous and early Cenozoic time [Dalmayrac, 1978; Wilson

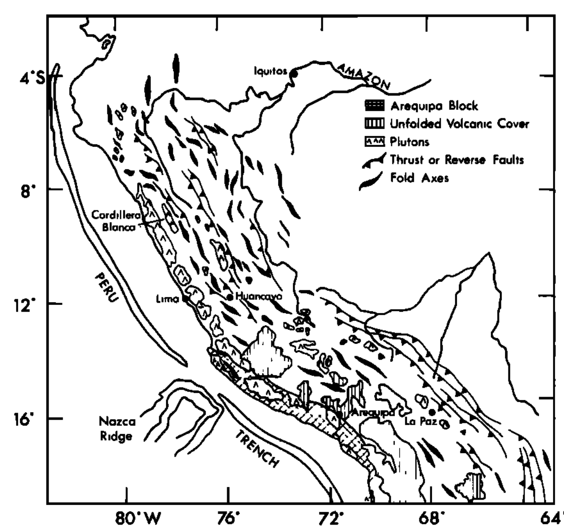


Fig. 2. Main structural features of the Peruvian Andes [after Audebaud et al., 1973].

et al.; 1967]. The folded belt is truncated and limited on its western side by the granitic plutons of the Cordillera Blanca emplaced 3-12 m.y. ago [Stewart et al., 1974].

In central Peru, the Mesozoic sedimentary rocks east of the Coastal Batholith are tightly folded in chevron folds. Fracture cleavage present in the Late Cretaceous red beds is absent in the overlying Tertiary volcanic rocks, indicating a Late Cretaceous to early Tertiary age of deformation [Megard, 1978].

In southern Peru, the Mesozoic sedimentary rocks are almost completely covered by Cenozoic volcanic rocks and the amount and age of deformation of the Mesozoic rocks are not well known. Recently, Vicente et al. [1979] mapped a subhorizontal east directed thrust fault with a minimum displacement of 15 km in the area near Arequipa (Figure 2). The thrust fault places rocks as old as Precambrian on Mesozoic sedimentary rocks. Although this thrust fault is now west of the active volcanoes, when thrusting occurred during the Late Cretaceous, it lay east of the volcanic arc active then.

The Altiplano and Central High Plateau

The Cordillera Oriental, to the east, and the Cordillera Occidental, to the west, bound the Altiplano and the high plateau of central Peru. North of about 10°S, the central high plateaus disappear and the western and eastern cordillera are adjacent. The width of the high plateau in central Peru is 10-50 km wide and increases considerably in the south to nearly 200 km near Lake Titicaca in the Altiplano. The thick sequences of Paleozoic and Mesozoic marine sedimentary rocks that are present in the high plateau and the Altiplano apparently were deposited in deep, northwest trending basins [e.g. Harrington, 1962; Lohman, 1970; Wilson, 1963] and later were capped by Late Cretaceous to late Eocene red beds.

The tectonic style in the central high plateaus is variable along strike. In the region around Huancayo in central Peru (Figure 2), zones of tight folds and thrust faults form narrow belts (10-30 km wide) that are separated by zones of open folds and undeformed rocks of similar width [Lepry and Davis, 1983; Megard, 1978]. These belts of intense deformation anastomose along strike. The geometry of folds and some faults suggests local detachment from their substratum [Lepry and Davis, 1983]. Where exposed, Paleozoic or Precambrian rocks appear to be involved in some of the structures.

Much of the Altiplano in southern Peru is covered by a thick sequence of mildly deformed continental molasse deposited during the Oligocene and Miocene [e.g., Newell, 1949]. Folding is less severe than in central and northern Peru, and the Tertiary sedimentary rocks are generally deformed in broad concentric folds cut by reverse faults [e.g., Chanove et al., 1969]. Northwest of Lake Titicaca, the thrust faults suggest that the sedimentary cover was detached from its underlying basement and deformed during Late Cretaceous and early Cenozoic time [Chanove et al., 1969; Laubacher, 1978].

Cordillera Oriental

The eastern cordillera is a broad zone underlain largely by pre-Mesozoic rocks located east of the central high plateaus. Precambrian crystalline rock and Paleozoic plutonic rocks crop out in large areas of the eastern cordillera, particularly in central Peru. The sedimentary rocks are composed of a thick section of mainly Paleozoic shallow marine and continental strata that are folded and frequently cut by reverse faults.

Precambrian rocks are generally low-grade metasedimentary rocks (phyllites or fine-grained schists) that are weakly to strongly foliated [Dalmayrac et al., 1980; Megard, 1978]. Only locally do the Precambrian crystalline rocks reach amphibolite to granulite grade. Lower Paleozoic rocks range from Ordovician to Devonian in age and are generally unmetamorphosed to weakly metamorphosed. The upper Precambrian and lower Paleozoic rocks are the oldest rocks exposed in the eastern Andes, and they form a very anisotropic and weak upper crust. Published material [e.g., Dalmayrac et al., 1980; Megard, 1978] attributes deformation of these rocks to a middle to late Paleozoic ("Hercynian") event.

The Sub-Andes

The sub-Andean zone consists of a belt of folded sedimentary rocks parallel to the mountain chain between the high Andes and the Brazilian shield. The limit between the Cordillera Oriental and the sub-Andes is generally shown to be formed by a zone of west dipping thrust or reverse faults [Ham and Herrera, 1963]. The rocks consist of shallow water and continental sedimentary rocks deposited intermittently from Paleozoic to Pliocene time. No evidence of major Andean magmatism has been found in the sub-Andes. The thickness of the sedimentary cover is not well known, but in northern Peru, where exploration for oil has been active, depths of up to 10 km have been reported in some basins [Audebaud et al., 1973; Rodríguez and Chalco, 1975].

The upper Tertiary sedimentary rocks are poorly dated and the deformation of the sub-Andes is difficult to date precisely [e.g., Audebaud et al., 1973]. In Peru, the sub-Andes are formed by a series of folds and faults active from at least Pliocene time to the present. Deformation is usually interpreted as the result of crustal shortening expressed by cylindrical folds cut by steep, west dipping reverse faults that pass from the sedimentary rocks into the underlying basement [e.g., Audebaud et al., 1973; Megard, 1978]. The age and intensity of tectonic deformation is typically shown to decrease steadily to the east [e.g., Dalmayrac et al., 1980]. Published cross sections are mostly schematic because data from drill holes or seismic profiles have not generally been available. Dense vegetation limits the exposure; the outcrops are small, and fault planes commonly are not exposed well enough to determine their dips. In northern Peru, oil exploration has produced a more detailed three-dimensional picture of the structure of the sub-Andes [Rodríguez and Chalco, 1975; Touzett, 1975].

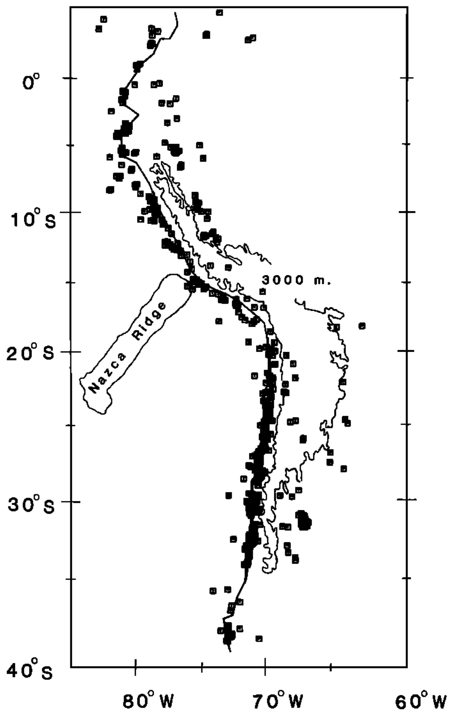


Fig. 3. Shallow focus seismicity in western South America from 1962 to 1979. Squares show the epicenters of earthquakes shallower than 70 km and registered by more than 40 stations reporting P waves to the International Seismological Centre catalog. Shown also is the 3000-m elevation contour.

There the sub-Andes can be interpreted as a thin-skinned fold-and-thrust belt.

Shallow Seismicity in South America

Epicenters of well-located shallow earthquakes (less than 50 km deep) occurring along the western margin of South America in the last 20 years or so define two distinct loci of seismic activity [Barazangi and Isacks, 1976, 1979]. To the west, one belt lies parallel to the coastline and marks the boundary where the Nazca plate subducts under South America [Chinn, 1982; Stauder, 1975]. The other belt of seismic activity occurs within the overriding continental plate and follows a trend parallel to the mountain chain. The majority of these crustal events take place in the transition zone between the eastern cordillera and the western margin of the sub-Andes. The distribution of epicenters of crustal events shows a remarkable quiescence of seismic activity in the high Andes and in the eastern part of the sub-Andes (Figure 3). There is also a paucity of seismic activity in the forearc region, a feature commonly observed in other areas of the world [e.g., Yamashina et al., 1978].

The spatial distribution of intracontinental seismicity shows variations along strike as well [Jordan et al., 1983]. The most seismically active areas occur under Peru and southern Ecuador between 2° and 13°S and under northern Argentina between 28° and 33°S (Figure 3). These regions correspond roughly to the two areas where teleseismic data suggest that subduction of the Nazca plate under South America takes place almost subhorizontally [Barazangi and Isacks, 1976, 1979; Hasegawa and Sacks, 1981; Megard and Philip, 1976]. In the Central Andes most of the intracontinental earthquakes occur in the eastern part of the cordillera, beneath the western sub-Andes (Figure 4). Seismicity along the eastern margin of the Altiplano and the Argentinian Puna, beneath which subduction of the Nazca plate occurs at a steeper angle (~30°), also shows the existence of a seismic sub-Andean margin, though less active than the two regions mentioned above (Figure 3). A remarkably quiescent segment of the sub-Andes is found along the eastern margin of the Altiplano of southern Peru and northern Bolivia between 13° and 18°S (Figure 4) [Jordan et al., 1983].

Fault Plane Solutions and Depth of Foci

Data and Method of Analysis

We studied all intracontinental events in the central Andes sufficiently large that a fault plane solution could be obtained with data from the World-Wide Standardized Seismograph Network (WWSSN) between 1962 and 1978 (Table 1). Constraints on the fault plane solutions and the depths of foci were obtained synthesizing long-period P waves for a spectrum of solutions and focal depths and comparing them to the observed waveforms, as others have done elsewhere [e.g., Jackson and Fitch, 1981; Kanamori and Stewart, 1976; Rial, 1978; Tréhu et al., 1981].

In this study, a point source embedded in semi-infinite half space was used to synthetize all available long-period P waves at WWSSN stations within an epicentral distance of 30-80°. Body waves at these distance ranges have paths that lie almost entirely within the lower mantle and are not affected by the local structure of the upper mantle [e.g., Langston and Helmberger, 1975]. At epicentral distances between 30° and 80°, P waves do not experience triplication or

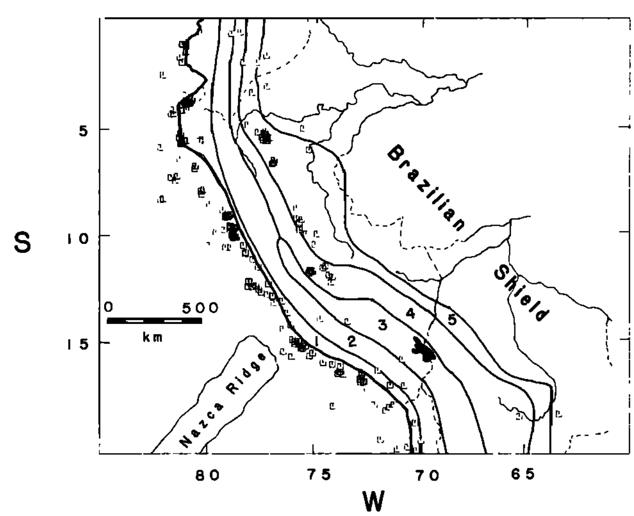


Fig. 4. Earthquakes shown in Figure 3 plotted on a physiographic map of the Central Andes. Intracontinental seismicity concentrated along the western margin of the sub-Andes in central and northern Peru.

TABLE 1. List of Earthquakes Studied: Data from the ISC Catalog

Event	Date	Latitude	Longitude	Origin Time	Depth, km	Magnitude
1	May 10, 1963	-2.20	-77.60	2222:42.0	33	6.7
2	Nov. 3, 1963	-3.50	-77.80	0310:12.7	33	6 7
3	Feb. 9, 1967	2.93	-74.83	1524:45.3	36	6.3
4	Jun. 19, 1968	-5.55	-77.20	0813:35.6	33	6.1
5	Jun. 20, 1968	-5.51	-77.30	0238:38.7	33	5.8
6	Dec. 1, 1968	-10.54	-74.81	1314:55.0	33	5.4
7	July 24, 1969	-11.84	-75.10	0259:20.9	1	5.9
8	Oct. 1, 1969	-11.75	-75.15	0505:50.0	43	5.8
9	Feb. 14, 1970	-9.84	-75.55	1117:16.4	36	5.8
10	Dec. 10, 1970	-3.97	-80.66	0434:38.0	15	6.3
11	Oct. 5, 1971	-14.20	-73.45	1033:46.3	54	5.7
12	March 20, 1972	-6.79	-76.76	0733:48.7	52	6.1
13	Feb. 23, 1973	-2.16	-78.33	0426:21.1	44	5.7
14	Sept. 27, 1974	2.72	-71.37	0409:01.6	44	5.5
15	April 9, 1976	0.85	-79.63	0708:50.0	19	6.0
16	May. 15, 1976	-11.62	-74.45	2155:55.0	5	5.9
17	Oct. 6, 1976	-0.76	-78.75	0912:39.0	33	5.7

diffraction near the core-mantle boundary that severely complicate the observed waveforms.

Synthetic seismograms were computed in the time domain by convolving the response of a shear dislocation in a semi-infinite half space with a far-field source time function, an attenuation operator, and the instrument response. The far-field source time function was assumed to be a symmetric trapezoid for both direct and reflected phases and was adjusted in an ad hoc manner to fit the observed P waves. The crustal structure at the receiver was assumed to be a single layer with an average velocity of 6 km/s. In none of the events studied did the crustal transfer function beneath individual stations introduce complications in the long-period waveforms. Plots of the fault plane

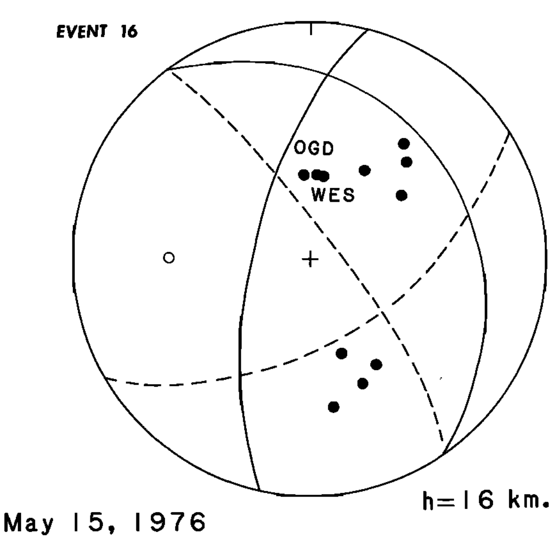


Fig. 5. Lower hemisphere equal-area projection of the focal sphere for the May 15, 1976, earthquake (event 16). Open circles indicate dilatational first motions, and solid circles indicate compressions. Two alternative pairs of nodal planes satisfying the data are shown.

solutions and examples of waveforms used in the analysis are shown in the appendix. Synthesized waveforms are also shown at several hypocentral depths for a few selected stations spanning a range of azimuths. This will permit the reader to judge the uncertainty in the estimates of focal depth (appendix).

Constraints on the Fault Plane Solution

The determination of fault plane solutions for earthquakes in the Andes suffers from a dearth of WWSSN stations to the west. Only a few stations on islands in the Pacific, with low magnifications, sometimes yield reliable first motions. Thus, since most of the events in the area show reverse faulting, the dip of nodal plane dipping west is generally very poorly constrained (appendix).

A clear example illustrating this problem is the earthquake of May 15, 1976 (event 16). First motions for this earthquake are scarce and allow two different families of fault plane solutions (Figure 5). The solid line depicts an almost pure dip-slip solution with reverse faulting, while the dotted line indicates nearly pure strike-slip motion. Only two stations, WES and OGD in the eastern United States, produced long-period P waves large enough to be useful in the visual comparison with synthetic waveforms. WES and OGD are stations in close proximity, and rays to them leave the source with similar takeoff angles Thus the orientation of the fault planes cannot be resolved well. Synthetic waveforms, however, allow us to determine which of the two alternative types of faulting is likely [e.g., Langston, 1979]. The strike-slip solution does not produce synthetic waveforms that resemble the observed P waves, but synthetic P waves corresponding to the dip-slip mechanism closely resemble the observed wave trains (Figure 6). Varying the strike or dip of the nodal planes of the dip-slip solution by as much as 15° will not change appreciably the shape of the synthetic waveforms observed at these two stations.

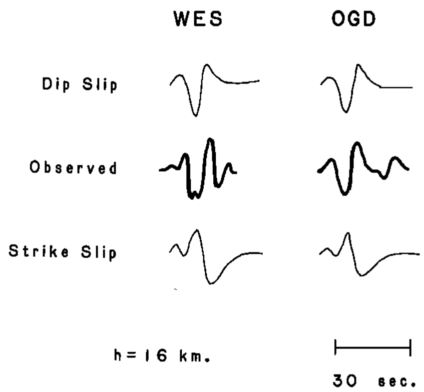


Fig. 6. Observed and synthetic long-period P waves at stations WES and OGD for the two alternative fault plane solutions shown in Figure 5. The observed P waves are matched better by the dip-slip solution. Each of the strike, dip, and rake of the nodal planes, however, may be varied as much as 15° without appreciably changing the shape of the synthetic pulses.

Errors in Depth Determination

A comparison of focal depths reported in the International Seismological Centre (ISC) catalog and those determined in this study using synthetic waveforms shows large discrepancies in many cases (Tables 1 and 2). The problem stems in part from the sparse coverage of the local networks and the poor azimuthal distribution of the WWSSN stations. One of the primary objectives in this paper is the accurate determination of the focal depths of these events.

To illustrate the accuracy of focal depth estimates, Figure 7 shows that for the earthquake of March 20, 1972 (event 12), for focal depths shallower than about 36 km the time interval between the primary and the surface-reflected phases is too short. Similarly, for depths larger than about 42 km, the interval between the direct and reflected phases is too large so that the synthetic P waves are too broad compared to the observed waveforms (Figure 7). Thus we consider the focal depth in this case to be

TABLE 2. Source Parameters

Event	Date	P Axis Trend, deg	P Axis Plunge, deg	P Axis Trend, deg	P Axis Plunge, deg	Depth, km	Seismic Moment M_o, dyncm	No. of Stations Used to Estimate Mo	Location
1	May 10, 1963	95.4	4.2	5.0	1.42	16	8.85×10^{25}	16	sub-Andes, Ecuador
2	Nov. 3, 1963	90.0	5.0	270.0	85.0	18	2.73×10^{25}	24	sub-Andes, Peru
3	Feb. 9, 1967	260.2	8.7	353.8	22.26	32	3.60×10^{26}	20	sub-Andes, Colombia
4	June 19, 1968	284.3	11.4	150.2	73.8	20	1.96×10^{26}	19	sub-Andes, Peru
5	June 20, 1968	82.8	12.5	330.4	60.8	16	4.95×10^{24}	16	sub-Andes, Peru
6	Dec. 1, 1968	80.8	5.0	260.0	85.0	18	5.75×10^{24}	3	sub-Andes, Peru
7	July 24, 1969	77.0	23.0	205.8	55.9	6	1.81×10^{25}	14	eastern cordillera, Peru
8	Oct. 1, 1969	52.6	21.9	207.4	66.1	5	9.84×10^{25}	17	eastern cordillera, Peru
9	Feb. 14, 1970	260.0	9.0	80.0	81.0	28	9.71×10^{24}	9	sub Andes, Peru
10	Dec. 10, 1970	73.0	10.0	253.0	80.0	32	6.76×10^{26}	15	forearc, Peru
11	Oct. 15, 1971	264.1	0.1	174.1	14.1	8	2.98×10^{24}	6	Altiplano, southern Peru
12	March 20,1972	267.0	10.0	87.0	80.0	38	4.23×10^{25}	26	sub-Andes, Peru
13	Feb. 23, 1973	97.8	22.5	321.9	60.0	10	9.47×10^{24}	13	sub-Andes, Ecuador
14	Sept.27, 1974	285.0	5.3	194.8	1.76	6	3.67×10^{25}	1	Colombia shield
15	April 9, 1976	264.0	33.0	84.0	57.0	22	9.69×10^{25}	12	forearc, Ecuador
16	May 15, 1976	16.5	9.1	281.2	25.4	18	1.25×10^{26}	23	sub-Andes, Peru
17	Oct. 6, 1976	274.4	24.2	145.2	54.6	16	5.24×10^{24}	2	high Andes, Ecuador

deeper than 34 km and shallower than 44 km if the average crustal velocity were 6.0 km/s. This would be equivalent to an error in focal depth of about 5%. It is noteworthy that the focal depth reported by ISC was 52 km.

The inferred depth is also proportional to the assumed velocity of P waves in the crust. Assuming P wave velocities to lie between 5.5 and 6.5 km/s, the range of plausible upper crustal velocities, there is an additional 5-10% uncertainty to the estimated focal depth. Synthetic seismograms computed using crustal velocities within this range corroborated the inference that higher or lower velocities lead to inferences of proportionally shallower or deeper sources.

Discussion of Results

Of the 17 earthquakes studied, 10 occurred along the transition zone between the sub-Andes and the Cordillera Oriental. The rest are distributed among the high Andes (four), the forearc in Ecuador (two), and the Brazilian shield in Colombia (one).

Events in the Forearc

As discussed above, the forearc is, in general, seismically quiet. Of the events studied, only two earthquakes beneath the coast of Ecuador appear to have taken place in the forearc. The fault plane solution of event 15 shows a nodal plane striking north-south and dipping east at a shallow angle, as do those of most interplate events in this region [Pennington, 1981]. The estimated focal depth of about 22 km appears to be too shallow for an earthquake reflecting the underthrusting of the Nazca plate beneath the arc. Errors, however, in both focal depth and epicentral location would place this earthquake as an interplate event on this poorly known inclined seismic zone.

The solution for event 10 also shows thrust faulting on a north trending plane that dips approximately 45° either east or west. Stauder [1975] had reported a fault plane solution for this event, indicating thrust faulting on nodal planes striking east-west, and attributed it to deformation of the overriding plate. The estimated focal depth of event 10 is 32 km, and it is not clear where it lies relative to the inclined seismic zone. The dip of 45° of the east dipping nodal plane is too steep for this event to occur along the gently dipping seismic zone that marks the plate boundary (Figure 9). This zone appears to dip more gently beneath Ecuador than it does farther south, but it is very poorly defined by teleseismic data (see Figure 5 of Barazangi and Isacks [1979]). Thus the depth of event 10 (32 km) is sufficiently deep that it is difficult to ascribe it to internal deformation of the overriding plate, and the dip of the east dipping nodal plane is too steep to reflect slip on the main inclined seismic zone. In any case, neither event 10 or 15 clearly indicate deformation of the overriding plate at the subduction zone.

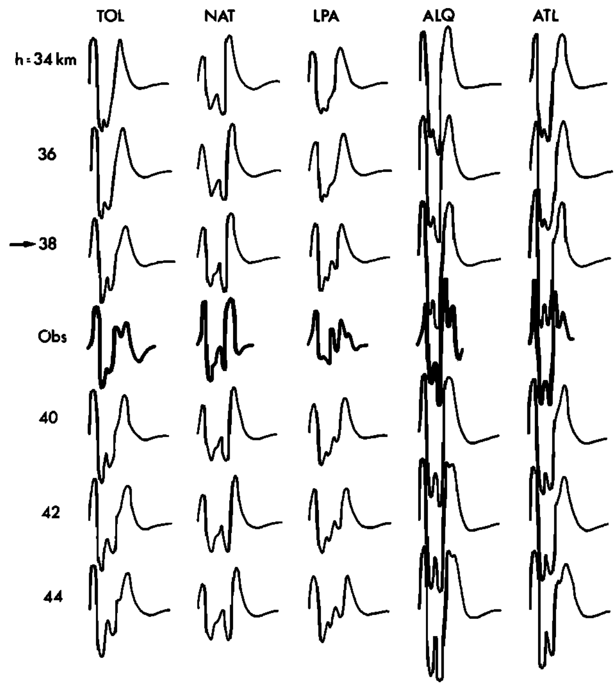


Fig. 7. Comparison of the recorded P waveforms from the event of March 20, 1972, with a suite of synthetic waveforms for different focal depths using the fault plane solution shown in the appendix. Synthetic P waveforms for depths shallower than 36 km or deeper than 40 km do not match the observed waveforms well.

The Brazilian Shield

The fault plane solution of event 14 in the Brazilian shield indicates strike-slip motion with the P axis oriented in an east-west direction. The orientation of the P axis parallel to the direction of subduction suggests that the Brazilian shield experiences compression in that direction and in response to convergence. This is consistent with the solutions of five shallow events in cratonic South America showing P axes oriented approximately east-west [Mendiguren and Richter, 1978].

Tectonics of the Sub-Andes

Focal depths and style of deformation. Most of the earthquakes studied are concentrated along the eastern flank of the Andes on the western edge of the sub-Andes. They occur beneath regions of low topographic elevation and consistently show thrust faulting approximately parallel to the mountain belt (Figure 8). In most cases, the nodal planes dip at steep angles, between 30° and 60° (Figure 11), and we cannot determine which of the two nodal planes was the fault plane. The events are too small for the body waves to show azimuthal differences in pulse shapes, due to the finiteness of the source, that could be used to select the fault plane. Moreover, the lack of local stations did not permit the recording and accurate locating of aftershock sequences. Nevertheless, because most of the thrust faults in the sub-Andes dip west

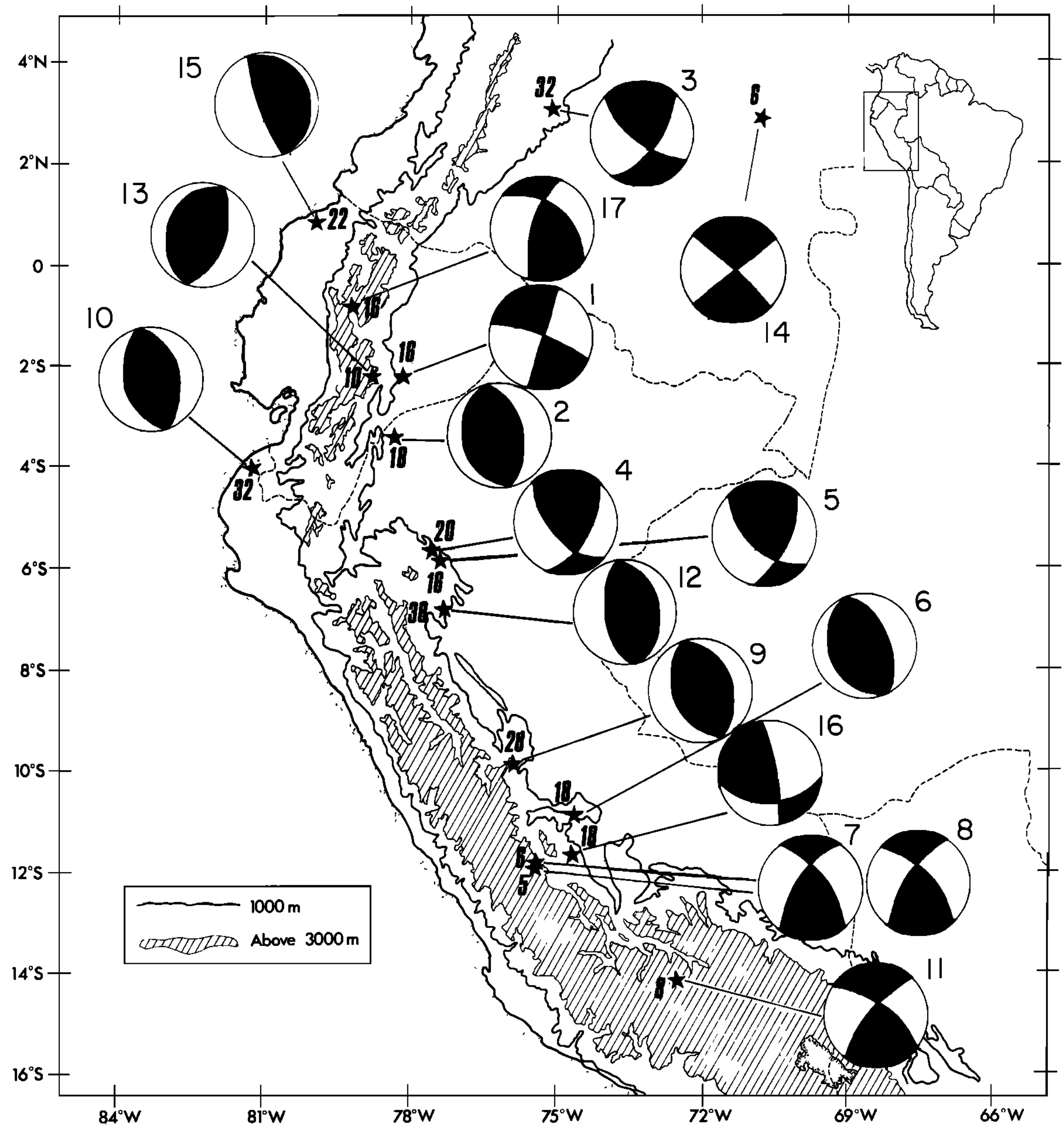


Fig. 8. Map of the Peruvian Andes summarizing fault plane solutions and focal depths. Shaded areas in lower hemispheric projections of focal spheres indicate quadrants with compressional first motion. Epicenters are plotted as stars and the numbers beside them indicate estimated focal depths in kilometers. Numbers next to balloons correspond to those in Table 1 and the appendix.

[Dalmayrac, 1978; Dalmayrac et al., 1980; Megard, 1978], we presume that the west dipping nodal planes are the fault planes (Figures 9 and 10). Although west dipping nodal planes are usually the more gently dipping of the two nodal planes, the dips are still much steeper (Figure 11) than those associated either with decollement in a thin-skinned tectonic environment as in the Canadian Rockies [Bally et al., 1966; Price and Mountjoy, 1970] or with surface exposures of large crystalline nappes, such as those in Norway [Gee, 1978]. At first glance, the steep dips appear to be more reminiscent of faults active in the early Tertiary (Laramide) in the Rocky Mountains of Colorado and Wyoming [e.g. Brewer et al., 1980; Smithson et al., 1979] or of those deduced from fault plane solutions of earthquakes in the Tien Shan [e.g., Tapponnier and Molnar, 1977] or the Zagros Mountains [Jackson and Fitch, 1981; McKenzie, 1972].

The earthquakes studied in the sub-Andes are relatively deep, ranging in depth from 8 to 38 km, compared with those in the Zagros Mountains with focal depths of 10-15 km [Jackson and Fitch, 1981]. Thus they indicate that most of the crust and possibly the uppermost mantle are involved in the deformation. The inferred depth of the deepest event is between 36 and 40 km. Unfortunately, we are not aware of any studies of crustal thickness in the sub-Andes that resolve the question of whether these deeper events occurred in the uppermost mantle or in the lower

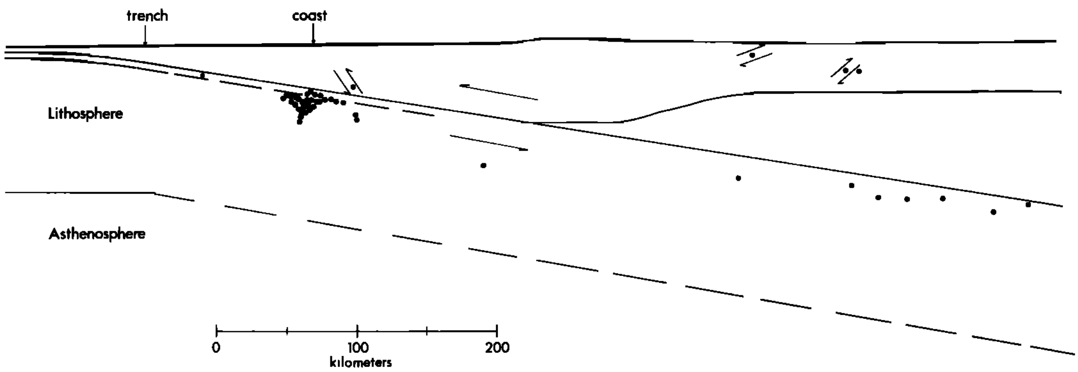


Fig. 9. Cross section of seismicity across Ecuador. Solid dots show position of earthquakes selected by Barazangi and Isacks [1979] in cross section B-B'. These hypocenters are used to determine the dip of the slab. The east dipping nodal plane of event 10 in the forearc appears to be too steep to be associated with slip on the main inclined seismic zone. The west dipping nodal planes of earthquakes 2, 4, and 5 in the sub-Andes are also projected on the cross section. The position of the Moho beneath the Andes is drawn so as to maintain isostatic equilibrium but is not constrained by other data.

crust. None of those studied here, however, occurred as deep as 80 km, as suggested by Stauder [1975]. Regardless of the depth of the Moho, it is clear that the earthquakes occurred in the crystalline basement and not in the overlying sedimentary cover. This fact is consistent with the inference that the earthquakes do not result from slip along a decollement between the crystalline basement and overlying sedimentary strata.

Relating the earthquakes to specific faults identified from geologic field work is difficult, both because of inaccuracies in the locations of the earthquakes and because of a lack of detailed geologic studies of the sub-Andes. The best studied segment of the sub-Andes is in Central Peru [Audebaud et al., 1973; Dalmayrac, 1978; Megard, 1978]. Numerous west dipping thrust, or reverse, faults can be inferred from the juxtaposition of rocks of different ages. Although, faults are drawn dipping at steep angles (60°-70°), the descriptions of these regions are not sufficiently detailed to constrain the dips below the surface. In the regions that we have seen, albeit only during a cursory field reconnaissance, the fault planes generally were not exposed, and faults could be inferred only from the juxtaposition of rocks of very different ages. The style of internal deformation of the rocks near the faults, however, suggested that they are thrust and not reverse faults. From the gentle west dipping cleavage present in Mesozoic and Cenozoic limestones and red sandstones and from the gentle west dips of some minor faults, we suspect that many of the thrusts dip at more gentle angles (less than 20°) than are shown in published cross sections.

One of the basic questions to be addressed by this study is to what extent does the deformation in the sub-Andes resemble that of the Canadian Rockies during the Late Cretaceous in one extreme or the Colorado-Wyoming Rockies during the early Tertiary in the other. The steeply dipping nodal planes of earthquakes in the basement of the sub-Andes are reminiscent of the type of faulting in the Colorado-Wyoming Rockies, where the basement is clearly involved and some fault planes are quite steep (>30°) [Brewer et al., 1980; Smithson et al., 1979]. Nevertheless, the

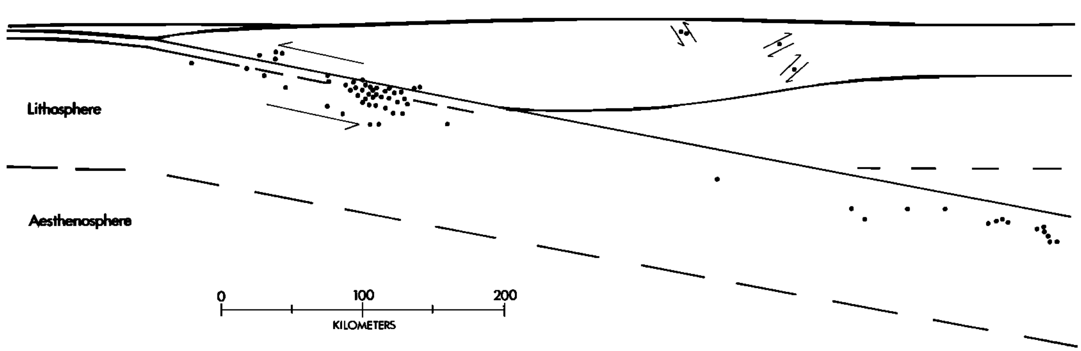


Fig. 10. Cross section for seismicity through central Peru (symbols as in Figure 9). The dip of the slab is inferred from the hypocenters of earthquakes selected by Barazangi and Isacks [1979] in their cross section C-C^1. West dipping nodal planes of two of the earthquakes studied here (6 and 16) from sub-Andes are projected onto the cross section. The fault plane of the Pariahuanca earthquakes (7 and 8) in the high Andes is known to dip east, as observed in the field [Philip and Megard, 1977].

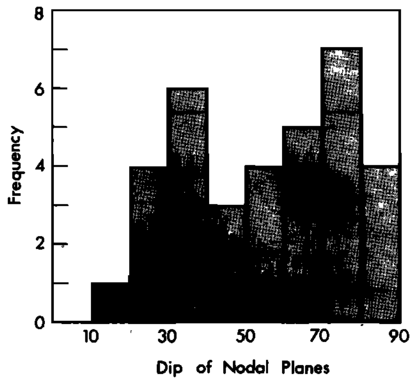


Fig. 11. Histogram of the dips of the nodal planes of the earthquakes studied.

similarity is far from perfect. In particular, whereas the ranges in Colorado and Wyoming apparently are bounded and uplifted by well-defined, localized fault zones [Gries, 1983], the earthquakes in the sub-Andes are relatively deep and occur only in the western margin of the province, while at the surface the most recently deformed rocks appear to be on the eastern margin. Thus, while the basement beneath the transition from the Cordillera Oriental and the sub-Andes undergoes crustal shortening, the youngest deformation of the cover occurs some 50-100 km to the east. In the Canadian Rockies, faults dip steeply at the surface but flatten with depth and detach the sedimentary cover from the underlying basement. As a result of this, a broad zone (~200 km) of folded and thrust-faulted sedimentary rocks overlies the basement, which dips gently and without any apparent major disruption [Bally et al., 1966; Price and Mountjoy, 1970]. The earthquakes in the Peruvian Andes clearly occurred within the basement; the nodal planes are too steep and the focal depths of these events are too deep for them to have occurred in the sedimentary cover or at its contact with the basement. There is no obvious systematic change with depth of the dips of the nodal planes as might be expected if fault surfaces curved smoothly with depth. Thus the earthquakes and brittle deformation of the basement in Peru probably do not occur in a setting similar to that of the wide zone of detached sedimentary rocks in the Canadian Rockies.

In Peru, the earthquakes might indicate deformation of the basement west of a detached sub-Andean terrane, and if there were a sequence of thick, competent sedimentary rocks in the sub-Andes of central Peru, they would show a style of deformation similar to that of the Canadian Rockies. Thus the shortening in the basement would occur west of the surficial expression of shortening in the sedimentary cover, and the two areas would be related by gently dipping, apparently aseismic decollement. Considering the weak and foliated nature of the Precambrian crystalline rocks that crop out in the Cordillera Oriental, it is possible that some of these crystalline rocks are also detached from a deeper and more rigid substratum.

The present regime of steep basement-involved thrust faults is similar to the style of basement deformation that probably prevailed in the high Andes during their formation. Throughout much of Peru the sedimentary cover is tightly folded, sometimes in localized zones of concentrated deformation and with occasional steep faults [e.g., Coney, 1971; Lepry and Davis, 1983; Megard, 1978]. Except for a portion in north-central Peru, neither a preferred vergence nor large overthrusts can be recognized. Therefore it is appropriate to suggest that the active tectonic regime that we observe today in the sub-Andes is similar to the process that produced extensive crustal shortening and is responsible for the formation of the Central Andes.

Rate and amount of crustal shortening. This interpretation motivates a comparison of the rate of crustal shortening obtained by adding the seismic moments of earthquakes with that implied by the geologic history. The definition of seismic moment is

$$M_0 = \mu S u \tag{1}$$

[Aki, 1966], where M_0 is the scalar seismic moment, S is the surface area of the fault, u is the average slip of the fault, and μ the shear modulus. Thus the cumulative slip on a single fault can be estimated by simply summing the seismic moments of earthquakes causing slip on the fault [e.g., Brune, 1968; Davies and Brune, 1971]. In the central Andes, however, seismic energy release does not occur along a single major fault. Instead, earthquakes occur on fault planes with different orientations distributed over a large seismogenic volume. In this case, it is more appropriate to follow Kostrov [1974] and use the seismic moment tensors to calculate the strain resulting from the motion of all the faults in the region [e.g., Wesnousky et al., 1982; Chen and Molnar, 1977].

The seismic moment tensor for an arbitrary shear dislocation may be expressed as

$$M_{ij} = \mu A[u_i n_j + u_j n_i] \tag{2}$$

where n_i are the direction cosines of the vectors normal to the fault plane, and u_i is the average slip in vectorial form. The principal values of the moment tensor, $-M_0$, 0, and $+M_0$, correspond to the P, B, and T axes of fault plane solutions [Gilbert, 1970]. Kostrov [1974] showed that the mean rate of irrotational strain in a volume V over a period of time t due to slip on n different faults within that volume is

$$\dot{\varepsilon}_{ij} = \frac{1}{2\mu Vt} \sum_{n=1}^{N} M_{ij}^{n} \tag{3}$$

In the region studied, the areal dimensions of the deformed volume are about 2000 km in length and 250 km in width. We assume a thickness of 40 km, corresponding roughly to the deepest intracontinental earthquake studied.

Diagonalizing $\dot{\varepsilon}_{ij}$ yields the principal directions of the average strain rate. Crustal shortening is computed by multiplying the horizontal component of the maximum compressive strain rate times the width of the deformed body in the direction of maximum compressive strain.

In the Central Andes, the principal value of compressive strain rate is 6×10^{-9} yr^{-1} oriented about 95° west of north. This orientation is consistent with the direction of subduction of the Nazca plate beneath South America. The average rates of crustal shortening is between 1.0 and 1.7 mm/yr. The smaller figure is obtained by not including event 10 in the calculations. As discussed above, it is possible that event 10 occurs in the main thrust contact between the Nazca and South American plate and may not be the result of crustal deformation of the continental plate. These rates of crustal shortening are lower bounds since they assume that the deformation is being taken up only by brittle fracture and disregard the possibility of fault creep and viscoelastic deformation. The rates also include only contributions from earthquakes with magnitudes between 5.4 and 6.3. Combining the frequency-magnitude ($\log N = a-b \cdot M_s$) and moment-magnitude ($\log M_o = 1.5 \cdot M_s+16$) [Hanks and Kanamori, 1979] relations, where N is the number of earthquakes, M_s is the surface wave magnitude, and M_o the seismic moment, the contribution due to earthquakes smaller that magnitude 5.4 can be estimated [e.g., Molnar, 1979]. Assuming a value of b equal to 1, the contribution to the total moment by events with $M_s < 5.4$ is only 37% of that of events with magnitudes between 5.4 and 6.3. Thus average crustal shortening rates would be of the order of 1.4-2.1 mm/yr.

The total amount of crustal shortening can be estimated assuming that crustal volume is conserved and excess material beneath the mountainous topography and in the root of the Andes corresponds to South American crust that subsequently has been substantially shortened. Assuming an average elevation of 4 km and a crustal thickness of 60 km [James, 1971], then for a crust originally 35 km thick, we obtain 190 km of crustal shortening. This estimate neglects the loss of crustal material through erosion or addition by magmatic processes. Even if 10 km of material had been eroded from the top of the Andes over a width of 250 km, these estimates of crustal shortening would increase by only 35%. Thus we do not consider ignoring erosion to be important for the significance of the following calculations. More problematic is the quantity and source of plutonic rocks in the Andes. To our knowledge there are no reliable estimates of the amount of such rocks in the Peruvian Andes, and we simply ignore them, recognizing that the amount of crustal shortening estimated here is an upper bound. One justification for ignoring the volume of batholiths is the recent isotopic evidence that the batholiths are not derived entirely from the mantle and that remelted crust is a major source for them [e.g., Allègre and Ben Othman, 1980; De Paolo, 1981]. Therefore the existence of batholiths may not imply a large addition of material to the crust. For contrast, from the structures and style of deformation in a geologic cross section traversing the Central Andes, Megard [1978] suggests crustal shortening to be of the order of 100 km.

Shortening of 190 km of Andean crust at a rate of 1.4-2.1 mm/yr would require 90-135 m.y. Although, subduction probably occurred throughout most of the Mesozoic [e.g., Helwig, 1973], paleogeographic interpretations have indicated the presence of a major transgression ending in the Late Cretaceous [e.g., Audebaud et al., 1973; Zeil, 1979]. Therefore much of the present Andes were at sea level about 70-80 m.y. ago, and the present topography and thickened crust must have formed since then.

Because of the short record of seismicity, our ignorance of recurrence rates, and the unavoidable approximate assumptions made in estimating the rate and amount of crustal shortening, the agreement between the average rates of shortening inferred from the seismic moments of 17 years of seismicity and from about 70 m.y. of geologic history is not a strong argument that the present tectonics of the sub-Andes are similar to those of the Andes as a whole throughout the last 70 m.y. or so. However, agreement within an order of magnitude suggests that if most of the topography, and implicitly the associated crustal root as well, developed in latest Cenozoic time, as often presumed [e.g., Audebaud et al., 1973], then the Andes must have been built by some process other than crustal shortening.

Earthquakes in the High Andes

Thrust faulting at high elevation. Only four large earthquakes occurred in the high Andes in the last 20 years: event 17, in Ecuador, events 7 and 8 in the Cordillera Oriental of southern Peru, and event 11 in the northern Altiplano (Figure 8). These earthquakes are shallow; focal depths of none of them are deeper than about 20 km. Among all of the earthquakes studied here only event 7 is known to have produced a surface break in the Andes. Events 7 and 8 occurred near the Huaytapallana fault, which crops out at an elevation of about 4600 m above sea level. The fault scarp indicates reverse faulting with a component of left-lateral strike-slip motion on a plane dipping about 60°NE and striking N50°W [Philip and Megard, 1977]. In the basins just west of this fault, Dollfus and Megard [1968] reported folding of Quaternary glacial terraces. The strike of the folds is in a direction N30°W and is consistent with the strain pattern deduced from the fault.

Event 11 is the only earthquake in the Altiplano large enough to be studied. It has a focal depth of 8 km, and the fault plane solution indicates almost pure strike-slip motion with north-south oriented T axes. In Ecuador, event 17 occurs at a depth of 16 km and shows a thrust mechanism with the P axes oriented approximately east-west.

Normal faulting in the high Andes. Although thrust faulting is the prevalent style of deformation throughout much of the Andes, normal faulting has been documented in a number of localities in the high Andes. Normal faults clearly offset Quaternary glacial moraines along the eastern margin of the Cordillera Blanca of central Peru [Dalmayrac, 1978; Yonekura et al., 1979]. Yonekura et al. [1979] recognized a system of normal faults on the satellite imagery more than 100 km long (Figure 12). The strike of the faults is roughly parallel to the Andes, indicating a component of extension perpendicular to the mountain range. Furthermore, in the Altiplano of southern Peru and northern Bolivia,

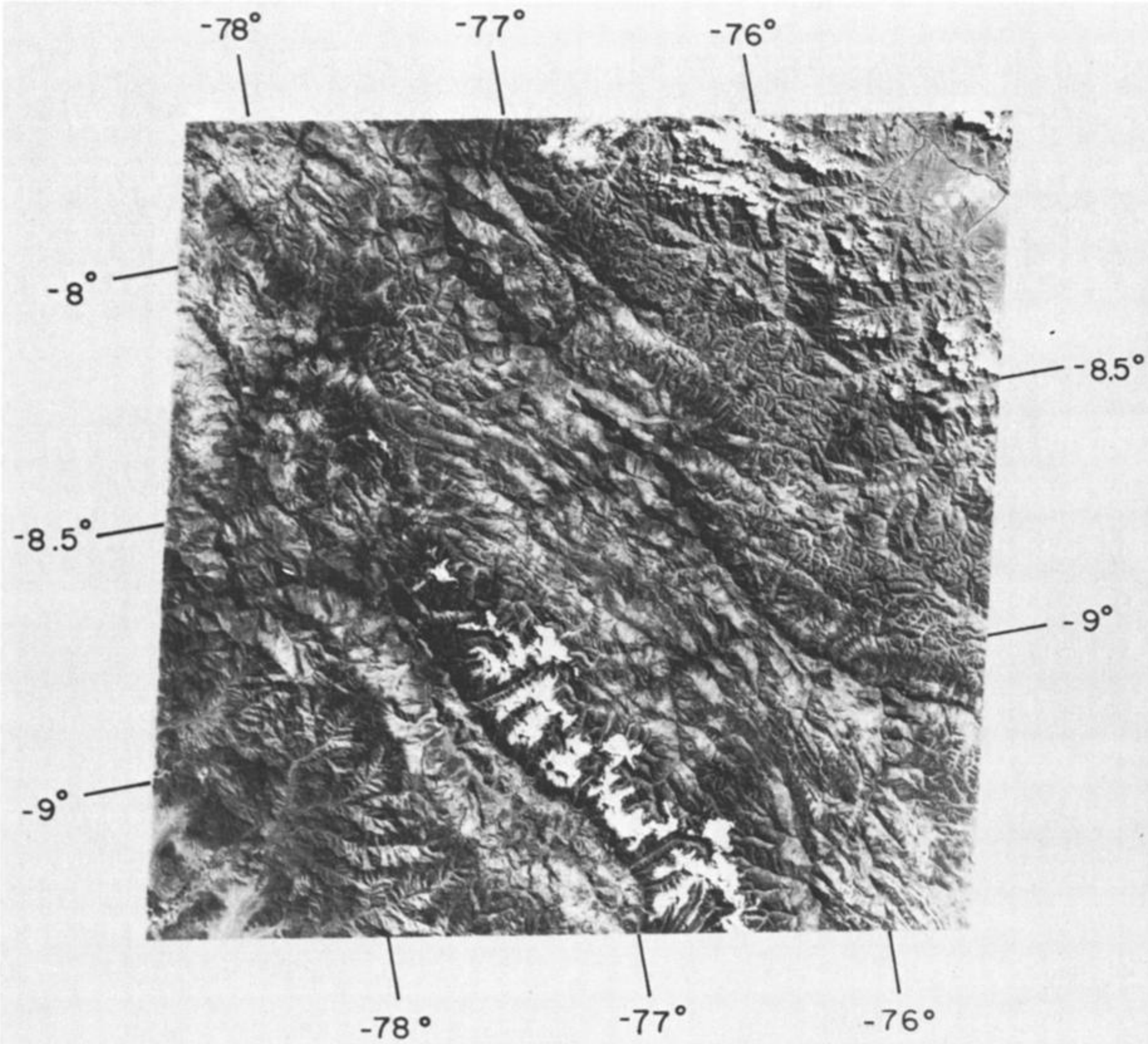


Fig. 12. Landsat image 2194-14345-7 of the area north of the Cordillera Blanca (snow-capped range). Steep escarpment to the west of the range defines zone of normal faulting.

normal faults offset Quaternary sediments [Lavenu, 1978; Lavenu and Ballivian, 1979; Mercier, 1981]. During field work in the summer of 1980, we discovered another clear, recently active normal fault near the northern edge of the Altiplano. The fault scarp is not prominent on satellite imagery, but it is clear on aerial photographs (Figure 13) and from the ground (Figure 14).

The only large earthquake recorded in areas where normal faulting has been reported is the 1946 Ancash earthquake [Richter, 1958]. The epicenter is located about 100 km east of the Cordillera Blanca. Silgado [1951] described the scarps produced by this event and concluded that normal faulting occurred on a plane dipping west at about 45°. Moreover, the fault plane solution (Figure 15) [Hodgson and Bremner, 1953] shows normal faulting with one of the nodal planes parallel to the strike of the fault scarps observed by Silgado [1951]. It has also been suggested that normal faulting was also the cause of both the 1949 Ambato earthquake in the high Andes of Ecuador (B. Foxall, personal communication, 1982) and the 1958 Maipu earthquake in Chile [Jordan et al., 1983].

Implications of normal and thrust faulting in the high Andes. The coexistence of normal faulting at high elevation and thrust faulting at the eastern edge of the mountain belt, within a few tens of kilometers of each other, can be explained by variations in the stress field that must balance the gravitational body force acting on the regions of high elevation and their associated crustal roots [Dalmayrac and Molnar, 1981]. To balance the lateral variations in the gravitational body force, there must be a gradient in stress. This balance can be maintained if the horizontal compressive stress exceeds the vertical stress in regions of low elevation, where thrust faulting prevails. Simultaneously, or alternatively, the vertical compressive stress can exceed the horizontal stress in regions of high altitude (and thickened crust), where normal faulting occurs. In portions of the Central Andes both seem to occur concurrently.

For topography in the form of a ridge on an elastic half space, the horizontal compressive stress induced by gravity acting on the ridge approaches a maximum on the flanks of the ridge and is minimum at the peak [McTigue and Mei,

Fig. 13. Aerial photograph of an active normal fault in the Altiplano of southern Peru. Arrow points to the fault scarp. Photo shows an area of approximately 8 by 8 km. The town of Sangarara (-13.95, -71.60) is near the center of the photograph.

1981]. For large characteristic slopes, the horizontal stress can become purely tensile at the peak of the ridge (Figure 16).

The two segments of the Central and Northern Andes where both thrust faulting and crustal shortening are observed in the high Andes coincide with the regions where aseismic oceanic ridges are being subducted: the Galapagos-Carnegie ridge in Ecuador and the Nazca ridge in southern Peru. It is qualitatively easy to ascribe an excess horizontal compressive stress to resistance to subduction of the buoyant ridges beneath these portions of the Andes, as has been suggested before to explain other phenomena [e.g., Kelleher and McCann, 1976; Pilger, 1981; Vogt et al., 1976].

Implications for Andean Evolution

In some ways, the active tectonics and tectonic setting are analogous to those present in the Himalaya and Tibet [Molnar and Tapponnier, 1975, 1978; Molnar et al., 1977]. Thrust faulting on the margins of Tibet occurs concurrently with normal faulting in the highest part of the plateau. The stress needed to cause slip on a thrust fault at low elevations apparently is less than that required to uplift the topography beyond a certain elevation. Consequently, the region of elevated topography seems to grow laterally instead of increasing its elevation.

The near absence of thrust faulting and in some cases extension along much of the high Andes suggest that no significant amount of crustal shortening is taking place there. Unless deformation occurs in a ductile fashion, the active tectonic regime of the high Andes is mild compared to that in the sub-Andes, and the rates of extension or shortening are probably low. This tectonic style, comprising extension in some areas and shortening in others, suggests a delicate balance between the compressive forces applied to the flanks of the Andes by the converging Nazca plate and Brazilian shield and the gravitational force applied to the elevated crust and its thickened root. It appears that the Andes transmit the stresses applied by one side to the other, much as Tibet seems to do between India and Eurasia [Molnar and Tapponnier, 1978]. Thus the forces driving the plates together apparently are not elevating the portion of the Andes that is already at high altitudes, increasing the large amount of gravitational potential energy already stored there. Instead, they seem to do work by breaking the crust where

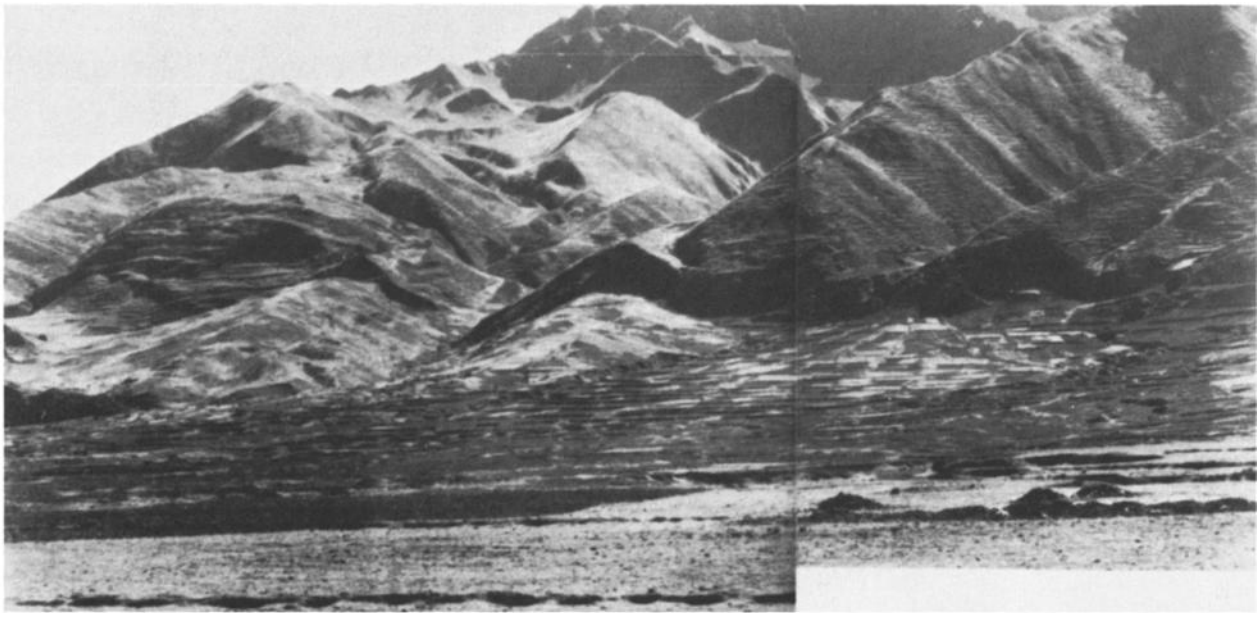


Fig. 14. Photograph of normal fault shown in Figure 13 taken from the ground at a distance of about 2 km from the fault using a lens with a focal length of 135 mm. Symbol arrow in Figure 13 shows location where the photograph was taken. The fault scarp clearly offsets Quaternary sediments and moraines.

it is thinner by underthrusting the Brazilian shield beneath the eastern margin of the mountain belt. This has the effect of thickening the crust where it is thin and causing topographic relief to grow progressively toward the east. In this manner, when the elevation reaches a critical height, faults are progressively created farther east, and new material is constantly appended to the Andes along their eastern margin, maintaining an average elevation of its highest part (Figure 17).

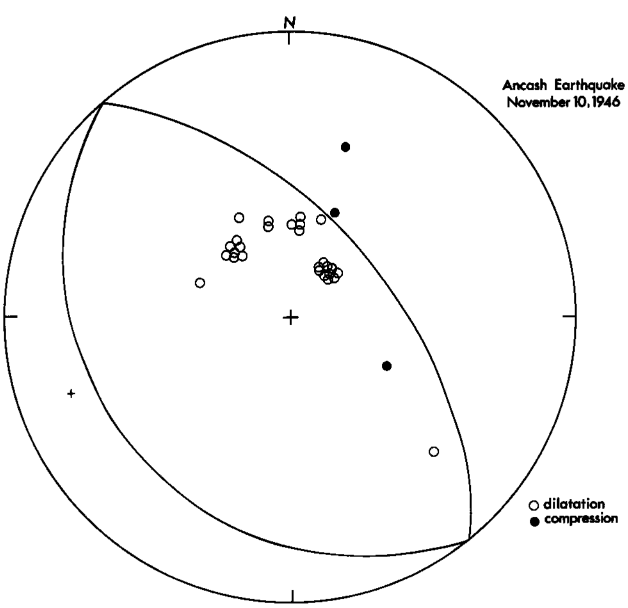


Fig. 15. Fault plane solution of the November 10, 1946, Ancash earthquake. Open circles indicate dilatational, and solid circles compressional first motions (from data collected by Hodgson and Bremner [1953]).

Crustal earthquakes in other tectonically active areas of the world usually occur at depths of less than 15 km [e.g., Chen and Molnar, 1983] and the lower crust apparently deforms in a

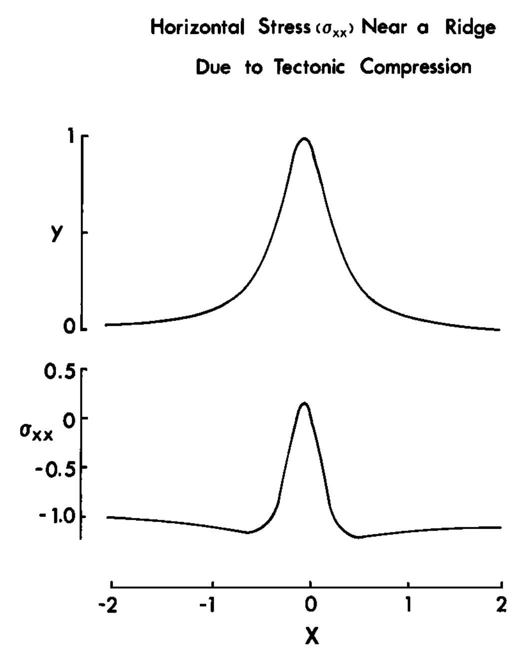


Fig. 16. Stress at top surface of an elastic, two-dimensional ridge subjected to horizontal compression perpendicular to and at a large distance from it. The horizontal compressive stress reaches a maximum (negative value) at the edges of the ridge. At the center of the ridge the value of the horizontal stress decreases, and for large characteristic slopes it may become purely tensile (positive) [after McTigue and Mei, 1981].

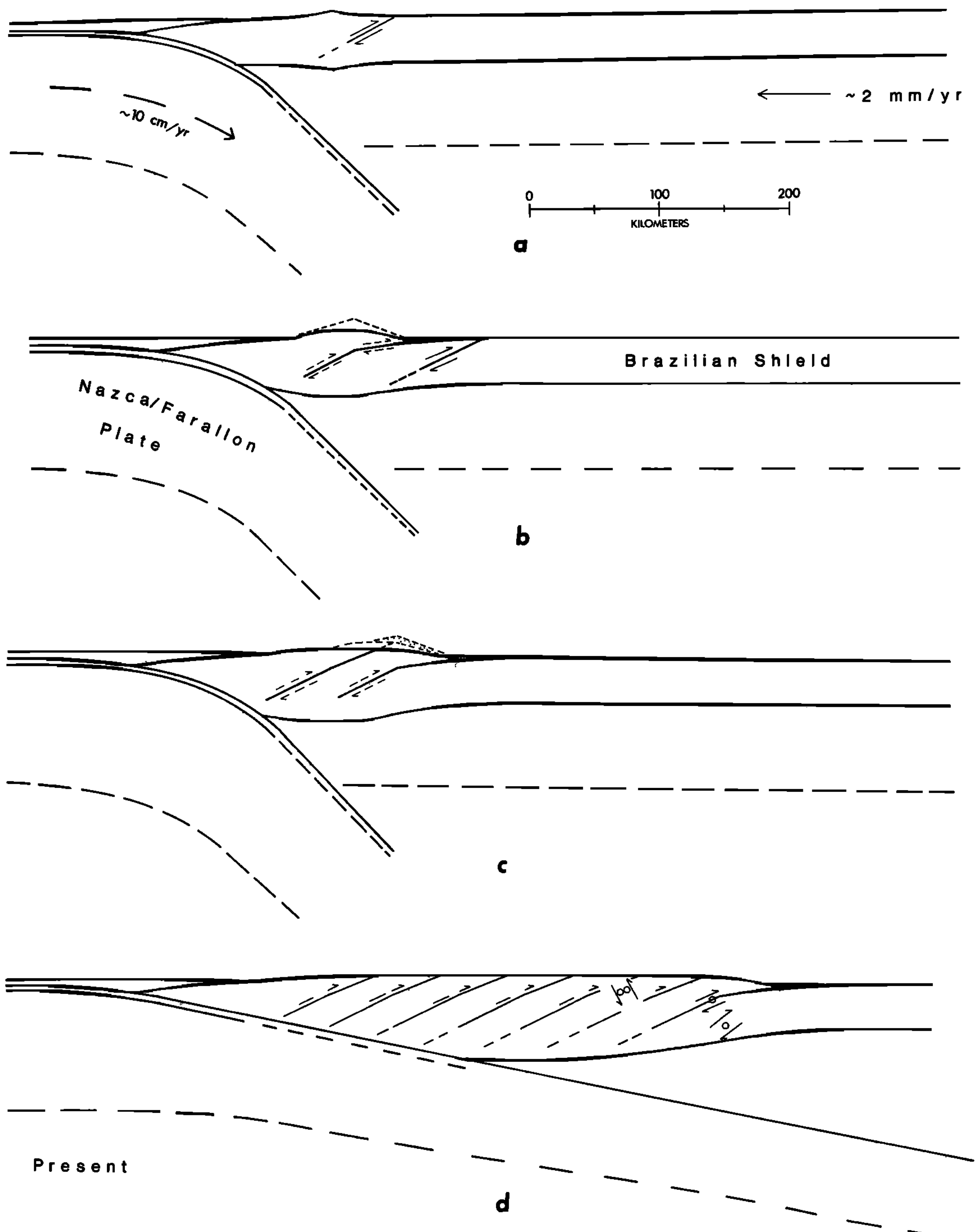


Fig. 17. Sequence of idealized cross sections across the Peruvian Andes through time showing the subduction of the Nazca plate beneath Peru and slow convergence (~2mm/yr) of the Brazilian shield with the Andean range. The trench and volcanic arc are arbitrarily held fixed. (a) Underthrusting of the Brazilian shield east of the incipient mountain belt. (b) and (c) The crust thickens and uplifts the topography to an equilibrium height and a new fault breaks to the east. (d) Present day topography and crustal thickness. Circles indicate earthquakes plotted in Figure 10. The regular spacing of faults in (d) is drawn to illustrate the lateral growth of the range and is not meant to imply such a regularity in a real cross section.

ductile manner. The presence of earthquakes in the lower crust beneath the sub-Andes, at depths of 20-38 km, indicates the presence of cold crustal material subject to brittle deformation. The existence of cold and brittle material at these depths suggests the presence of an abnormally low geothermal gradient in the sub-Andes, which, in turn, is probably due to the underthrusting of shallow and cold basement rocks from the Brazilian shield under the eastern margin of the Andes.

Extension in the high Andes is probably due to body forces produced by gravity and to buoyancy forces exerted by the crustal root of the Andes [Dalmayrac and Molnar, 1981]. This delicate balance of thrust faulting along the eastern piedmont and extension in the highest part of the Cordillera helps maintain an average equilibrium topography. In this scenario, one can imagine that the cessation of subduction along the western margin of the mountain belt may reduce the horizontal compressive stress required to support the topography. The Andes might then collapse by extensive normal faulting.

Summary

By comparing synthetic and observed long-period P waveforms we determined fault plane solutions and depths of foci of all earthquakes occurring in the Andes of Peru, Ecuador, and southern Colombia, sufficiently large to be studied with data from WWSSN. In general, fault plane solutions indicate thrust faulting along nodal planes striking northwest approximately parallel to the mountain belt. Nodal planes dip at high angles of about 30°-60°. The fault plane solutions of these intracontinental earthquakes seem to reflect deformation of the South American plate probably due to the convergence with the Nazca plate.

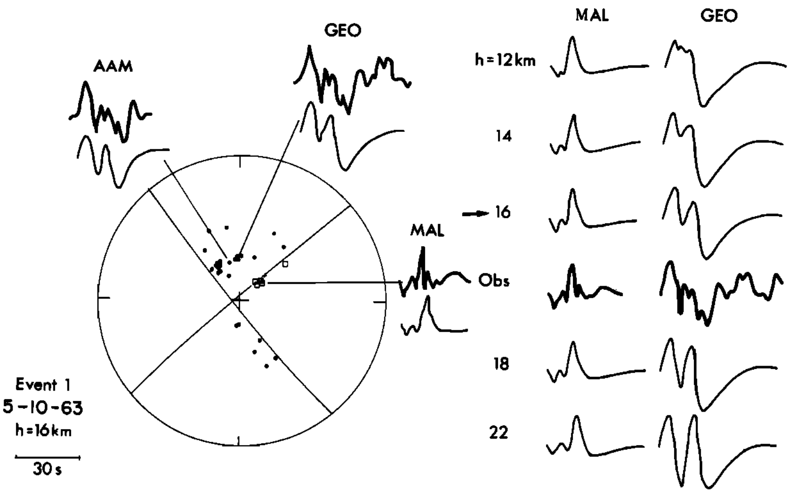
Most of the earthquakes occur on the western margin of the sub-Andes in the transition zone from the sub-Andes to the Cordillera Oriental and beneath regions of low elevation (~1000 m). The focal depths of earthquakes range from 10 to 38 km, suggesting that the lower crust and possibly the uppermost mantle are involved in the deformation. The earthquakes in the sub-Andes are too deep and their nodal planes are too steep to associate them directly with decollement of the sedimentary rocks from the underlying basement in a thin-skinned tectonic environment. These earthquakes, however, may reflect deformation in the hinterland of a detached fold and thrust belt in the sub-Andes, and therefore they might reflect the style of deformation that took place in the crystalline terrain west of the fold and thrust belt of the Canadian Rockies.

A few earthquakes occur in the Cordillera Oriental. Earthquakes there are shallow with the focal depths of none of the four events studied exceeding 16 km. Two events that occurred in the Cordillera Oriental in southern Peru are the only two earthquakes in the last 20 years associated with surface fault breaks [Philip and Megard, 1977].

While thrust faulting at high angles is prevalent on the eastern side, normal faulting has been documented in a number of localities in the high Andes. The normal faults trend in a direction parallel to the mountain belt, suggesting that some extension takes place in a direction normal to the strike. They occur in areas of high topographic relief, and faulting associated with one earthquake (the 1946 Ancash earthquake) reflects this extensional regime.

Excess crustal material in the topography and the root of the Andes indicates that as much as 190 km of crustal shortening may have taken place. A summation of the seismic moment of earthquakes in the sub-Andes yields a rate of shortening of 1.4-2.1 mm/yr. If these rates were comparable to those in the geologic past, the formation of the Andean topography and crustal root would have occurred in the last 90-135 m.y. Unless the deformation is aseismic and ductile, no significant amount of crustal shortening is taking place in the high Andes. Most of the shortening, as reflected by the seismicity, takes place in the western sub-Andes. We infer that underthrusting of the Brazilian shield along the eastern side of the massif thickens the crust and creates topographic relief. New faults are successively created farther east when the topography reaches a critical height, so that the mountain belt grows eastward in areal extent.

Appendix



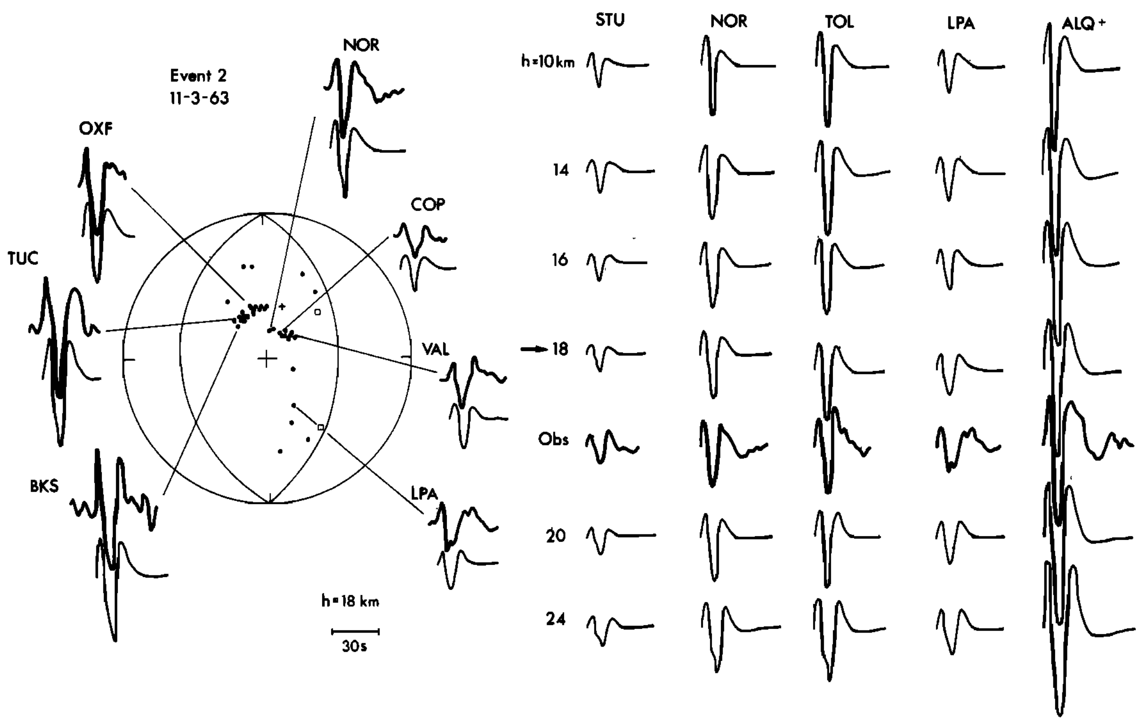


Fig. A1. Fault plane solutions and synthetic waveforms for the earthquakes listed in Tables 1 and 2. First-motion data are plotted on equal-area lower hemispheric projections where the solid circles indicate compressional first motions and open circles dilatational first motions. Observed long-period P waves and synthetic waveforms for the fault plane solutions shown are also presented for a few stations. Not all the stations used in the analysis are shown since many of them are clustered at similar azimuth and epicentral distance from the source and have very similar waveforms. Synthetic waveforms computed at various focal depths are also shown versus the observed long-period P waves at a few stations spanning a range of azimuths and epicentral distances. Arrow indicates preferred focal depth. + idicates amplitude was reduced in half for plotting purposes.

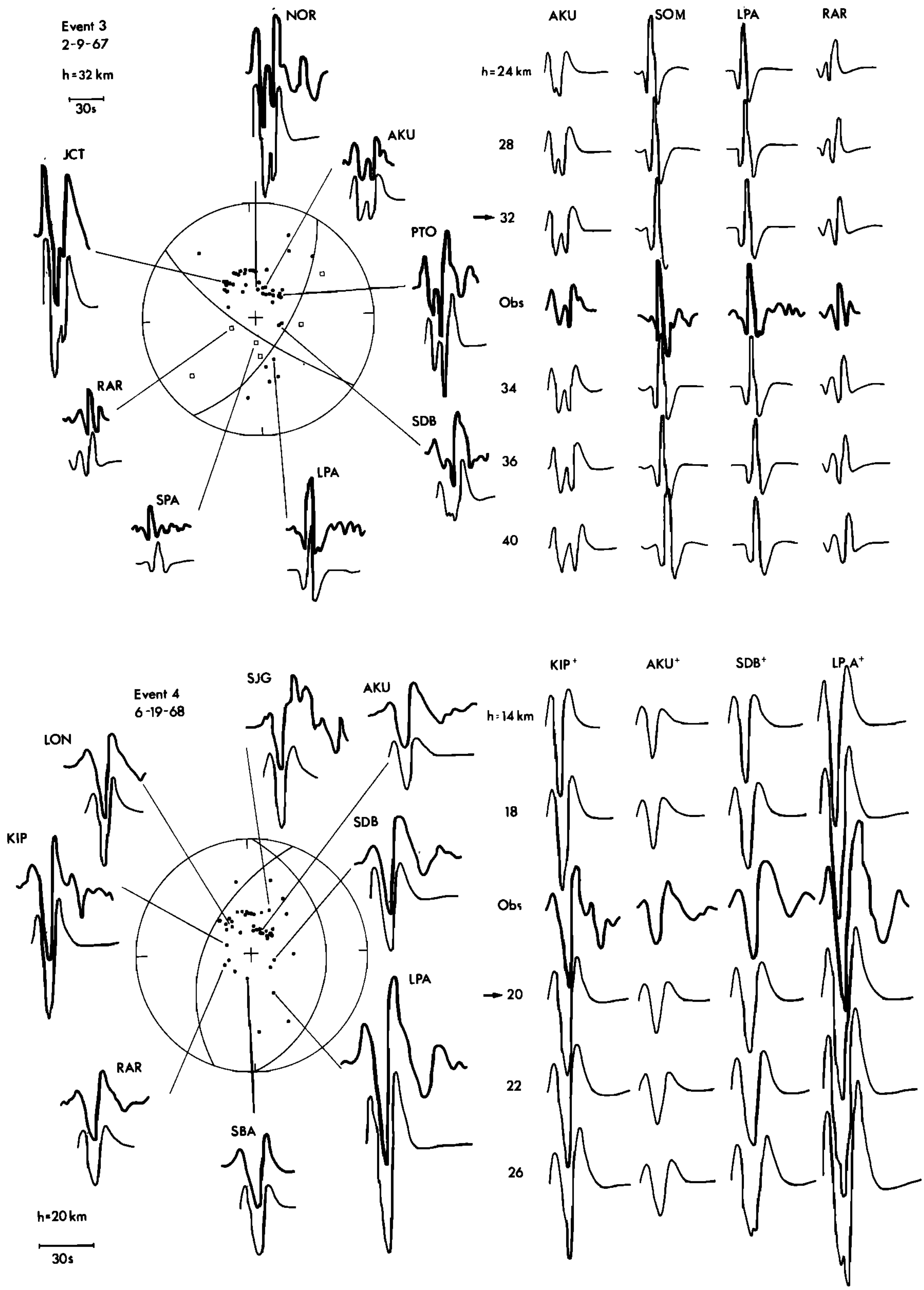
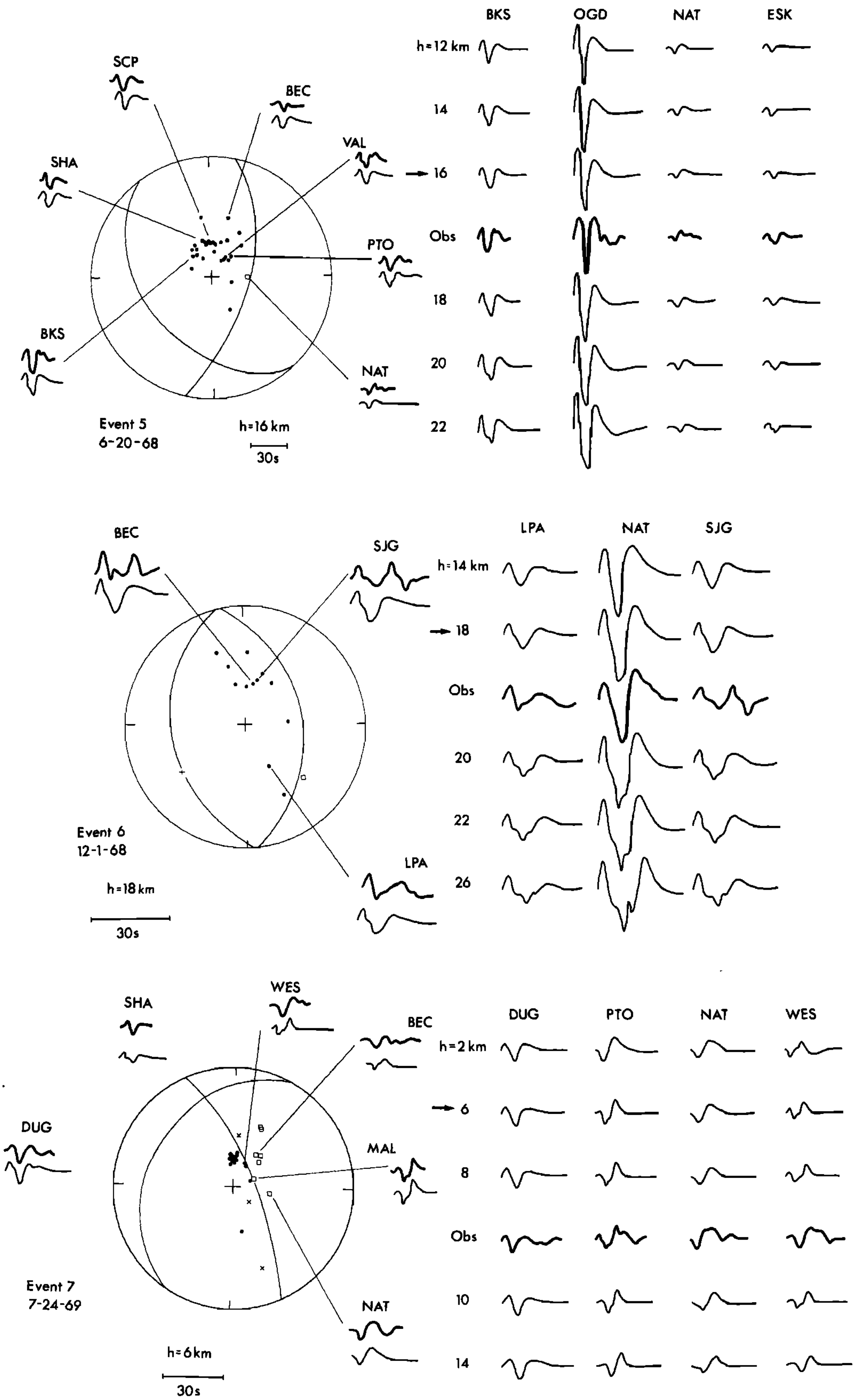


Fig. A1. (continued)



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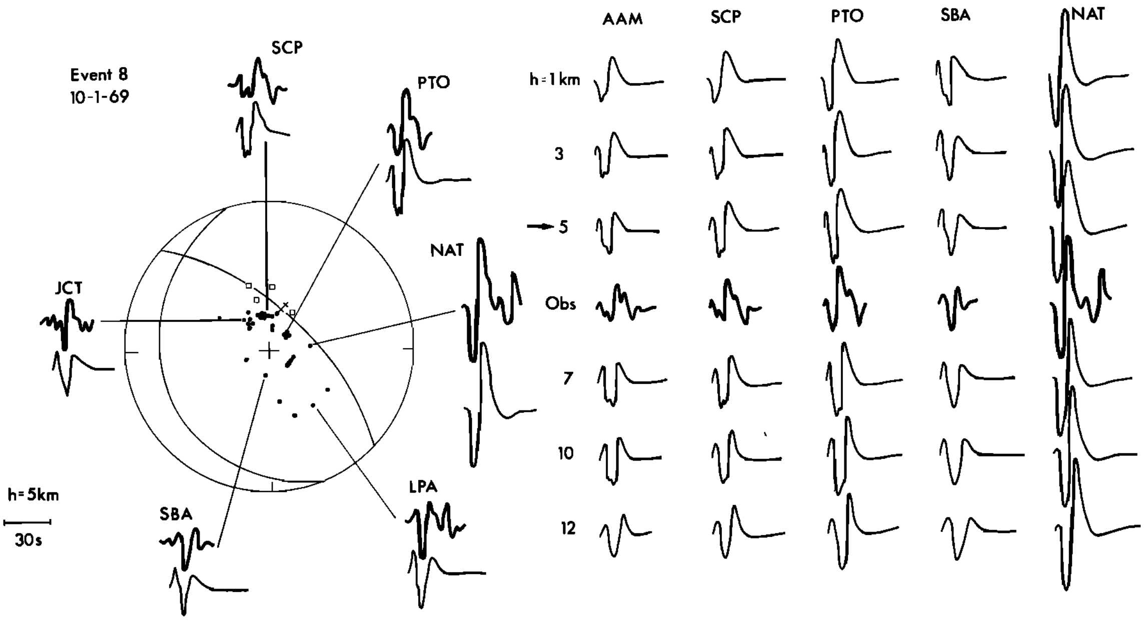
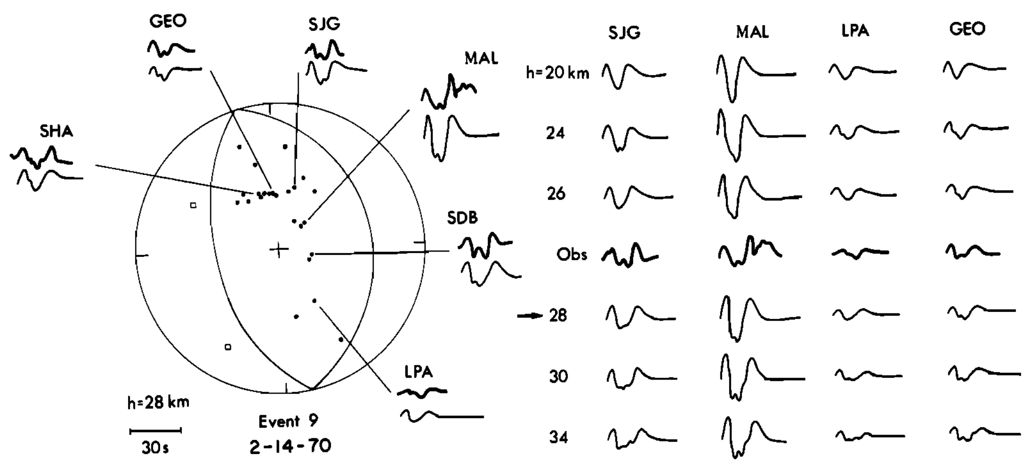




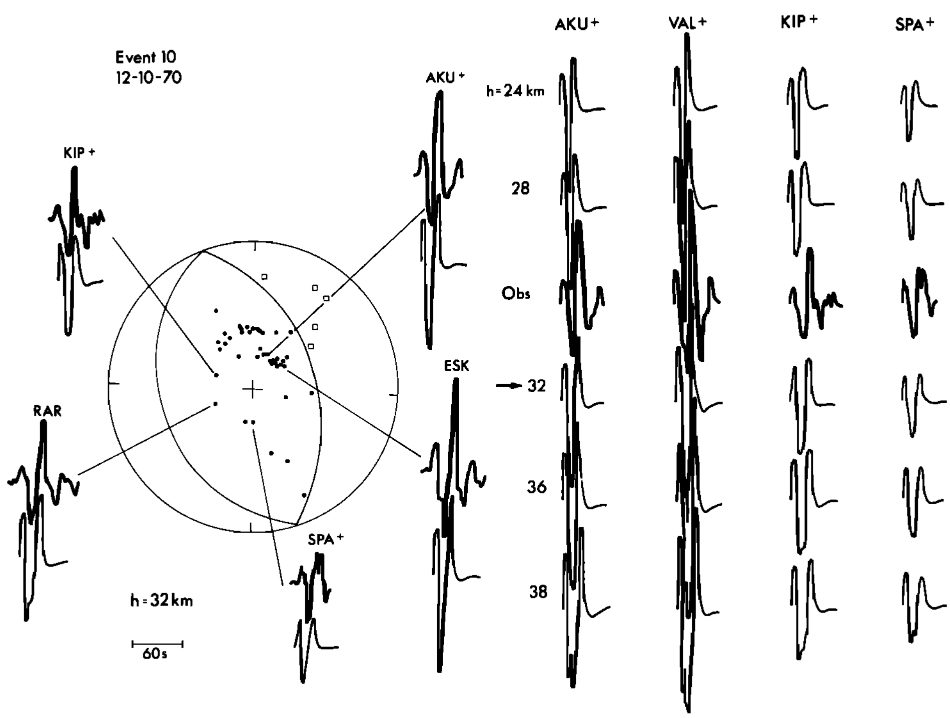
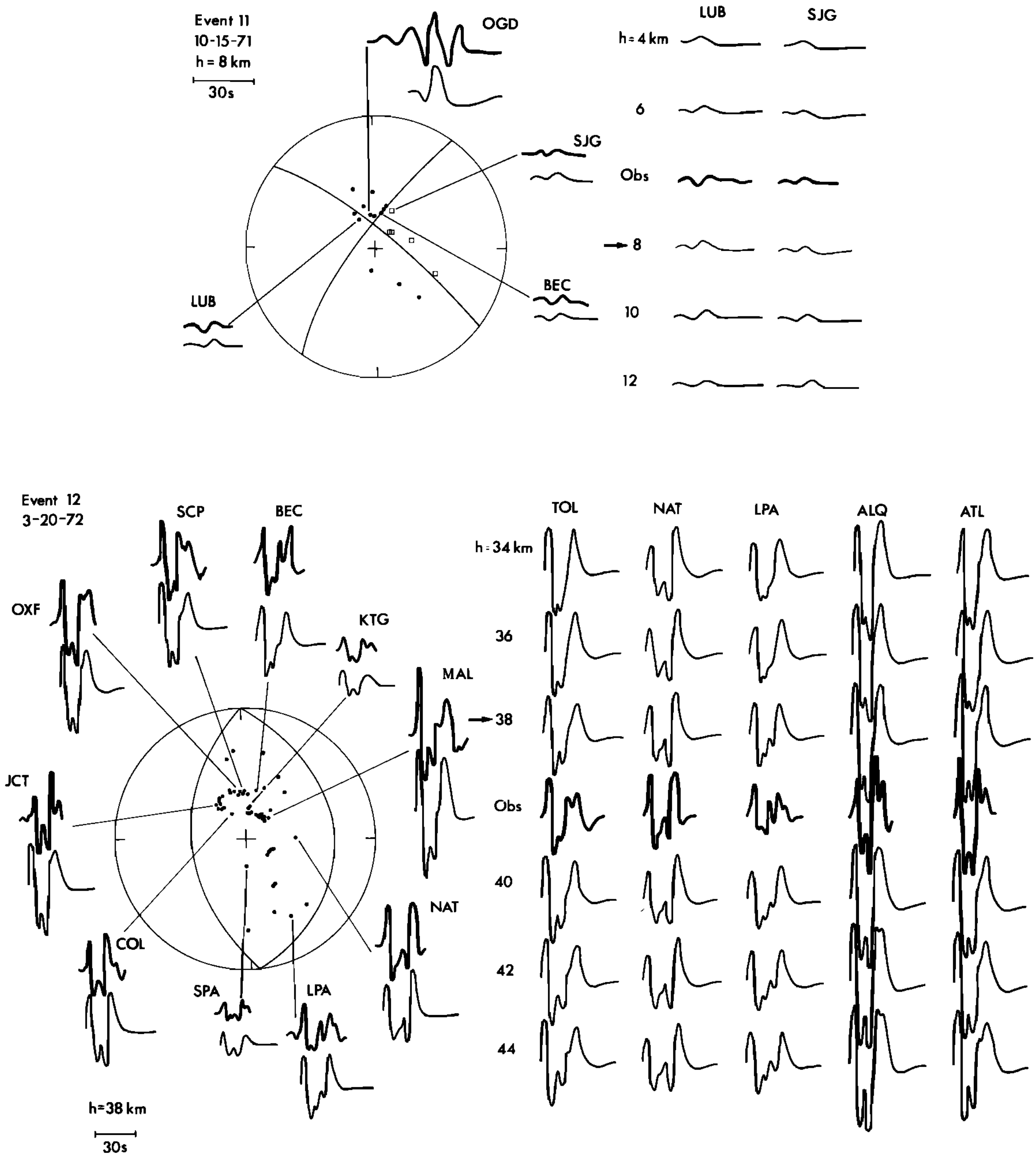


Fig. A1. (continued)



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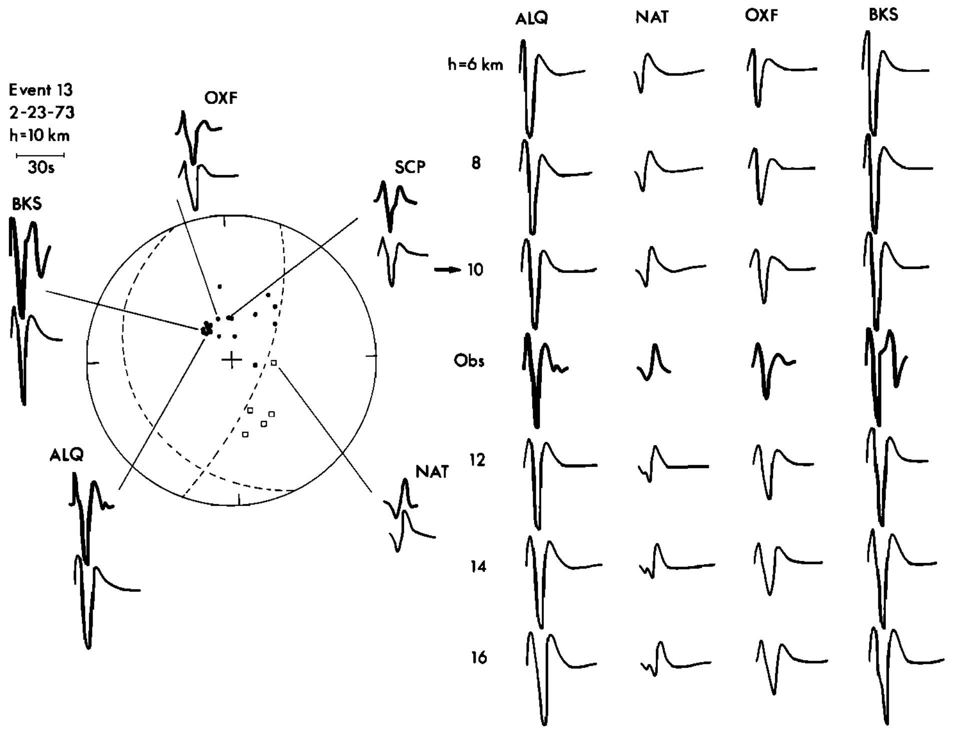

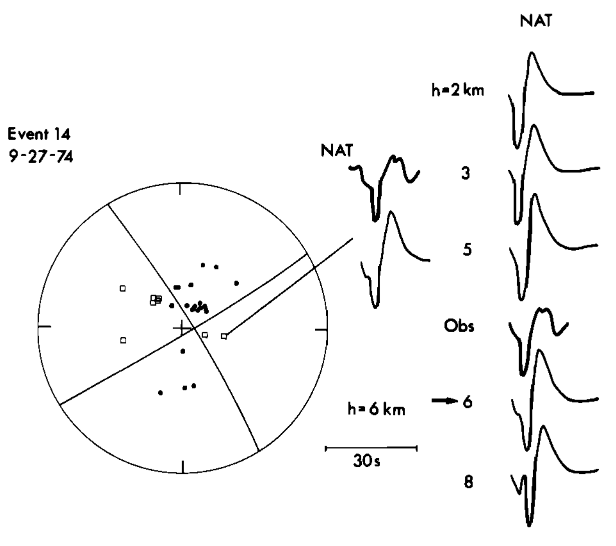

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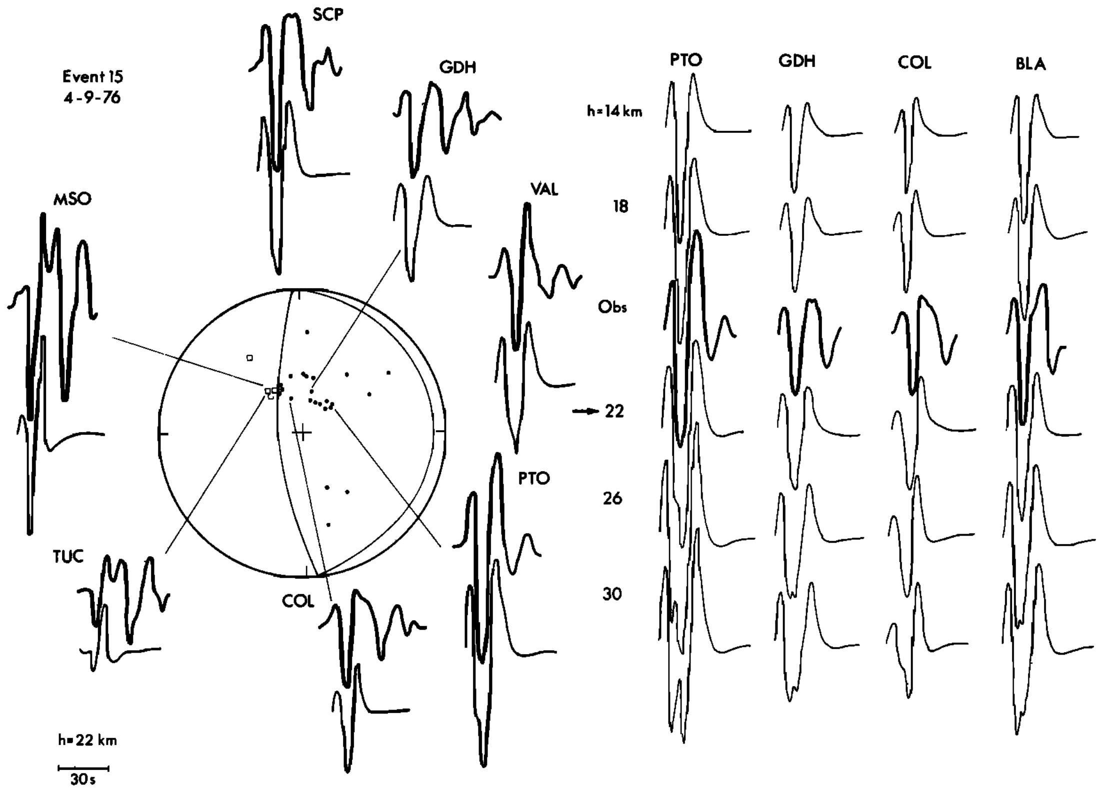

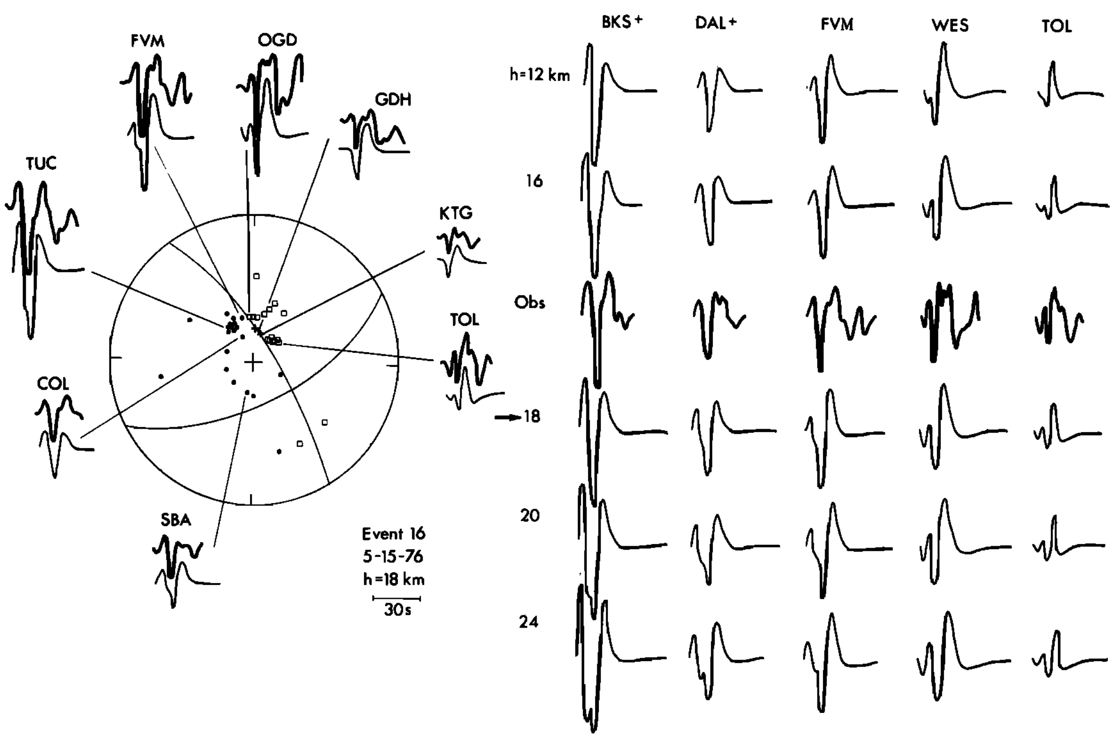

Fig. A1. (continued)

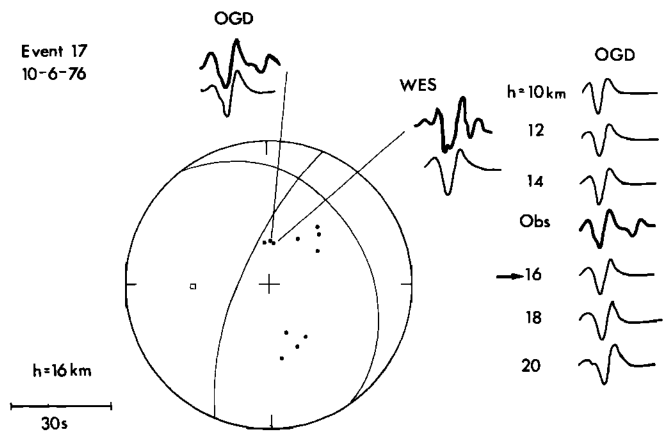

Fig. A1. (continued)

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