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The deep earthquakes of 1921–1922 in Northern Peru

Emile A. Okal *, Craig R. Bina

Department of Geological Sciences, Northwestern University, Evanston, IL 60208, USA

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Abstract

We present a detailed analysis of the 1921 and 1922 deep events in Northern Peru, which we identify as the only reliably located earthquakes in the vicinity of the great 1970 deep Colombian shock. All other reported seismicity in a three-dimensional radius of 550 km centered on the 1970 hypocenter is erroneous, and can be relocated to other seismic provinces. The 1921–1922 earthquakes relocate approximately 230 km to the south of the 1970 event, at depths of 630 km and 660 km, respectively. A combination of centroid moment tensor inversion and forward modeling using ray theory and reflectivity synthetics yields focal mechanisms indicating down-dip compression and seismic moments of 1.2×10^{27} dyn cm and 6×10^{27} dyn cm, respectively. The 1921, 1922 and 1970 events take place on a piece of subducted Farallon plate aged approximately 55 Ma, i.e., anomalously young to support deep seismicity. We suggest that this unusual seismic activity may be due to a local heterogeneity on the slab, whose nature, however, remains unknown.

1. Introduction

The deep Colombian earthquake of 31 July 1970 (1.46°S, 72.56°W; $h = 651$ km) is by far the largest event ever detected and measured at the bottom of any subduction zone (Gilbert and Dziewonski, 1975; Furumoto and Fukao, 1976; Okal and Geller, 1979). Its seismic moment, 2×10^{28} dyn cm, remains unchallenged 23 years later: the largest centroid moment tensor (CMT) solution computed deeper than 400 km is the recent 9 March 1994 earthquake south of Fiji, nearly an order of magnitude smaller (G. Ekström, personal communication, 1994). The exceptional size of the 1970 event can be grasped by noticing that it was felt from Mexico City to Buenos Aires. In terms of instrumentation, it may have been one of the most influential earthquakes ever, as it inspired the development of the IDA (International Deployment of Accelerometers) network (Gilbert, 1986; Agnew et al., 1986).

The Colombian event is also unique in that no foreshocks or aftershocks are documented, and no background seismicity appears in catalogs of modern earthquakes, the closest hypocenter being a 1963 shock at the northern end of the Peru–Brazil subduction segment, 535 km away (see Fig. 1). This distance is more typical of a regime of sparse seismicity in otherwise stable regions than of the general clustering of earthquakes observed at plate boundaries and within slabs (Frohlich and Davis, 1990).

* Corresponding author.

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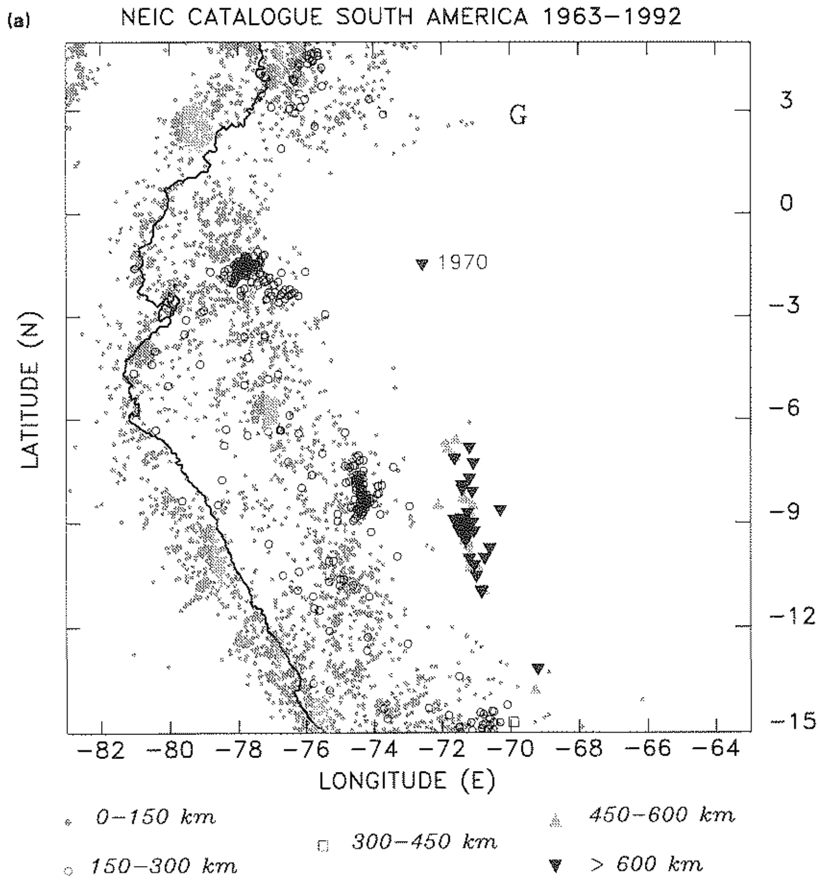

Fig. 1. (a) Regional seismicity in the vicinity of the 1970 epicenter, as extracted from the NEIC catalog. Various symbols are used to bin the events by depth. These events have not been relocated; however, the three erroneous earthquakes discussed in the Appendix have been suppressed. The symbol 'G' identifies the Guaviare seismic zone in Eastern Colombia. (b) Same dataset as in (a), plotted in cross-section along the north–south direction.

The occasional occurrence of earthquakes in subduction zones below the maximum depth of abundant seismicity is not unique; we refer, for example, to the New Zealand deep earthquakes of 1953, 1960 and 1975 (Adams, 1963; Adams and Ferris, 1976), complemented by a recent earthquake (14 September 1991), or to the two isolated intermediate events 100 km below the termination of abundant seismicity in the South Sandwich Islands (7 October 1974 (286 km); 1 January 1979 (288 km)). However, in all cases these were small events of magnitude $m_b = 5$. The only earthquake in recent times comparable with the 1970 Colombian earthquake is the Spanish event of 29 March 1954 ($h = 630$ km; $M_0 = 7 \times 10^{27}$ dyn cm (Chung and Kanamori, 1976)). However, the Spanish earthquake took place in an area featuring intense shallow seismicity. Furthermore, it was followed in 1973 and 1990 by two small shocks at similar depths ($m_b = 4.0$, $h = 634$ km, and $m_b = 4.1$, $h = 625$ km, respectively). The absence of comparable seismicity in the case of the Colombian event is not an artifact of detection capabilities, as earthquakes of magnitudes as small as 3.5 are routinely detected (and were in the 1970s) at the bottom of other segments of the South American slab, in areas which are not instrumented significantly better than Eastern Colombia.

On the other hand, catalogs of historical seis-

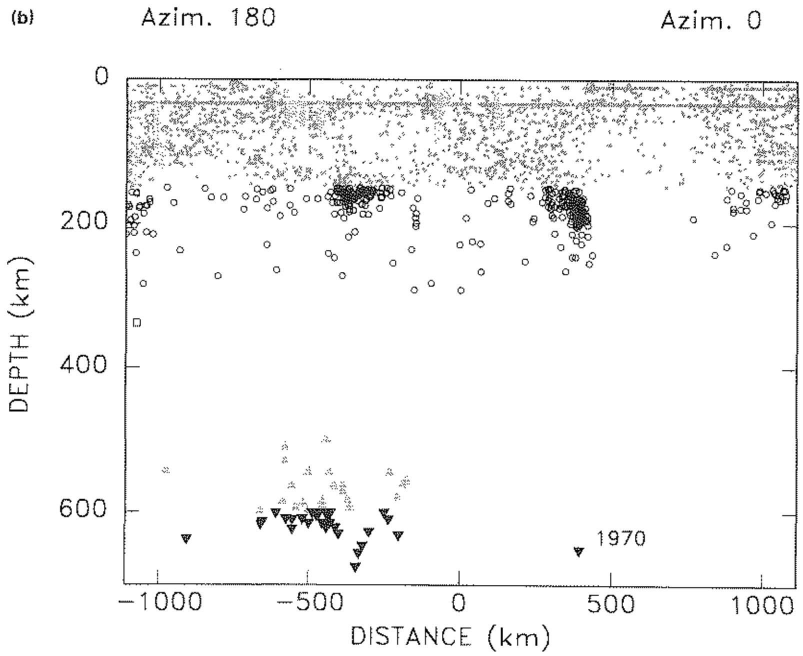

Fig. 1 (continued).

micity list two earthquakes in the vicinity of the 1970 event, on 18 December 1921 and 17 January 1922. The International Seismological Summary (ISS) locates the 1921 event at 2.5°S, 71°W and the 1922 event at 2S°S, 72°W. They propose depths of 540 km for the 1921 earthquake, and 475 km for the 1922 shock (although the ISS listing of the latter on the National Earthquake Information Center (NEIC) tape gives no depth information). Gutenberg and Richter (1954) placed both events at 2.5°S, 71°W and 650 km depth. These researchers assigned 'unified' magnitudes M of 7.9 (1921) and 7.6 (1922). As a result, the 1921–1922 earthquakes are prominently displayed on world seismicity maps (e.g., American Association of Petroleum Geologists, 1982), actually with more emphasis than the 1970 event ($m_b = 7.1$), an artifact of the disparity in the magnitude scales used.

The occurrence of these two earthquakes in the vicinity of the 1970 shock raises a number of important questions pertaining to the origin of seismicity and the mechanism of stress release at the bottom of the South American subduction zone. The mere fact that the 1921–1922 earthquakes occurred in an apparent doublet whereas the 1970 event was a single, isolated shock, points to a different pattern of activity. In this framework, we seek to answer at least partially the following questions:

(1) can the 1970 event be considered a spatial repeat of the 1921–1922 sequence? (i.e. are the locations similar?)

(2) Are the 1921, 1922 and 1970 shocks the only ones in the area, or are there historical hypocenters in their vicinity?

(3) Can the focal mechanisms of the 1921–1922 sequence be worked out, and how do they compare with that of 1970?

(4) Can an estimate of the seismic moment of the 1921–1922 shocks be obtained? Are those shocks truly larger than the 1970 earthquake, as suggested by seismicity maps?

(5) More generally, what is the relationship of the 1921, 1922 and 1970 events to the Andean subduction process, as it is defined further south?

2. Relocations and reassessment of nearby seismicity

The unusual character of the 1922 event elicited the early interest of a number of scientists. Seeking to explain the abundance of what we now describe as depth phases at a time when the existence of deep earthquakes remained a matter of speculation among a few visionaries, Byerlee (1924) invoked a complex sequence of three shocks, taking place within a few seconds of each other, in Ecuador, Brazil and Venezuela, respectively. Later, Inglada (1944) proposed a hypocentral location at 2.67°S, 71.73°W ($h = 602$ km) ('VIyO' in Fig. 2), based primarily on a detailed investigation of the Toledo records. Most of these early reports concentrated on the 1922 event, suggesting that it was the larger of the two, contrary to the 'Pasadena' magnitudes later assigned by Gutenberg and Richter.

2.1. Relocation of 1921–1922 events

We relocated the events of 18 December 1921 and 17 January 1922 using P and S arrival times listed in the ISS, and the techniques described by Wysession et al. (1991). These include an interactive iterative least-squares algorithm, offering the possibility of deleting erroneous stations, and of either constraining the depth or letting it float. In addition, we use a Monte Carlo technique to inject noise into the dataset of published arrival times, to explore the precision of our relocations.

In the case of the 1921 earthquake, a floating depth relocation using 20 P and S times converges on 4.17°S, 72.13°W at 655 km, with the standard deviation of residuals being $\sigma = 2.43$ s, a very good figure for an event that far back in time. However, the Wiechert records at Uppsala (see below) suggests a slightly shallower depth (630 km) on the basis of the time difference sS − S. We prefer to constrain the depth at this value, which results in a very similar epicenter, 4.11°S, 72.04°W, with no appreciable deterioration in the quality of the relocation ($\sigma = 2.56$ s).

In the case of the 1922 event, a similar strategy using a dataset of 39 ISS times yields a floating depth relocation at 3.76°S, 71.89°W ($h = 635$ km)

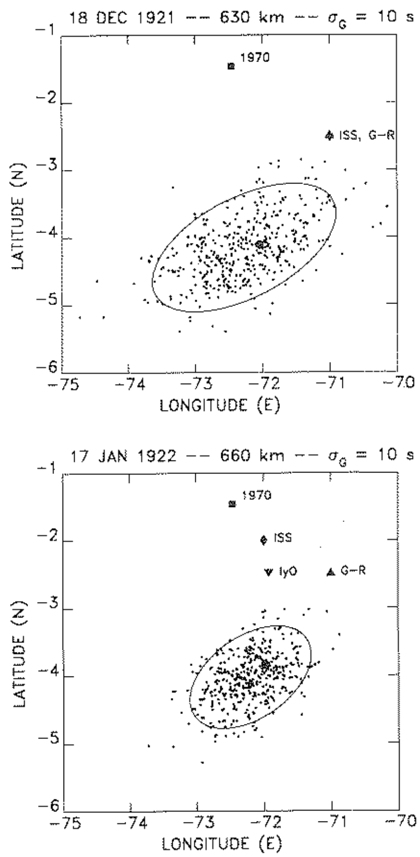

Fig. 2. Results of the relocation of the 1921 and 1922 events. In both cases, the large dot is the epicenter obtained with the depth constrained from the observation of sS − S at Uppsala; the small dots are the population of Monte Carlo epicenters obtained by injecting noise into the dataset (see Wysession et al. (1991) for details) with corresponding 95% confidence ellipse shown. Various symbols give the location of other estimates of the epicenters.

with $\sigma = 4.22$ s, whereas constraining the depth at 660 km, as suggested by sS − S at Uppsala, moves the epicenter SW a few kilometers to 3.84°S, 71.98°W, with no detectable change in $\sigma = 4.22$ s, indicating that the dataset of arrival times cannot resolve depth. Our preferred epicenter is approximately 160 km south of the location proposed by Inglada (1944).

For the purpose of running statistical tests using the Monte Carlo approach, we selected a standard deviation $\sigma_G = 10$ s for the Gaussian deviate errors added to the arrival times (this choice is in keeping with our experience in applying the Monte Carlo algorithm to historical earthquakes — see Wysession et al. (1991) for details). On the basis of the sS – S evidence from the Uppsala records, we held the depths fixed at 630 km and 660 km, respectively, in all tests. The results of the Monte Carlo experiments show that the two epicenters are indistinguishable, but significantly south of the 1970 earthquake (see Fig. 2). Longitudinal control is relatively poor, especially in the case of the 1921 shock, with acceptable epicenters extending as far east as 71°W. It is thus most likely that the 1921–1922 shocks took place under Peru, across the border from the 1970 Colombian event.

2.2. Nearby recent seismicity

In very general terms, the seismicity of Colombia, Ecuador and Northern Peru reflects the sub-

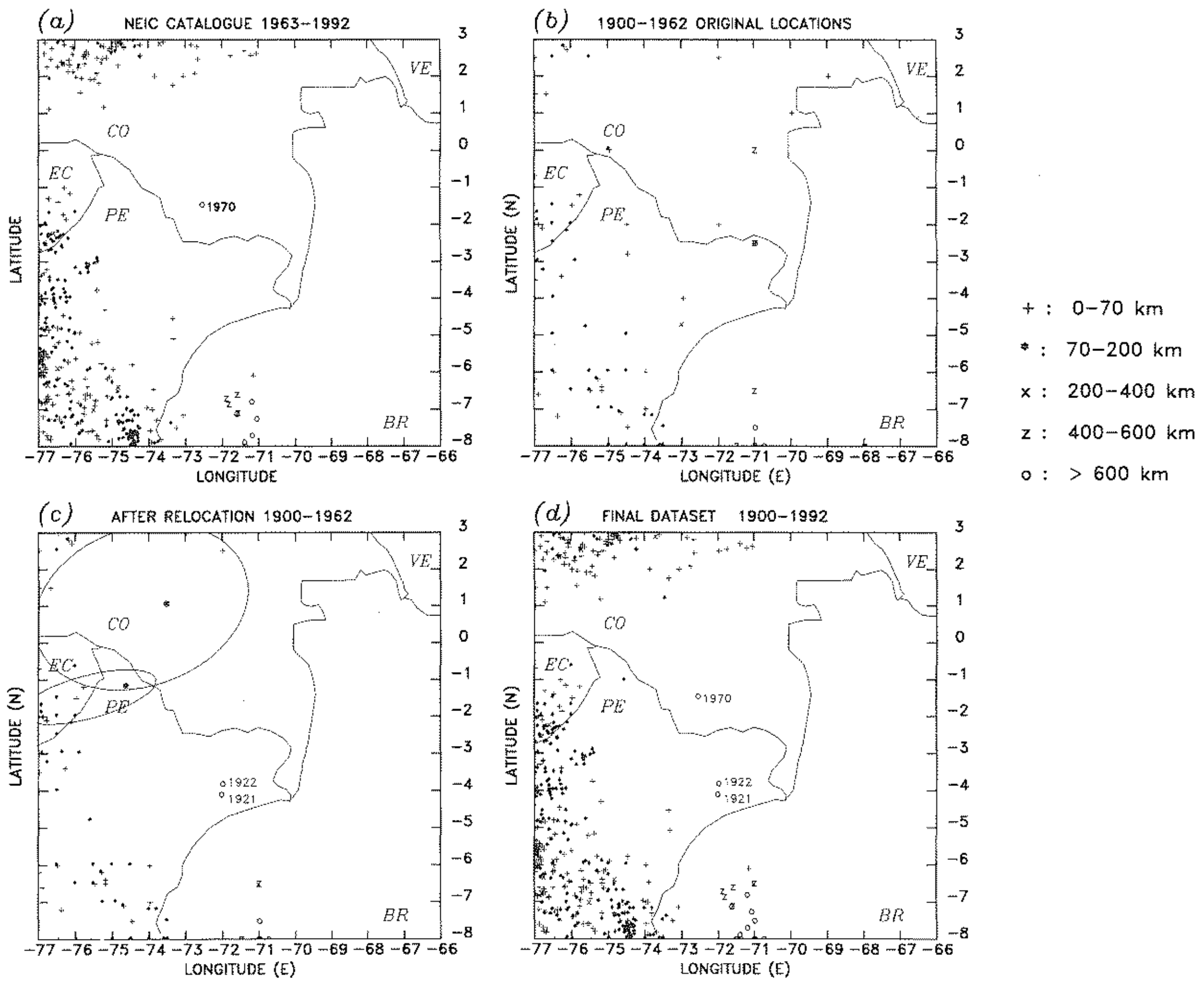

Fig. 3. (a) Original locations of regional seismicity in the vicinity of the 1970 Colombia event, as resulting for the NEIC catalog for the years 1963–1992. Political boundaries are shown by solid lines, and countries identified by two-letter codes. (b) Same as (a) for historical earthquakes (1900–1962). (c) Result of the relocation of historical earthquakes (note that all the seismicity in the area of the 1970 shock relocates to the seismic belts, with the exception of the two shocks in 1921 and 1922). Monte Carlo ellipses traced in the case of two apparent outliers show that they are not significantly detached from the seismic belts. (d) Final map of relocated epicenters, covering the full period 1900–1992. (See text for details).

duction process along the Colombia–Ecuadorian shoreline, as well as the Andean orogeny. The following characteristics are defined by modern (i.e. post-1963) hypocenters (see Figs. 1 and 3(a)):

(1) seismicity extends down the Wadati–Benioff plane to a maximum depth of 292 km (20 November 1964; $m_b = 3.7$); it is strongly clustered at two sites around 3°S, 78°W and 8°S, 75°W. Further down, except for a few obviously erroneous or unresolvable events (discussed in the Appendix), there is no recorded seismicity down to 500 km, where activity starts again under the Peru–Brazil border, south of 6.5°S, and remains fairly continuous with depth, down to 672 km (see Fig. 1(b)).

(2) In addition, the 1970 deep shock stands out alone at 651 km depth, under Eastern Colombia.

(3) A finger of active seismicity extends eastwards approximately 300 km into the Brazilian shield, between latitudes 2°N and 3°N, in the Guaviare province of Colombia. It reaches $m_b = 5.6$, $M_s = 5.8$ on 27 September 1974. The latter event contributed 150 stations to the US Geological Survey (USGS) and its location is beyond doubt. We refer to this line of active seismicity as the Guaviare seismic zone.

2.3. Nearby historical seismicity

In addition to the 1921–1922 earthquakes, historical seismicity (as compiled for the years 1900–1962 from the NEIC tape) shows a number of scattered epicenters in the general area of this study (see Fig. 3(b)). Because of the general absence of modern activity at these sites, we proceeded systematically to relocate (1) all earthquakes whose hypocenter is given in a zone of post-1963 quiescence, and (2) all earthquakes originally given a depth greater than 200 km. Details of the relocations are found in the Appendix. The resulting hypocenters are given in Fig. 3(c). It is immediately apparent that the general patterns outlined above for post-1963 seismicity also hold for historical seismicity. We show the Monte Carlo ellipses for two events whose relocated hypocenters (identified by dots in Fig. 3(c)) plot somewhat outside of the zone of modern seismicity: it is clear that these ellipses have a substantial intersection with these zones, and that the events are therefore not necessarily anomalous. The only remaining isolated events are the deep 1921 and 1922 shocks. In particular, the general trend running in a SW–NE direction in Fig. 3(b) coincides in azimuth with the major axis of many of our Monte Carlo ellipses; it simply reflects the lack of resolution of the corresponding datasets in this direction.

Thus, the 1921–1922 shocks are the only reliable hypocenters in the vicinity of the 1970 earthquake. After 1963, the NEIC catalog is certainly complete down to magnitude $m_b = 5.0$, and probably down to $m_b = 4.5$. For historical events, these thresholds would be $M = 6$ and $M = 5\ 3/4$, respectively (Gutenberg and Richter, 1954). Despite this unavoidable inhomogeneity in detection thresholds, the conclusion of this section is that the three shocks of 1921, 1922 and 1970 are exceptionally isolated at the bottom of the Andean subduction zone.

3. Seismological modeling

3.1. Data

In the present study, we use seismograms of the 1921, 1922 and 1970 earthquakes from the Uppsala Wiechert seismometer. This system (Kulhánek, 1987) has the advantage of featuring a relatively long pendulum period ($T = 10$ s), and of having been in uninterrupted operation ever since 1904. This allows the rather rare opportunity of a direct comparison of historical and modern earthquakes, as recorded on the same instrument. In addition, Uppsala (UPP) enjoys natural polarization (its back-azimuth to the 1970 epicenter being a near-perfect 269.6°), which allows direct visual interpretation of the original seismograms as radial and transverse components without the need to rotate them. Finally, original records of the yearly calibrations are available in the UPP bulletins, and thus provide optimal control on the precise instrument response at the time of the event. For these reasons, and even though we gathered a few seismograms at other stations (mostly written on shorter-period instruments), we elected to focus on the UPP records. The original smoked-paper records were directly digitized at Uppsala, and later processed to remove pen curvature, and interpolated to a constant time interval $\delta t = 1$ s. Fig. 4 compares records of all three earthquakes.

The most obvious feature in Fig. 4 is the difference in source complexity: whereas the waveshapes of the 1921 and 1922 earthquakes are relatively simple, particularly on the NS (SH) components, the 1970 event shows complexity, in agreement with the 60 s source duration described by many authors (e.g. Gilbert and Dziewonski, 1975), suggesting that the two historical events have a generally smaller size. Furthermore, an examination of the EW (SV) records shows striking differences in SKS/S amplitude ratios, these being substantially greater than unity (1921), about unity (1922) and substantially less than unity (1970), pointing to necessary differences in focal mechanism between the three events.

3.2. Focal mechanism and moment: 1922 event

In this section, we present the results of a one-station CMT inversion of the 1922 earthquake, based on the UPP records. As part of an

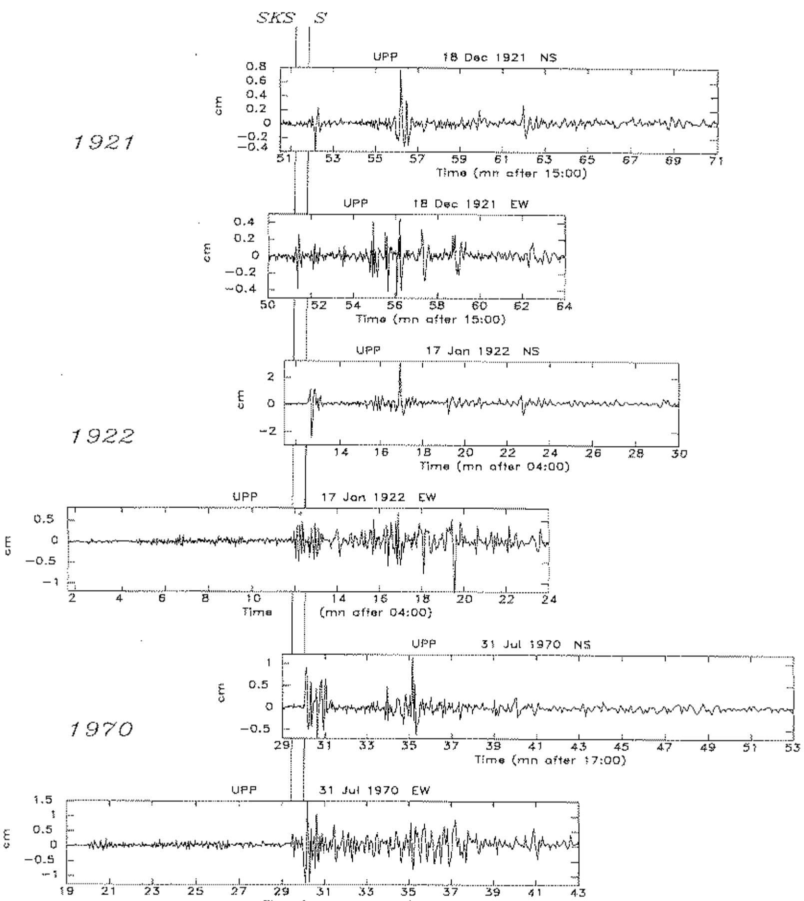

Fig. 4. Comparison of the 1921, 1922 and 1970 events, as recorded on the same Wiechert instrument at Uppsala. The time-scale is the same on all seismograms, but the vertical scales vary. The seismograms have been aligned to match the arrival times of S. Computed travel times for SKS are also shown. (Note the waveform complexity of the 1970 records, and the variation in SKS/S amplitudes. See text for details.)

independent project discussed more in detail in Huang et al. (1994), we have conducted a feasibility study of the application of the standard Harvard CMT algorithm (Dziewonski et al., 1983) to deep historical earthquakes. The possibility of using a dataset restricted to as little as one station was initially suggested by Buland and Gilbert (1976). In the case of deep earthquakes, the excitation of numerous overtone branches with significantly different kernels allows stable inversions for datasets as restricted as those consisting of two horizontal records at a single, long period (not necessarily broad-band) station, with obvious applications to large, deep, historical earthquakes (Huang et al., 1994). The 1922 event was one of 20 earthquakes selected for analysis as part of that study.

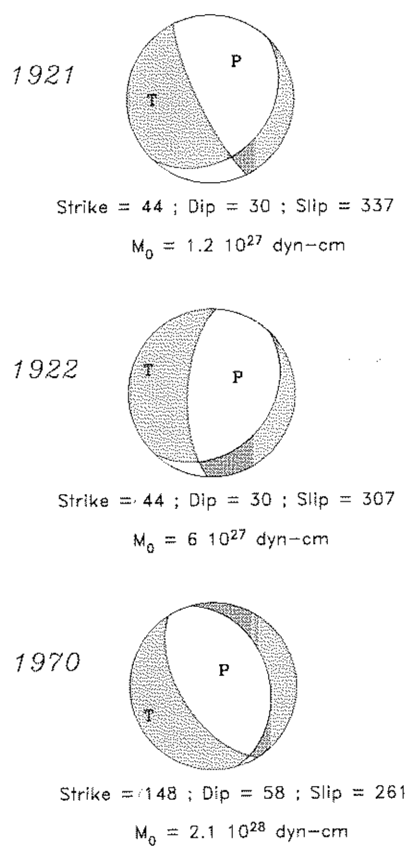

Fig. 5. Final preferred double-couples compared with the 1970 focal solution. The 1922 mechanism is the best double-couple resulting from the CMT inversion; the 1921 mechanism is derived from the former by forward-modeling the UPP record (see text for details).

Single-station CMT inversions must be conducted with a constrained epicenter, as they have no resolution in the direction perpendicular to the great-circle path involved. However, the inversion can and does resolve epicentral distance, and of course focal depth. In the case of the 1922 event, inversion of the UPP records converged acceptably only for a distance of approximately 91.5°, which requires moving the epicenter to the vicinity of 3°S, 71°W, at the extreme eastern end of the range of epicenters obtained from our Monte Carlo relocations, approximately 50 km south of Gutenberg and Richter's (1954) solution.

The results of the inversion are shown in Fig. 5 in the form of the best-fitting double-couple. The full five-dimensional deviatoric solution features a minor double-couple contributing 16% of the total seismic moment. The focal mechanism of the best double-couple ($\phi = 44°$, $\delta = 30°$; $\lambda = 307°$) clearly expresses the down-dip compression characteristic of large earthquakes at the bottom of subduction zones; in particular, its P-axis is only 16° away from that of the nearby 1970 shock. The tension axis, on the other hand, is significantly rotated from that of the 1970 shock, a simple expression of its generally variable character for deep events at the bottom of subduction zones.

3.3. *Forward modeling: synthetic seismograms of the 1922 event*

We complemented the CMT inversion by forward-modeling the UPP records using both a reflectivity code (Kennett, 1983; Clarke, 1993) and a simple generalized ray-theory approach, following Okal (1992), itself based on a code originally written by Stein and Wiens (1986). This allowed us to fine-tune the source time function at frequencies higher than used in the CMT algorithm (0.022 Hz). Fig. 6(a) compares the original data with ray-theory synthetics; although the CMT inversion gives a half-duration of 9 s, we find that the waveforms are best fitted by quasi-triangular but shorter time functions, lasting only 15 s; longer, or more rectangular time functions would give rise to broader S and sS waveforms. The resulting value of the best-fitting moment is also in better agreement with the inversion. Fig. 6(b) similarly compares the UPP records with reflectivity synthetics build for a source with rise, top and fall times of 6 s, 3 s and 6 s, respectively, and for a moment of 6.7×10^{27} dyn cm. The synthetics give an excellent fit to the relative

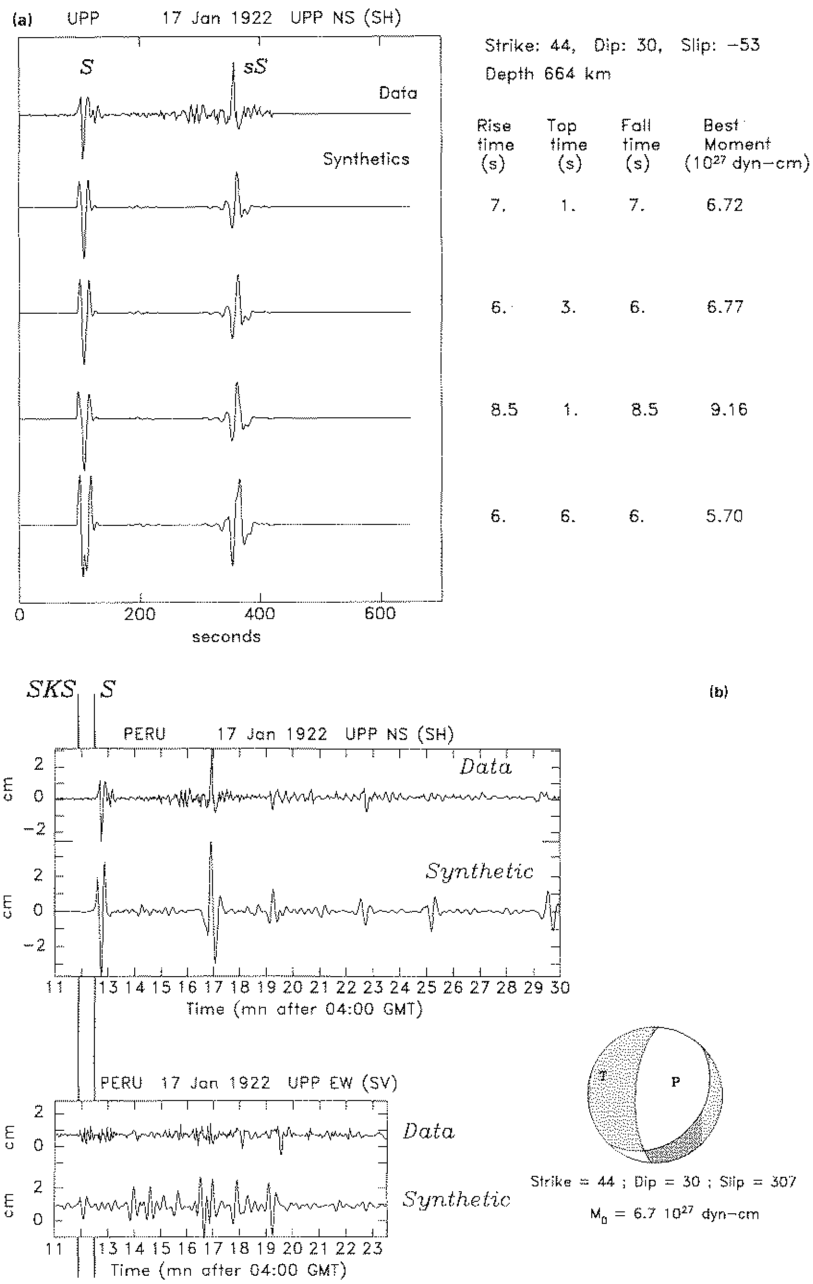

Fig. 6. (a) SH-Waveform modeling of the 1922 record at UPP. This procedure constrains the source time function. (b) Reflectivity synthetic seismograms computed at UPP for the final model of the 1922 source, and compared with the observed records.

amplitudes $(sS/S)_{SH}$ and S_{SV}/S_{SH}; they also give a reasonable fit to SKS/S_{SV}. They would, however, require a somewhat lower value of the moment; the combination of the CMT inversion and the two sets of body-wave synthetics suggest 6×10^{27} dyn cm, with a precision of 30%.

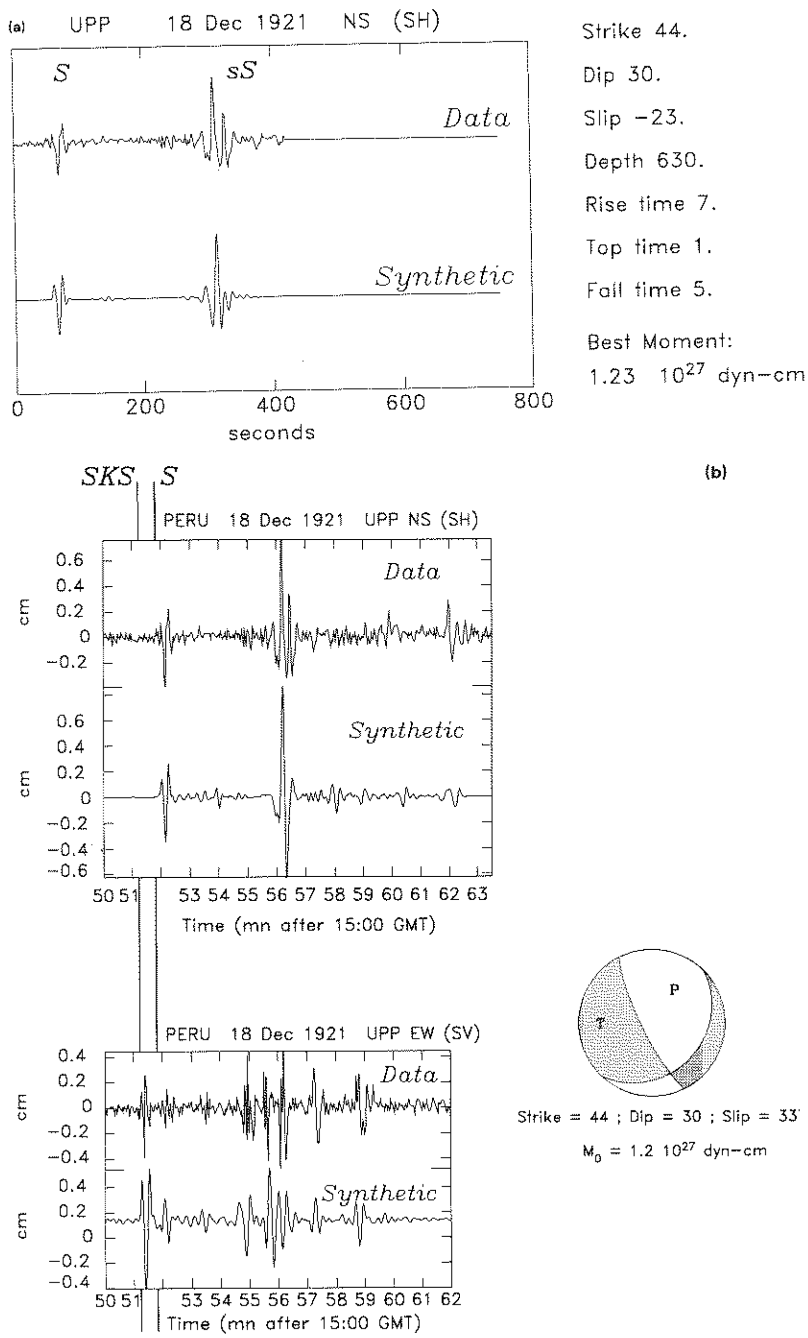

Fig. 7. (a) SH-Waveform modeling of the 1921 record at UPP. (b) Reflectivity synthetic seismograms computed at UPP for the final model of the 1921 source, and compared with the observed records.

This preferred moment makes it one of the largest events ever modeled at such depths. Although it remains three times smaller than the Colombian shock, it is comparable in size with the 1954 deep Spanish earthquake, more than three times larger than the 1982 Banda Sea earthquake, the largest deep event in the published CMT catalog, and at least twice as large as the recent 1994 deep Fiji event.

3.4. Focal mechanism and moment: 1921 event

In the case of the 1921 event, the CMT solution failed to converge, i.e. the inversion did not result in a satisfactory goodness of fit. However, on the basis of ray theory and reflectivity synthetics, we propose a tentative mechanism, adapted from the 1922 focal solution by keeping one fault plane, but increasing the strike-slip component. The resulting focal solution, ($\phi = 44°$; $\delta = 30°$; $\lambda = 337°$) requires a seismic moment of 1.2×10^{27} dyn cm for a source duration of 13 s. Figs. 7(a) and 7(b) compare the UPP data with the ray theory and reflectivity synthetics, respectively. Both techniques model the $(sS/S)_{SH}$ ratio very well, and the reflectivity synthetics also give a good fit to SKS/S_{SV}.

The mechanism we tentatively propose for the 1921 event shares its general nature (downward compression) with the 1922 source, but the P-axis dips only 49°, and stands 26° away from that of the 1970 event. A comparison of the two moments, 1.2×10^{27} dyn cm (1921) and 6×10^{27} dyn cm (1922), ranks the second earthquake as the larger one. This was clearly indicated by the relative amplitude of the dominant (NS) components at UPP, but is in contrast to the magnitudes assigned by Gutenberg and Richter (1954). Additionally, both events are clearly smaller than the 1970 event. In particular, the moment of the 1922 event is large, but not exceptional, as deep shocks of comparable size have been documented, in particular further south along the Andean Wadati–Benioff zone (Furumoto and Fukao, 1976) albeit in spatial clusters of abundant seismicity.

We obtained source durations of 13 s (1921) and 15 s (1922) from body-wave modeling. In the framework of Vidale and Houston (1993), these would correspond to a 'scaled duration' of 5.7 s and 3.8 s, respectively. Even though these authors' approach is different from ours, and direct comparison may not be warranted, it is interesting to note that the 1921–1922 earthquakes fit the general pattern defined by Vidale and Houston for events in their depth range.

4. Discussion

We have shown that the three events of 1921, 1922 and 1970 make up an exceptional seismic location in an otherwise inactive portion of the South American slab. Any explanation of this activity will have to account for (1) the absence of deep seismicity ($h \geq 300$ km) north of 6.5°S; (2) the isolated occurrence of the three shocks; (3) their large size — this last being particularly impressive as their cumulative moment reaches about 3×10^{28} dyn cm. On the other hand, the mechanism of the events (principally down-dip compression) fits the mold of seismicity at the bottom of slabs, and does not appear anomalous. Similarly, their scaled durations, computed in the formalism of Vidale and Houston (1993) do not make them exceptional. Finally, it is worth noting that the 1921, 1922 and 1970 events, when combined with the Peru–Brazil region between 7°S and 12°S and the Peru–Bolivia focus at 13°S, make up a linear string of very deep seismicity which can be regrouped in four clusters (Southern Colombia, Northern Peru, Peru–Brazil and Peru–Bolivia). However, the patterns of occurrence are very different at the two southern clusters, with the Peru–Brazil location featuring abundant seismicity scattered both in time and space (along the vertical and north–south directions), and the Peru–Bolivia location having exhibited some modest background of deep seismicity and a significant earthquake in 1994 at the same hypocenter as the large 1963 event. Thus, the various clusters probably correspond to different seismogenic regimes.

In general, the absence of deep seismicity in slabs has been attributed by Kirby et al. (1991) to the thermal characteristics of the descending slab, with the product of the age of the sinking material by its convergence rate controlling the development of deep seismicity: only those slabs initially old enough and sinking fast enough will create a large enough thermal anomaly to sustain earthquakes below 300 km. In this respect, the South American slab would not be expected to feature deep seismicity, as the subducted lithosphere is too young. However, Engebretson and Kirby (1992) have explained deep South American earthquakes by proposing that an age discontinuity of as much as 60 Ma exists in the subducted lithosphere under the Andes. In their model, they noted that the Farallon (proto-East Pacific Ridge) spreading center initially developed around Chron 34 (82 Ma), by cutting into lithosphere generated at the Phoenix spreading center, and aged 140 Ma. Phoenix lineated lithosphere is documented in the southwestern Pacific Basin, south of the Manihiki Plateau, and west of the Pandora Escarpment (Engebretson et al., 1991; Mammerickx, 1992). The South American slab, the conjugate of that basin with respect to the Farallon spreading center, should then feature a deep plug of Phoenix-generated material, separated from the rest of the slab by an age discontinuity of the order of 60 m.y. There still remains the problem that deep seismicity is not continuous along the South American subduction zone, but rather there exist significant gaps in deep seismicity from 11°S to 13°S, from 14°S to 16.5°S, and from 17°S to 19°S.

In this very general framework we seek to explain both the termination of abundant and relatively regular seismicity north of 6.5°S, and

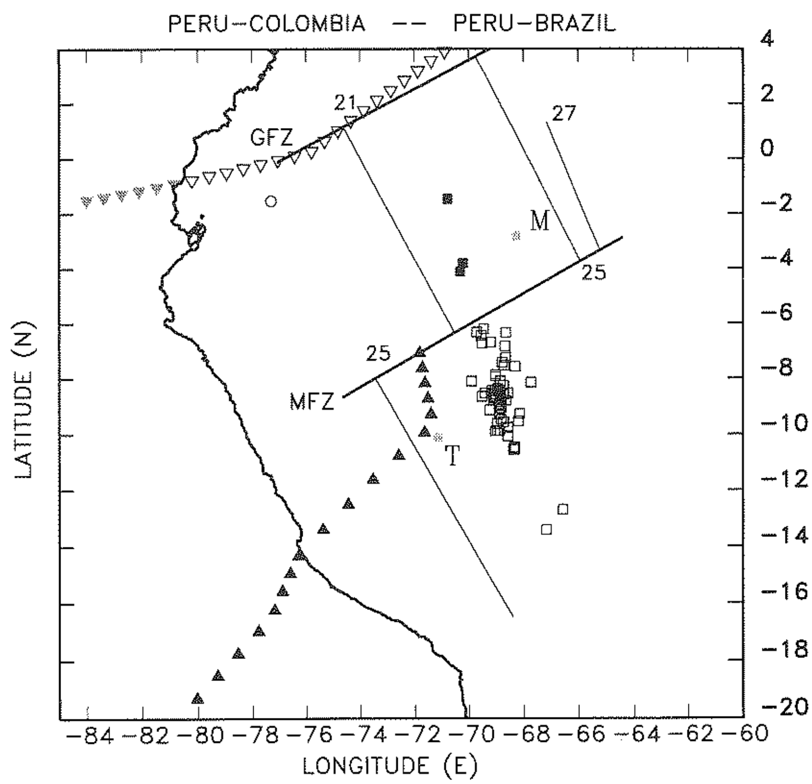

Fig. 8. Reconstruction of the relative positions of the hotspot tracks and of the seismic locations on the Nazca plate. This figure is a map of the Farallon–Nazca plate, as it would cover South America, were it not subducting. The solid lines are the Marquesas–Mendaña and Galápagos–Grijalva fracture zone systems, together with selected magnetic anomalies. The squares are the location of deep seismicity, rotated back onto the surface of the Earth (□, Peru–Brazil cluster; ■, Peru–Colombia events). ○, Approximate location of the strong cluster of intermediate seismicity under Central Ecuador. The triangles are the loci of the hotspot tracks (upward pointing: Nazca; downward pointing: Galápagos). The small dots labeled M and T indicate (for reference only) the rotated positions of the islands of Hiva-Oa (Marquesas; M) and Tahiti (T).

the presence of the three anomalous deep Peru–Colombia shocks. There are several ways of addressing the problem.

4.1. Old Phoenix lithosphere?

We must first assess the possibility that the deep Peru–Colombia shocks occur in a piece of Phoenix-generated lithosphere, along the lines of Engebretson and Kirby (1992). For this purpose, we map the actual deep seismic hypocenters back onto the Pacific plate, and examine whether they fall on Phoenix material in the model of Engebretson et al. (1991). This procedure involves several steps. We first rotate the subducted slab onto the surface of the Earth by computing the total length along the slab down to a particular hypocenter, and unfolding it back on the surface of the Earth along a great circle perpendicular to the local hinge of the subduction, i.e., to the plate boundary; the resulting epicenters are shown as solid (Peru–Colombia) and open (Peru–Brazil) squares in Fig. 8. We then rotate the corresponding points back onto the Pacific plate using the Cenozoic plate motions of Gordon and Jurdy (1986). In practice, because we do not know the age patterns on the subducted Nazca plate (before 25 Ma, this area was part of the Farallon plate, to simplify the language, we occasionally refer to the downgoing slab as a piece of 'Nazca' plate, even though we mean 'Farallon–Nazca plate'), we proceed backwards by mapping Pacific plate locations of known age back onto the Nazca plate, as it would lie over South America, had it not subducted. We use the Pacific floor ages of Acton and Petronotis (1992). Fig. 8 shows the rotated anomalies 21, 25 and 27, delineated by the traces of the Mendaña (rotated Marquesas) and Grijalva (rotated Galápagos) Fracture Zones (hereafter MFZ and GFZ, respectively). For reference, the rotated positions of Tahiti (T) and Hiva Oa (M) are also given, even though these young islands were never involved with Nazca plate ocean floor. We estimate that these rotations are probably accurate within 400 km.

The bottom line from Fig. 8 is that the part of the Nazca plate bearing the deep Peru–Colombia hypocenters maps onto Pacific lithosphere in the vicinity of the present-day Marquesas Islands, where Farallon magnetic lineations are abundant; Mammerickx (1992) has documented the ENE-trending Galápagos Fracture Zone all the way to the Manihiki Plateau, and the MFZ to longitude 167°W. Thus, it cannot be a Phoenix remnant and we can rule out the 'old Phoenix plug' interpretation of the deep Peru–Colombia shocks. Moreover, our rotation suggests that the age of the portion of slab involved in the Peru–Colombia cluster is at most 55 Ma, and thus much too young to be seismogenic, in the framework of Kirby et al. (1991).

It is tempting to make a similar argument for the deep Peru–Brazil events immediately to the south (open squares in Fig. 8): they map approximately 300–600 km west of Tahiti. A potential problem arises, as the ages of that part of Pacific lithosphere are poorly if at all known, owing to the overprinting of the magnetic record by the Society and Cook hotspots. In other words, the northern prolongation of the Pandora Escarpment, recognized south of the Austral Fracture Zone as the line of age discontinuity between Phoenix-and Farallon-generated lithosphere (Mammerickx, 1992), is unmapped, and thus the origin and age of the lithosphere bearing the deep Peru–Brazil hypocenters cannot be ascertained. However, for this lithosphere to be Phoenix generated would require a 400 km right lateral offset in the Farallon Ridge at, or immediately north of the Austral Fracture Zone, as it initially broke into Phoenix material; although such an offset is documented further east in the vicinity of the Southern Tuamotu Islands, the age gap across the Austral Fracture Zone decreases westwards, and is hardly documented along a line Tahiti–Raivavae (Acton and Petronotis, 1992).

4.2. 'Detached' events at the forefront of the slab?

A second possibility would be to consider the deep Peru–Colombia shocks as 'detached' events, occurring a few hundred kilometers continentwards of the slab, in the manner of the 1989 Paraguay, 1990 Sakhalin or 1982 Izu-Bonin events. This interpretation would not explain the deep Peru–Colombia activity, as a consistent explanation of such earthquakes has yet to be given (see Lundgren and Giardini, 1994 for a review); it would simply make them less singular. In the absence of a well-developed Wadati–Benioff zone in the area of the Peru–Colombia cluster, it is of course impossible to picture the location of the shocks relative to the slab itself, but Fig. 9 does not argue for such a scenario, as the deep shocks are reasonably aligned with two clusters of events, identifying the slab at lesser depths, this alignment involving only a minor amount of warping in the interpolated slab. The size of the Peru–Colombia events does not affect the argument, as both small and large events are found in such 'detached' locations (Lundgren and Giardini, 1994).

4.3. Activity on the track of one or more major hotspots

In this section, we explore the possibility that the deep Peru–Colombia and/or Peru–Brazil seismicity occurs along the remnants of the Nazca Ridge, the Carnegie Ridge, or both, borne by the Nazca plate lithosphere as it sinks into the mantle.

The suggestion that oceanic structures such as island chains and plateaux have a profound effect on the assimilation of lithosphere as it is being subducted is not new. Pilger (1981) has observed that areas where significant structures, such as the Juan Fernandez chain and the Nazca Ridge, subduct or have subducted (27–33°S and 6–15°S, respectively), are characterized by lower-dipping zones of intermediate seismicity, as well as by a general absence of present-day arc volcanism. More recently, Kirby and Engdahl (1993) have reported a significant correlation between the level of seismic activity at intermediate depths along the South American slab and the presence of structural features as inferred by the corresponding offshore bathymetry. However, these authors have been unable to trace this effect below a depth of 325 km. It should be noted that

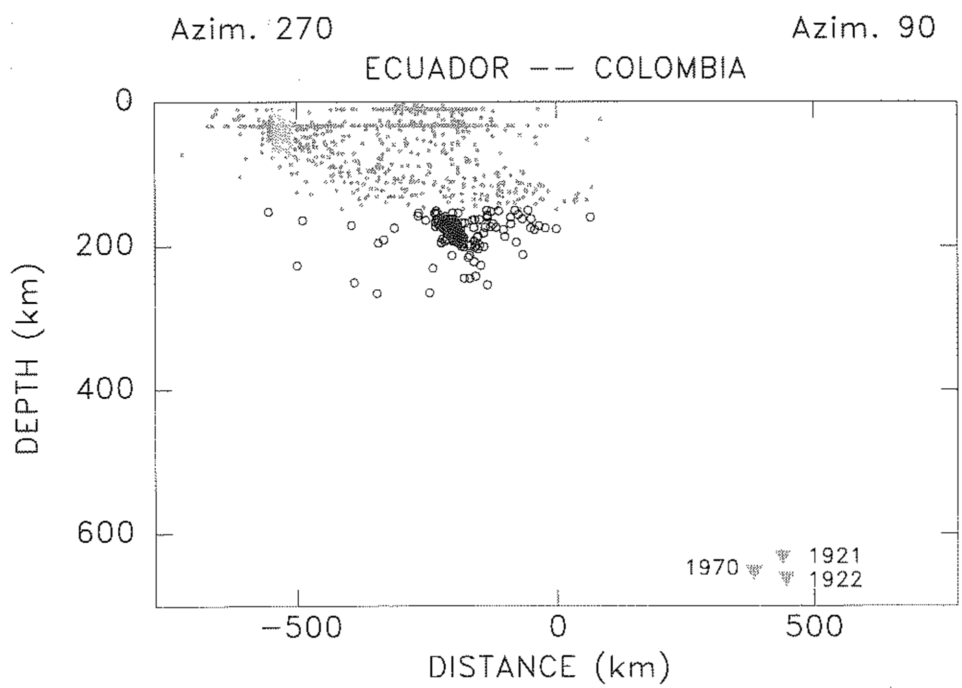

Fig. 9. East–west cross-section of seismicity dataset used in Fig. 1(a) between latitudes 4.5°S and 0.5°N. Although obviously incomplete, this figure does not suggest that the deep Peru–Colombia events are anomalously detached in front of the subducting slab.

Figs. 1 and 9 suggest that one such intermediate-depth cluster exists at 180–200 km under central Ecuador, directly updip from the 1970 event. Finally, Pilger (1981) has commented in very general terms on the variation of the morphology and activity of the Benioff zone in areas where large structures subduct.

In Fig. 8, we plot as solid triangles reconstructions of the presumed position of the Nazca ridge hotspot on the Nazca plate. These were obtained by rotating the Northern Tuamotu chain using the poles of Gordon and Jurdy (1986), and assuming that the Northern Tuamotus and the Nazca Ridge were formed on-ridge, and are thus coeval with the ocean floor (Okal and Cazenave, 1985; Talandier and Okal, 1987; Woods and Okal, 1994). In this respect, we interpret the Northern Tuamotu Plateau as conjugate of the Nazca Ridge, and use a sea-floor age of Chron 25 (60 Ma) in the vicinity of Rangiroa (Schlanger et al., 1976); the individual islands (e.g. Rangiroa itself) represent a later overprinting of the structure, by an off-ridge source contained entirely in the Pacific plate. Gordon and Jurdy's (1986) model extends back to Chron 27 (64 Ma), which also corresponds to the intersection of the Northern Tuamotu Ridge with the MFZ. Also, the large offset at the MFZ on the Pacific plate (about 1000 km or 10 Ma in the Eocene) implies that the hotspot cannot be tracked on the Nazca plate further back in time, when the age of the corresponding plate suddenly becomes 10 Ma younger. Before 64 Ma, the hotspot was off-ridge, entirely in the Pacific plate, and it came in contact with the ridge around Chron 27, owing to the presence of the MFZ. For that reason, we stop the formal reconstruction at 64 Ma.

Our reconstruction comes within 200 km of the deep Peru–Brazil seismic segment. This distance is comparable with the width of the Tuamotu Plateau and of the Nazca Ridge in its exposed part, and is at any rate less than the estimated accuracy (400 km) of the reconstruction. We speculate that the activity in the Peru–Brazil deep seismic zone may be related to the presence of Nazca Ridge material on the subducting slab. We emphasize that this assumption would be incompatible with simultaneously interpreting that portion of the slab as Phoenix generated, as the conjugate location of the hotspot (the Northern Tuamotus) is on a portion of the Pacific plate well documented to have been Farallon generated. It should be noted also that the northern termination of the Peru–Brazil deep seismic cluster corresponds remarkably well to the trace of the MFZ on the Farallon plate.

It is more difficult to track the Carnegie Ridge onto subducted Nazca lithosphere. Available models (Hey, 1977; Minster and Jordan, 1978; Gordon and Jurdy, 1986) all fail to reproduce the overall trend of the Carnegie Ridge (N95°E) on the Nazca plate and give it a more northerly position. Only the massive segment of the Carnegie Ridge east of 84°W has its azimuth properly modeled. This would be consistent with the idea that only the eastern part of the Carnegie Ridge may be the direct expression of the hotspot, before it moved under the Cocos plate at about 12 Ma (Lonsdale and Klitgord, 1978). The western part of the Carnegie Ridge would then result from lateral leaking or piping into the ridge, as suggested, for example, by Morgan (1978) and Schilling et al. (1982). In this respect, it is legitimate to use Gordon and Jurdy's (1986) rotations to map the previous path of the hotspot on the plate only before 12 Ma. In addition, Lonsdale and Klitgord (1978) have suggested that the hotspot became active only around 22 Ma (although it is not clear that hotspots can be turned on or off over very short periods of time — we prefer to invoke a temporal variation in activity possibly linked to rising blobs (Schilling, 1973, 1985), or controlled by structures in the overriding lithosphere (McNutt et al., 1989)). Assuming, nevertheless, that the hotspot was indeed present on the Nazca plate during the Oligocene and Eocene, and proceeding to rotate it from its position at 12 Ma, at the western end of the massive part of the Carnegie Ridge (84°W), we find that it approaches the Central Ecuador cluster of intermediate seismicity but misses the Peru–Colombia deep seismic zone by 600 km (inverted open triangles in Fig. 8).

In conclusion, it is probable that the deep Peru–Brazil cluster of seismicity takes place on the track of the Nazca hotspot within the accuracy (400 km) of the reconstructions. It is much more difficult to associate the Peru–Colombia shocks with this hotspot, which is not expected to have imprinted the Farallon plate north of the MFZ or with the Galápagos hotspot, whose wake is too far north.

The mechanism by which seismicity could be enhanced at the locus of a past center of volcanism on the plate remains speculative. A tentative explanation is suggested by the absence of present-day arc volcanism at locations where hotspot tracks (the so-called 'aseismic ridges') are subducted (Pilger, 1981). Extensive hydrothermal circulation near spreading ridges (Williams et al., 1974) in regions of minimal sedimentary cover (Davis and Lister, 1977) results in significant hydration of the oceanic crust. Dehydration of oceanic crust and accumulated sediments during subduction is thought to contribute fluids to the overlying mantle wedge, resulting in partial melting and consequent arc volcanism. The reheating of the oceanic crust associated with hotspot volcanism may result in partial dehydration of the crust before subduction. Subsequent subduction of such partially dehydrated crust would liberate less fluid, inducing less partial melting in the overlying mantle wedge and thus yielding less arc volcanism. Furthermore, if such an anomalously dehydrated state is maintained to the maximum depths of seismicity, then one might invoke the observed correlation between low water content and greater metastable persistence of low-pressure phases (Young et al., 1993), to argue for greater retardation of any putative seismogenic phase transformations and consequent anomalous seismicity at depth. We reiterate, however, that this tentative interpretation is proposed as a speculative model.

5. Conclusions

The 1921–1922 deep earthquakes in Northern Peru are, together with the 1970 Colombian deep shock, the only earthquakes known to have occurred below 300 km in that section of the South American slab; the closest hypocenter is 250 km south, under the Peru–Brazil border. All other reported seismicity in the area, notably historical events, is erroneous and can be relocated to shallower segments of the Wadati–Benioff zone, or to the local shallow seismic belts, including the Guaviare seismic zone.

This situation is made remarkable by the size of the events in question ($M_0 = 1.2 \times 10^{27}$ (1921), 6×10^{27} (1922) and 2.1×10^{28} (1970) dyn cm) and by the total absence of recorded foreshock and aftershock activity. Moment tensor inversion and/or bodywave modeling indicate focal mechanisms expressing down-dip compression whereas tension axes vary. Source durations for the two historical earthquakes fall within the range observed by Vidale and Houston (1993) for deep earthquakes at the bottom of subduction zones. The 1970 event ($\tau_{\text{scaled}} = 10$ s) fits at the high end of these authors' values. In this respect, the events are anomalous only from the standpoint of their exceptional isolation in space and time.

At this point, we cannot propose a satisfactory explanation for the occurrence of the deep Peru–Colombia shocks. Plate reconstructions indicate that they most probably took place in a fragment of Farallon-generated lithosphere, aged approximately 55 Ma; when compared with other subduction zones, this age would be too young for a remanent thermal signature strong enough to allow deep seismicity. The deep Peru–Colombia shocks must therefore be explained by a more localized circumstance, presumably in the form of a heterogeneity on the descending slab. Hotspot tracks would appear to be potential candidates, but plate reconstructions (at least from currently available models) argue against placing any tracks in the immediate vicinity of the Peru–Colombia hypocenters.

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Note added in proof (22 June 1994)

The occurrence of the deep Bolivian earthquake of 9 June 1994 ($M_0 = 2$ to 3 times 10^{28} dyn-cm) invalidates certain statements in the very first paragraph of the Introduction, regarding the character of unicity of the 1970 Colombian event. While the 1994 Bolivian shock promises to be the subject of intense study in the forthcoming months, preliminary results indicate many similarities with the 1970 event (depth, moment, location in a zone of sparse deep seismicity, felt area extending into North America), with also a number of significant differences (occurrence of at least one large aftershock, location 200 km East of the presumed position of the slab).

It is clear that none of the conclusions of the paper regarding the 1921 and 1922 events are affected; together with the 1970 shock, they remain exceptional, intriguing earthquakes, if somewhat less unique since 09 June.

Appendix

Relocations of historical events

We present here a detailed analysis of the relocation of historical seismicity originally located in the vicinity of the 1921–1922 shocks. None of the 17 earthquakes studied can be conclusively relocated outside of either the South American Benioff plane systems or the Guaviare zone of shallow seismicity in Central Colombia. The three deep events of 1921–1922 and 1970 thus stand out as exceptional occurrences. In the following discussion, events are listed in order of date.

28 April 1911; Original location 0°N, 71°W; 600 km (G-R); $M_{PAS} = 7.1$

At the time of this event, the ISS reported arrival times only to the nearest minute, and the identification of the phases reported is in doubt. Thus no attempt was made to relocate the event based on ISS times. However, an examination of the Golitsyn records (G-R) at Pulkovo (PUL) puts strong constraints on the depth of the event (560 km from the time difference pP − P) and the epicentral distance (104° from the relative times of P, SKS and S). An acceptable epicenter would be 9.5°S, 71.2°W, in the area of the 9 November 1963 shock at the bottom of the Peru–Brazil subduction segment. This hypocenter was confirmed by a CMT inversion of the PUL seismograms (Huang et al., 1994), which yielded a down-dip compressional mechanism, with a moment of 3.7×10^{26} dyn cm (Fig. A1). The geometry, size and inverted depth (557 km) of this event are indeed typical of CMT solutions documented in the area.

16 March 1918; original location 1°N, 70°W (ISS)

The dataset for this earthquake is very poor, with no times available from South American stations. A dataset of seven non-emergent times converges on 14.37°S, 72.68°W ($h = 661$ km; $\sigma = 3.45$ s), an aseismic location in modern times. We prefer a somewhat shallower location at 12.86°S, 70.91°W ($h = 530$ km; $\sigma = 3.62$ s), whose Monte Carlo ellipse ($\sigma_G = 10$ s) covers a large section of the Peru–Brazil deep subduction segment. Although the data provide little latitudinal control, the event is clearly much further south than reported in the ISS.

21 December 1918; original location 1°N, 70°W (ISS)

The dataset of 14 ISS times converges on 0.37°S, 72.23°W at 22 km depth ($\sigma = 3.8$ s), but there is a strong trade-off between depth and longitude. We favor a deeper solution (1.15°S, 74°W; $h = 200$ km; $\sigma = 4.3$ s). Although this epicenter comes closest to the 1970 deep shock, its Monte Carlo ellipse run for $\sigma_G = 10$ s extends to the site of known modern seismicity at that depth (2°S, 77°W; see Fig. 3(c)). Deeper solutions cannot be worked out.

9 September 1923; original location 0°N, 75°W (ISS)

This event is listed only as a possible solution requiring two shocks separated by 46 s, in an attempt by the ISS to unravel a dataset of 15 times of particularly poor quality. We could not obtain a relocation satisfying a significant part of the ISS dataset, and concur that the shock must be multiple. However, the evidence for an epicenter east of the Andes is totally speculative.

28 January 1924; original location 1°N, 70°W (ISS)

There are only five times available for this event. The solution converges on 1.07°N, 73.53°W ($h = 78$ km; $\sigma = 2.04$ s), and the Monte Carlo ellipse ($\sigma_G = 10$ s) intersects significantly the active West–Colombian seismic belt and the Guaviare seismic zone (see Fig. 3(c)).

10 March 1924; original location 2.8°S, 74.5°W (ISS)

The ISS times converge on 5.54°S, 79.83°W at 192 km; however, in agreement with the pattern of modern seismicity, we prefer to constrain the depth to a shallower value (100 km), resulting in a hypocenter within the Benioff plane (4.41°S, 78.13°W) with only a slight deterioration of the residuals ($\sigma = 2.15$ s as opposed to 1.87 s). Both

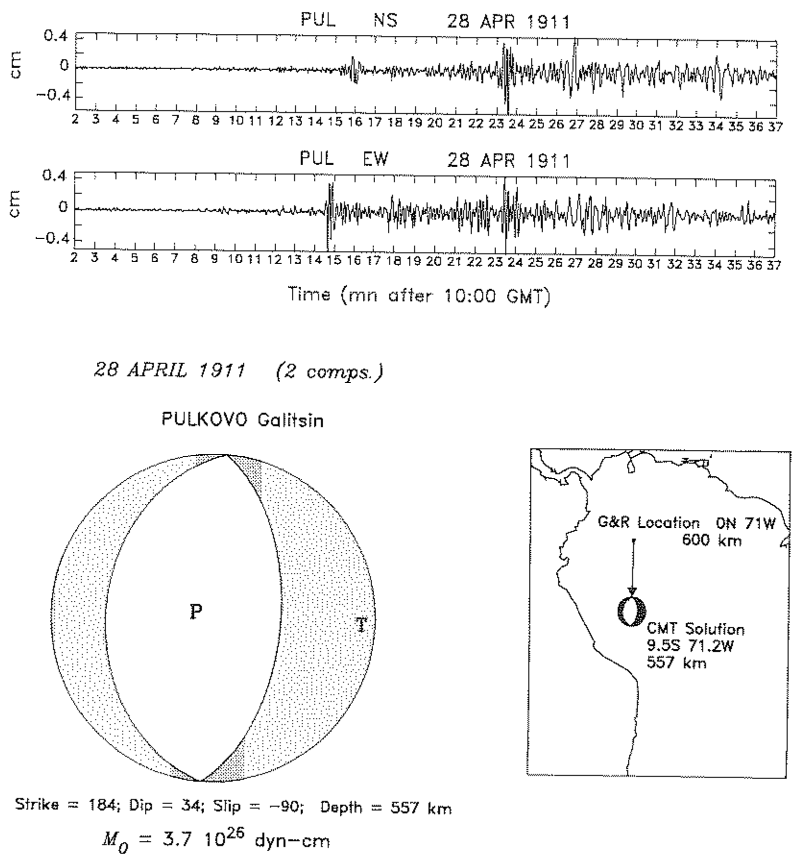

Fig. A1. CMT inversion of the deep 1911 Peru–Brazil event. Top: Digitized Pulkovo seismograms. The traces start at 10:02 h GMT, i.e. approximately 10 min after origin time. Bottom left: inverted CMT solution. (Note down-dip compressional character.) Bottom right: map of the proposed relocation vector for the event. After Huang et al. (1994).

this event and the earthquake 6 days later (see below) have very poor longitudinal resolution, with Monte Carlo ellipses ($\sigma_G = 10$ s) extending over 8° of longitude. They intersect around 4°S, 78°W, suggesting that the two earthquakes may indeed share a common source.

16 March 1924; original location 2°S, 72°W (ISS)

A dataset of 10 ISS times converges on 3.18°S, 77°W at 208 km ($\sigma = 3.64$ s); however, in agreement with the pattern of modern seismicity in the area, we prefer to hold the depth to a slightly shallower value (160 km), resulting in the epicenter 2.96°S, 76.24°W; with only little deterioration of the solution's quality ($\sigma = 4.21$ s).

23 June 1925; original location 0°N, 75°W; 191 km (ISS); $M_{PAS} = 6\frac{3}{4}$

Our relocation moves the event shallower and westwards, to 0.66°S, 76.01°W ($h = 159$ km; $\sigma = 2.86$ s). This is comparable with the G-R estimate for this earthquake: (0°N, 77°W; 180 km). Although there is little longitudinal control, with the Monte Carlo ellipse ($\sigma_G = 10$ s) extending from 81°W to 73°W, the results suggest that the earthquake falls within the Ecuadorian section of the Benioff plane.

7 March 1926; original location 2.8°S, 74.5°W (ISS); $M_{PAS} = 6\frac{1}{2}$

Eighteen out of 22 ISS times converge on a solution within the Benioff plane at 5.27°S, 77.45°W ($h = 93$ km; $\sigma = 3.2$ s). The ISS noted that their solution is 'far from satisfactory', and proposed as an alternative an intraplate epicenter in the Nazca plate, which is equally unsatisfactory. We believe most of the data are explained by our solution, which is also in reasonable agreement with Gutenberg and Richter's (1954) (5°S, 76.5°W, 150 km).

14 July 1926; original location 2.5°S, 71°W (ISS)

The ISS data (4 P and 5 S times) are mutually incompatible, and insufficient to achieve any reliable location. The standard deviation of residuals for the ISS solution is 37 s for a shallow source, and 66 s for a deep source.

24 November 1930; original location 2°S, 74.5°W (ISS); $M_{PAS} = 6\frac{1}{4}$

With the exception of a large S residual at La Paz (also present in the ISS solution), the entire dataset converges easily on 3.44°S, 77.51°W ($h = 135$ km; $\sigma = 2.35$ s), within the Benioff plane in Southern Ecuador. This solution is closer to Gutenberg and Richter's (1954) (2°S, 77°W; 100 km).

24 May 1934; original location 4°S, 73°W (CGS)

A dataset of 11 ISS times converges on (1.39°S, 78.33°W) at 156 km, within the Ecuadorian segment of the Benioff plane, and only within 1° of the ISS solution (0.5°S, 78.1°W; shallow). The CGS solution has a standard deviation of 52 s for this dataset, and is clearly erroneous.

2 August 1937; original location 4.7°S, 73°W; 400 km (CGS)

The dataset for this shock, listed by the ISS only as "Shock probably Central America [sic]", is extremely poor. A possible solution fitting eight of the 13 times is at 4.5°S, 78°W and 70 km ($\sigma = 2.9$ s), but this solution remains speculative. An alternative solution would place the event at 6.40°S, 71.18°W (650 km) in the Peru–Brazil deep seismic zone, fitting nine out of the 13 reported times.

24 April 1938; original location 2°N, 69°W (CGS)

No times are listed in the ISS, and the BCIS gives only three P times (TUC, PAS and LPZ), and an S at LPZ. This limited dataset would converge to an epicenter north of the Guaviare zone, at 4.71°N, 70.83°W. There is no resolution in the WSW–ENE direction, with the Monte Carlo ellipse extending 15° across Central Colombia and Northern Venezuela, even when run with a Gaussian deviation $\sigma_G = 5$ s, a conservative figure for 1938. There is no evidence that the event is outside the Colombian seismic belt. Although the earthquake took place 6 h after a moderate event of intermediate depth in Argentina, the available times are incompatible with interpreting it as an aftershock.

3 September 1944; original location 3°N, 71.5°W (CGS)

This event is given two significantly different locations by the CGS and ISS. The CGS solution, at the north of the Guaviare Zone, is grossly in error, and our relocation converges on 3.17°N, 76.71°W at 116 km, well within the Northern Colombia Benioff plane, and in basic agreement with the ISS solution (3.3°N, 77.2°W).

13 September 1960; original location 5°S, 74.5°W; 119 km (CGS)

This event is not reported in the ISS, but on the basis of BCIS times, we relocate the earthquake at 15.00°S, 75.51°W, with the available times failing to constrain the depth. Residuals for the CGS solution are clearly unacceptable, with a standard deviation approaching 2 min. We suggest that the erroneous CGS epicenter is the result of a typographic error (5°S instead of 15°S).

Recent events

In addition, we discuss the following more recent events, which we have found in error.

23 January 1964; original location 2.5°S, 80.1°W; 418 km (USGS); $m_b = 3.6$

This hypocenter is clearly in error, as it is located seaward of the Benioff zone, right under the port city of Guayaquil. The earthquake is also conspicuous in that it is the only event at that depth in the Colombia–Ecuador subduction system. The ISC location is poor, achieving only a standard deviation $\sigma = 5.1$ s. We note that the depth of the event is totally controlled by the station BOG. If BOG is eliminated, the solution converges to a shallow source for which both P and S residuals at BOG are close to 60 s, strongly suggesting that BOG is in error by 1 min. Effecting such a correction results in an acceptable hypocenter at 0.56°S, 78.57°W (10 km) with $\sigma =$ 2.9 s.

7 December 1964; original location 5.5°S, 80.3°W; 345 km (USGS); $m_b = 3.5$

This small earthquake is very poorly located. The ISC proposes a depth of 203 km, with $\sigma =$ 4.64 s on 15 P times. The times at TRN (early) and SPA (late) are mutually incompatible. Depending upon which one is ignored, the hypocenter is moved up or down, but with little effect on the epicentral location. We prefer the shallow solution, which relocates offshore near the trench (5.06°S, 82.25°W; $h = 33$ km; $\sigma = 2.66$ s on 13 stations), at the cost of ignoring the emergent TRN arrival, and keeping the impulsive one at SPA.

21 March 1974; original location 4.533°S, 73.396°W; 48 km (ISC); $m_b = 4.8$

This modern event is challenging in several ways: it is located in an area of very low modern seismicity, with the only other known shock 50 km further south, and the ISS solution features excessive residuals (13 s for P and 32 s for S) at the closest station (Quito), with an additional 23 s residual for S at Huancayo. We attempted to relocate the event, but all efforts at moving the source failed, and the proposed hypocenter is robust. The origin of the excessive residuals at QUI and HUA is unclear.

12 February 1991: original location 9.47°S, 70.61°W; 396 km (NEIC); $m_b = 4.4$

This location is labeled 'poor' by the NEIC. The dataset published by the ISC relocates at 483 km when using its eight P times ($\sigma_P = 0.61$ s), and 503 km ($\sigma_{PS} = 0.79$ s) when adding the three reported S times. However, the USGS hypocenter results in $\sigma_P = 0.79$ s, and one must conclude that the dataset lacks depth resolution. There is no reason to place the event in the otherwise quiet depth range 300–500 km.

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