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Thrust and extensional faulting under the Chilean coast: 1965, 1971 Aconcagua earthquakes

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Summary. Earthquakes under coastal northern Chile, from 18° S to 33° S occur at two very different depths. The shallower interplate events have the typical thrust mechanisms associated with subduction of the Nazca Plate under South America. The deeper (close to 70 km) events have normal fault mechanisms with a tension axis approximately parallel to the dip of the downgoing slab. We study in detail a couple of $M_s \approx 7.5$ events that took place about 60 km north of Valparaiso, Chile. The 1971 July 9 event was a shallow thrust interplate event at 40 km depth, while the 1965 March 28 earthquake had a normal fault mechanism and a depth of 72 km. We carry out a detailed analysis of long period *P*- and surface waves observed in the WWSSN network of stations in order to improve fault plane solutions and depth determinations. For the 1971 event we find a seismic moment of 5.6×10^{27} dyne cm from both *P*- and surface waves. The source area for this event is relatively small compared to that of other underthrust events with similar moments. Assuming a circular fault shape and a rigidity of 7×10^{11} dyne cm^{-2} we find a stress drop of 38 bar. The 1965 event is more complex; seismic moments of 1.8 and 1.0×10^{27} dyne cm are found from surface and body waves respectively. The stress drop estimated from *P*-waves and a circular fault model is close to 91 bar.

The existence of earthquakes at two different depths near the Chilean coast may be interpreted in terms of a double layered Benioff zone, as has been found in Honshu, Kuriles or central Peru. Unfortunately the hypocentral determinations of events in this region are not precise enough to separate the two layers. An alternative method to demonstrate the existence of a double Benioff zone is to look for compressional events associated with the upper layer of the down-going slab. A search among available fault plane solutions is ambiguous; three shallow coastal events are found in Stauder's catalogue with horizontal pressure axes normal to the trench, but they might be intra-continental plate events.

Introduction

The Peru–Chile trench is the site of some of the largest earthquakes, yet few of these events have been studied in detail. Recent analysis of the seismicity and fault mechanisms (Isacks & Molnar 1971; Stauder 1973, 1975; Barazangi & Isacks 1976, 1979; Isacks & Barazangi 1977; James 1978; Snoke, Sacks & James 1979; Santo 1969; Hanus & Vanek 1978; Hasegawa & Sacks 1979) have demonstrated that this subduction zone is segmented. Barazangi & Isacks (1976) proposed a division into five zones: two of them under central Peru and north-central Chile would be under low angle subduction, while the rest would have steeper, 30° dip Benioff zones. Another model proposed by James (1978) and Snoke *et al.* (1979), proposes a uniform subduction angle of 30°. Recently, Hasegawa & Sacks (1979) have revised this model proposing 30° subduction down to 100 km and then horizontal subduction under central Peru. Whatever the result of the controversy, it is clear that the segmented nature of the downgoing slab should have important consequences for the nature and mechanism of large, destructive earthquakes under the Peru–Chile coast.

Several large events in this area have been studied in detail using teleseismic data: the 1960 southern Chile events (Kanamori & Cipar 1974; Plafker 1972), the 1960 slow events (Kanamori & Stewart 1979); the 1966, 1970 and 1974 Lima earthquakes (Abe 1972; Dewey & Spence 1979). Most of these earthquakes are of the usual underthrust type at the contact between the Nazca plate and the overriding South America plate. Isacks & Barazangi (1977) and Dewey & Spence (1979) have remarked that the 1970 Peru earthquake was a very

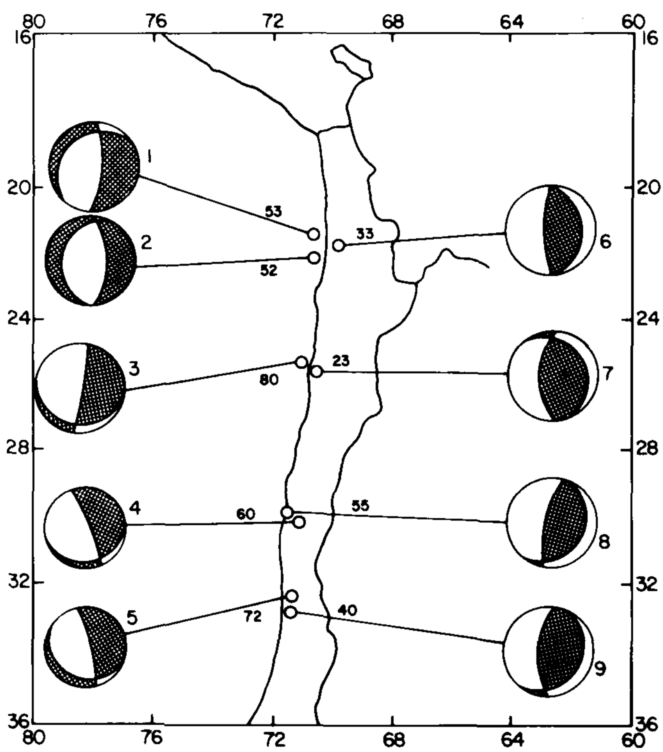

Figure 1. Selected coastal events in northern Chile chosen so as to put in evidence a systematic pairing of shallow underthrust events (numbers 6–9) and deeper normal fault ones (1–5). Depths, indicated by the small numbers near the epicentres, are from ISC bulletins except for events 5 and 9 which we study in this paper. Focal mechanisms are from Stauder (1973) except again for events 5 and 9.

complex normal fault event that occurred probably inside the downgoing slab. Careful relocalization by Dewey & Spence (1979) showed that this event was located 30 km under the contact between the two plates.

We have reviewed available fault mechanisms for the Peru–Chile trench (Stauder 1973, 1975) and found a systematic pairing of underthrust and normal events all along the Pacific coast, from Lima to Santiago. Underthrust mechanism events are shallow (40–50 km) while normal fault ones have intermediate depths, as seen in Fig. 1. Location and especially depth control are poor all along this area, so that a careful study of the mechanism and depth of these events is necessary. We have selected for study the events of 1965 March 23, 1965 March 28 and 1971 July 9 in the Aconcagua province, 60 km north of Valparaiso, Chile. We have tried to obtain refined estimates of the fault plane solutions and fault parameters from surface and body waves. Depth has been redetermined modelling long-period *P*-wave pulses with direct and reflected phases. These two earthquakes are also of great interest from an engineering viewpoint since they produced extensive damage; Pereira, Crampien & Saragoni (1979) have stressed significant differences in the intensity distribution produced by the two earthquakes.

Normal and thrust faulting under the Peru–Chile coastline

A study of available fault mechanism for South-American earthquakes (Stauder 1973, 1975; Isacks & Molnar 1971) shows a clear pattern of normal and thrust-faulting under the Pacific coast: all along the subduction zone from 18° S to 33° S, shallow underthrust events are found at the same locations as deeper normal fault ones. We have found a total of five such pairs in Chile which we have listed in Table 1, and plotted in Fig. 1 together with the depths as given by ISC and the fault mechanisms determined either by Stauder (1973) or ourselves. Most of the deeper events were also studied by Isacks & Molnar (1971) but their mechanisms do not differ significantly from those of Stauder.

The presence of the deeper normal fault events at essentially the same place where shallower thrust events occur had already been noticed in the Lima region of Peru. Abe (1972) showed that the 1970 May 30 event was a large ($M_s = 7.8$) normal fault intraplate event and that it was deeper than the 1966 October 17 ($M_s = 7.5$) underthrust earthquake. The aftershock zones of these two events are very close to each other. Abe (1972) interpreted the 1966 event as due to underthrust of the Nazca plate under the South American plate and the 1970 event as normal faulting of the downgoing slab. Further study of the 1970 event by Barazangi & Isacks (1977) showed that some of its aftershocks had down-dip compressional mechanisms and proposed that these events corresponded to the upper layer of a double Benioff zone similar to that which has been found under north Honshu by Hasegawa, Umino & Tagaki (1978) from small earthquakes located with a local network.

Table 1. Shallow thrust and deeper normal fault earthquakes under the Chilean coast.

No.	Date	Time	Coordinates	Depth	Magnitude	Mechanism
1	1967 December 25	10:41	21.5 S 70.4 W	53	5.8	Stauder
2	1970 June 19	10:56	22.2 S 70.5 W	52	6.2	Stauder
3	1965 February 23	21:11	25.7 S 70.5 W	80	6.2	Stauder
4	1963 March 10	10:51	30.0 S 71.2 W	60	6.3	Stauder
5	1965 March 28	16:33	32.4 S 71.2 W	72	6.4	This paper
6	1967 December 21	02:25	21.8 S 70.0 W	33	6.3	Stauder
7	1966 December 28	08:19	25.5 S 70.7 W	23	6.9	Stauder
8	1967 September 26	16:11	30.0 S 71.5 W	55	6.0	Stauder
9	1971 July 9	03:03	32.5 S 71.2 W	42	6.6	This paper

This opens the intriguing possibility that the large down-dip extensional events may be part of the lower layer of a double Benioff zone; or, perhaps, that some common mechanism may be responsible for the pairing of large extensional and thrust events and of the double layer. A more detailed examination of the source mechanism and depth of some minor compressional events listed in Stauder's (1973) catalogue is being pursued to clarify this point.

La Ligua earthquakes of 1965 and 1971

The coastal region of the Aconcagua Province in Chile, to the north of Valparaiso has been extremely active since 1965. This area corresponds to the extreme north of the rupture zone of the great 1906 Valparaiso Earthquake. Two major earthquakes ($M_s \approx 7.5$) occurred in that region on 1965 March 28 and 1971 July 9 (see Fig. 2). Their coordinates, origin times etc. are given in Table 2 together with those of a smaller event ($M_s = 6.0$) of 1965 March 22. This event may be interpreted as a foreshock of the 1965 March 28 event, yet as we shall see they have very different fault plane solutions and depths. Fault plane solutions for the two events of 1965 March have been obtained by Stauder (1973). In addition Isacks & Molnar (1971) gave a solution for the larger 1965 March 28 event. We have re-determined fault plane solutions for all these events (Fig. 3) which we obtained from first motions of *P*- and *S*-waves and the long-period surface wave observations. The best constrained solution is the one for 1971 in which we used data from the Chilean long-period stations. Our prefered solutions for the 1965 March 28 event, based on *P*- and surface waves is somewhat

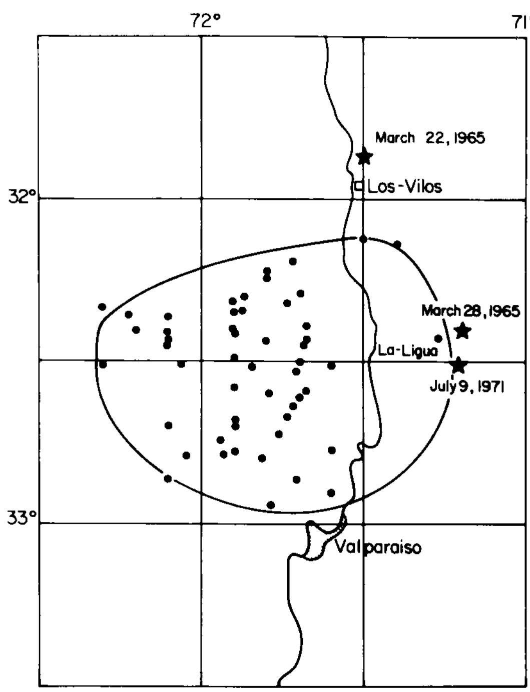

Figure 2. Epicentral area. Stars mark the epicentres for the two large earthquakes of 1965 March 28 and 1971 July 9 and of the smaller one of 1965 March 22. Dots indicate the aftershocks of the shallow underthrust event of 1971 July 9. Epicentral data for all these events are from ISC bulletins. The 1965 March 28 earthquake had only two aftershocks listed by ISC, located at practically the same place as the main event.

Table 2. Source parameters of the Aconcagua Earthquakes of 1965 and 1971.

Date	1965 March 22	1965 March 28	1971 July 9
Time	22:56:26.5	16:33:14.6	03:03:18.7
Latitude (S)	31.95	32.4	32.54
Longitude (W)	71.5	71.2	71.15
Depth (km)	46	72	40
m_b	6.0	6.4	6.6
M_S	6.0	7.5	7.5
M_0 (P) (dyne cm)	4×10^{25}	1×10^{27}	5.6×10^{27}
M_0 (surf) (dyne cm)	–	1.8×10^{27}	5.6×10^{27}
$\langle T \rangle$ (s)	<1	6	16
$\Delta\sigma$ (bar)	–	91	38
Slip (m)	–	1.6	1.6–5.7
Dip	70° W	80° E	66° W
Strike (°)	191	348	180
Rake (°)	106	−95 to −105	87

different from those found by Isacks & Molnar (1971) or Stauder (1973). The two large events are significantly different: 1965 March 28 is a normal fault, intraplate event (H=70 km), while the event of 1971 July 9 is a typical shallow dipping thrust fault. Its depth (40 km) indicates that it is an interplate event associated with the slip of the Nazca plate under the South American one. The 1965 March 22 foreshock is also shallow and has

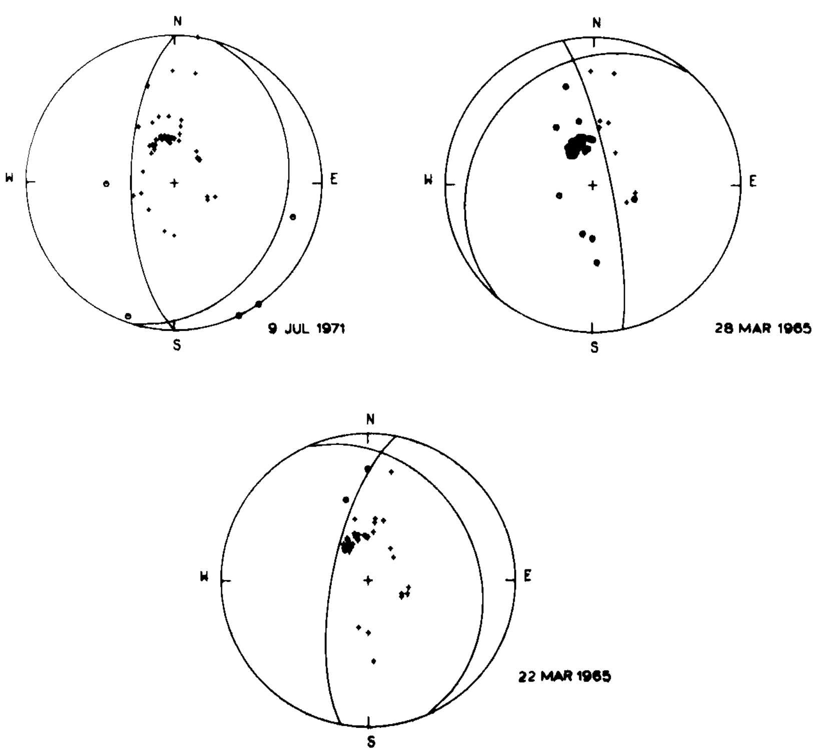

Figure 3. Fault plane solutions for three Aconcagua events. There are lower hemisphere equal area projections of P-wave first motions as observed on long-period stations. Dilatations are indicated by circles, compressions by crosses. Fault planes are chosen to satisfy additional constraints from S-wave polarization and surface wave radiation patterns.

a thrust mechanism indicating interplate slip. It is difficult to decide whether this is a true foreshock of the 1965 March 28 earthquake or whether their time separation (one week) is a matter of pure chance. In this context, it is interesting to remark that the 1967 December 21 shallow thrust, northern Chile event had a large deeper, normal faulting aftershock on 1967 December 25 (see Fig. 1 and Table 1).

Another major difference between the two large events is that the 1971 one had a large number of aftershocks as determined by ISC (see Fig. 2), while only two aftershocks with essentially the same hypocentre accompanied the normal fault 1965 event. This may be due in part to a significant improvement in instrumentation in central Chile and Argentina during this time. However, local networks were installed in both cases a few days after the events (Kausel, private communication; Eisenberg, Husid & Luco 1972). Data obtained in 1965 are not sufficient to perform hypocentral location due to poor time control on most stations. After the 1965 March 28 event the number of aftershocks registered by the local network was very small, on the order of five per day. From $S-P$ times read at the epicentral area near La Ligua one concludes that about one-half of these events were located at least 50 km away, probably near the hypocentral zone of the smaller 1965 March 22 interplate event. In contrast the large 1971 event had an important series of aftershocks, local stations registered tens of aftershocks a day during three months after the main event. Unfortunately most of these data are still unpublished. In Fig. 2 we plot the epicentre of the aftershocks located by ISC. If we considered this to be a good approximation to the fault area, we may conclude that the 1971 earthquake started inland and propagated towards the trench. We have not found enough directivity in our data to confirm this observation.

Finally, these two events appear to be different in their intensity distribution: Pereira *et al.* (1979) observe that the 1971 event had isoseismals that are elongated in a north–south direction (Eisenberg *et al.* 1972) while the 1965 event had isoseismals elongated across Chile (Kausel 1965). It is difficult to see how to relate these observations to a source mechanism or depth of these earthquakes. Both events produced intensity X to IX in the La Ligua area and significant damage to structures along the coast down to the port of Valparaiso.

Long-period surface wave study

We have studied the long-period Love and Rayleigh waves of the two larger events with the purpose of calculating very long-period seismic moments and to refine the nodal plane solutions. We chose unsaturated multiple passage Love or Rayleigh waves, usually G_2, G_3, R_2 or R_3 as observed in the WWSSN long-period stations. The records are windowed for the appropriate group velocities, digitized, smoothed and filtered to remove surface waves with periods under 60 s. The data are then equalized to a distance of 270° using model 1066B of Dziewonski & Gilbert (1975) for phase velocities, and attenuation derived from model SL8 of Anderson & Hart (1978). A standard long-period WWSSN instrument is assumed. The technique is essentially the same proposed by Kanamori & Stewart (1976) and a full description may be found in Deschamps, Lyon-Caen & Madariaga (1980). The data are now compared with synthetic seismograms calculated for the same earth models as above (i.e. 1066B for velocities, SL8 for attenuation). The source model is a point double couple at the hypocentre, represented by a moment tensor. Synthetics are constructed by superposition of fundamental modes, whose excitation is calculated by a method derived from Dziewonski & Gilbert (1975) and fully described in Deschamps *et al.* (1980). The synthetics are calculated at the equalization distance ($\Delta = 270°$) and convolved with the instrument response, the attenuation operator and the source time function. For earthquakes of the size studied here, where the source area is less than a 100 km diameter, a point source is a reasonable approximation. A Heaviside step function was assumed for the source time function.

In Figs 4 and 5 we compare the equalized, filtered data for the 1965 normal fault event with synthetic seismograms. As is typical of earthquakes with an almost vertical fault plane, the Rayleigh waves in Fig. 4 have a two-lobed radiation pattern. Theoretical Love waves in Fig. 5 are four-lobed and are very sensitive to the orientation of fault planes. Unfortunately we do not have enough data to fix unequivocally the shallow fault plane. Considering all the available data: first motions for P- and S-waves and the surface waves, we finally came up with the source mechanism represented in Fig. 3. The almost vertical fault plane in this solution is very well constrained and is very close to that proposed by Stauder (1973). The shallow dipping plane is less constrained but is different from that of Stauder, his solution is inconsistent with our Love and Rayleigh waves. In Fig. 6 we show the observed peak-to-peak amplitudes of Rayleigh and Love waves as a function of azimuth. We select the fault plane solution by fitting the theoretical radiation pattern (continuous line) to the observations. We finally find for the steeper fault plane a strike of $\phi = 348°$, a dip of $\delta = 80°$ E and a rake that may vary from $-95° < \lambda < -105°$. For ϕ and δ, $\pm 1°$ uncertainties are appropriate. The seismic moment was obtained from a comparison of the amplitudes of observed and calculated surface waves. We obtain $M_0 = 1.8 \times 10^{20}$ Nm (1.8×10^{27} dyne cm). Let us finally note that this seismic moment is calculated at a slightly longer period than that of the 1971 events, because most long-period stations had a seismometer of $T_s = 30$ s in 1965 March.

Next, we study the surface wave radiation for the large thrust event of 1971 July. In Figs 7 and 8 we compare observed and calculated Rayleigh and Love waves respectively. As in the previous case theoretical Rayleigh waves are two-lobed while Love waves are four-lobed. The observed data agree quite closely with the synthetic records calculated for the fault plane mechanism of Fig. 3 which agrees also with observed first motion data for P-waves. Thus in this case we have an excellent control of the fault planes, which, referred to the almost

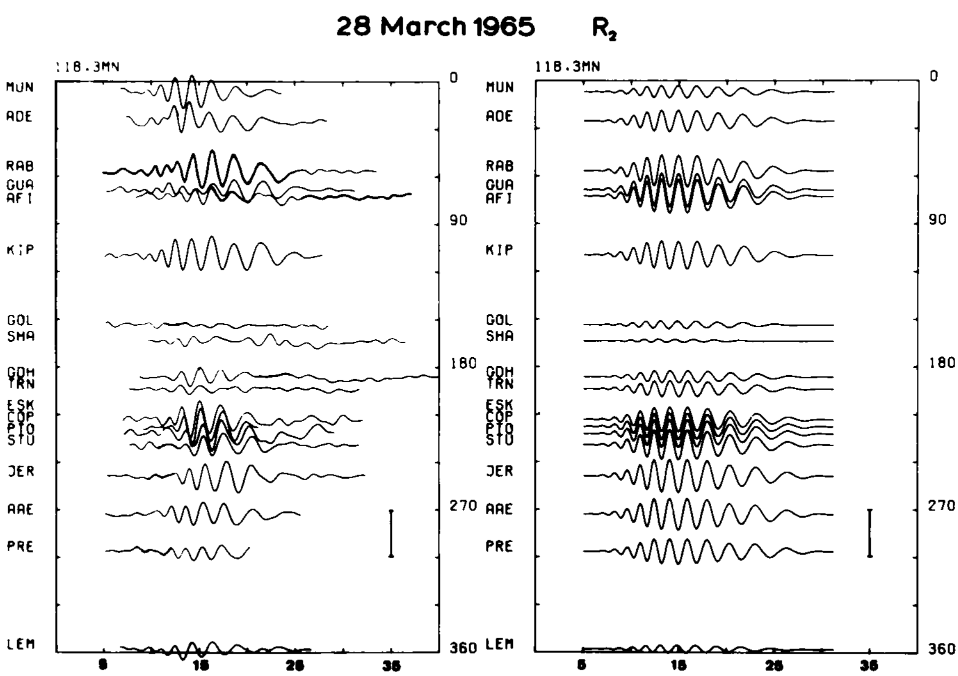

Figure 4. Observed (left) and theoretical (right) long-period Rayleigh waves for the 1965 March 28 intermediate depth normal fault event. Vertical line represents 2 cm on a WWSSN record of magnification 1500.

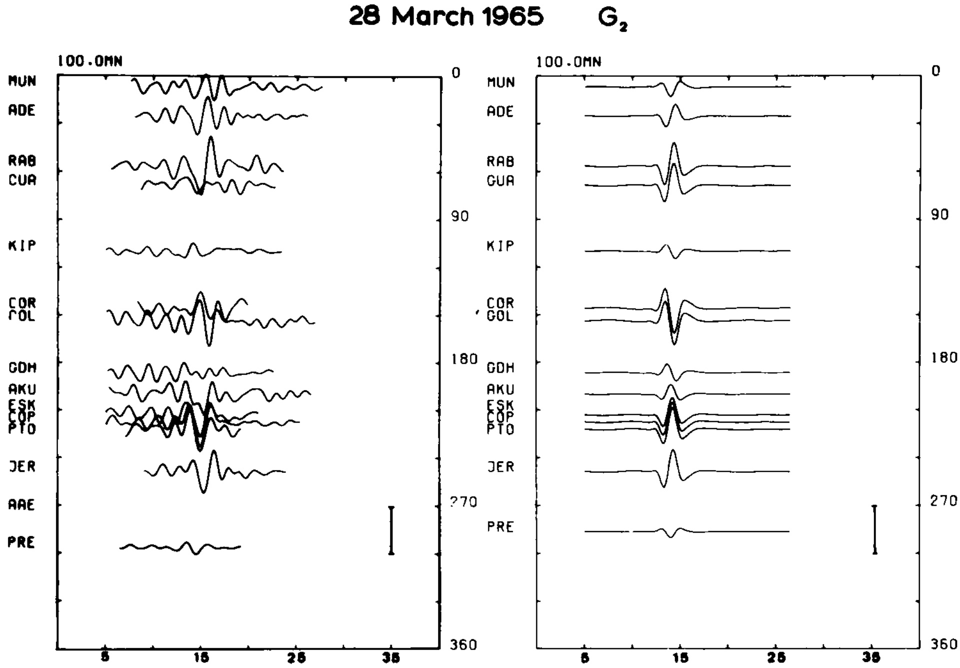

Figure 5. Observed (left) and theoretical (right) long-period Love waves for the 1965 March 28 event. The vertical line represents 2 cm on a WWSSN record of magnification 1500.

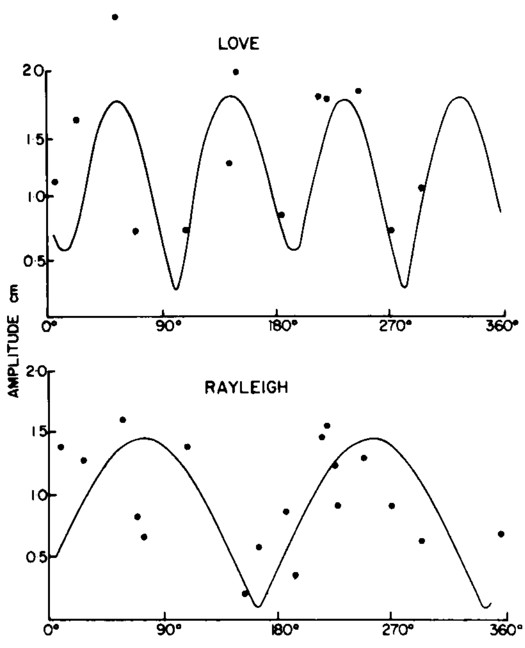

Figure 6. Theoretical (continuous line) versus observed (dots) maximum amplitudes of Love and Rayleigh waves as a function of azimuth. Theoretical curves are obtained for a fault plane with strike 348°, dip 80° E and rake −100°.

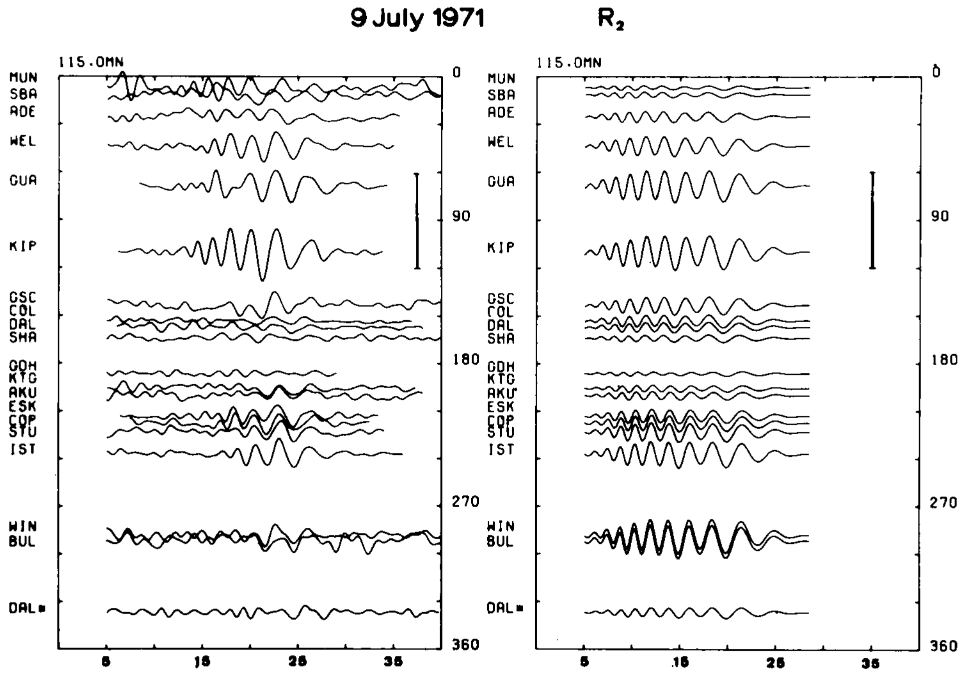

Figure 7. Observed (left) versus theoretical (right) long-period Rayleigh waves for the 1971 July 9 earthquake. The vertical line represents 2 cm on a WWSSN record with amplification 1500.

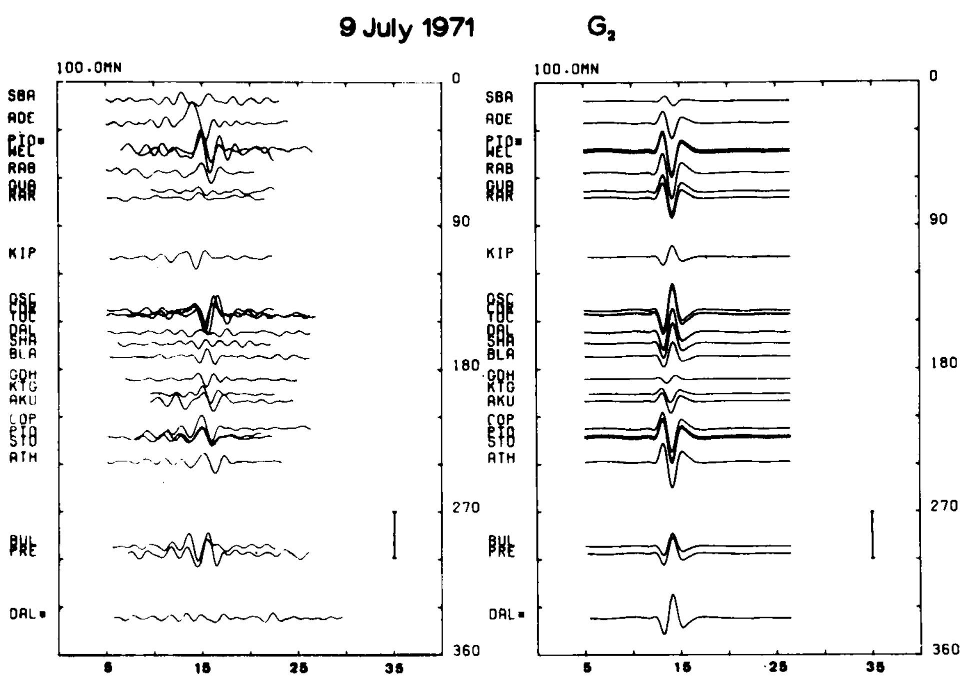

Figure 8. Observed (left) versus theoretical (right) long-period Love waves for the 1971 July 9 earthquake. The vertical line represents 2 cm on a WWSSN record with amplification 1500.

vertical plane, are given by $\phi=180°$, $\delta=66°$ W and $\lambda=87°$. There is some slight directivity in the radiation of Rayleigh waves, but this is mostly due to the KIP (Kippapa, Hawaii) record. We think that the point source model is quite adequate for this event. The seismic moment is estimated as $M_0=5.6\times10^{20}$ N m (5.6×10^{27} dyne cm).

Long-period *P*-wave study

The most serious problems with hypocentral determinations in most of the Pacific coast of America is poor depth control. This control has improved significantly in recent times due to the regional networks installed in Chile, Peru and Argentina. Crucial for the interpretation of the events of La Ligua is a good control of source depth. We shall try to determine their depth from the modelling of the complete *P*-wave train, including direct and surface reflected arrivals. The procedure used is to compare synthetic *P*-wave seismograms directly to observed ones. In this form we avoid the difficult problem of deconvolving the instrumental responses. Observed *P*-waves are simply digitized and brought to a common amplification and time-scale. Only those WWSSN records have been used for which signals could be traced with confidence. For the smaller 1965 March 22 event some noise is seen preceding the *P*-waves; we have not attempted to eliminate this noise from the records. Synthetic seismograms are calculated for a simple point double couple excitation, and propagated by ray theory in a spherical earth model (1066B in our examples). Initially only free surface reflections were included following Kanamori & Stewart (1976) and Deschamps *et al.* (1980). Later, we realised that sub-Moho reflections were important especially for the deeper 1965 March 28 event. Then a simple crustal model with a 33 km crust overlying a normal mantle was assumed in the source area. This model is a very schematic structure determined for most of the coastal area of the Peru–Chile convergence zone (Kulm, Schweller & Masias 1977). No refraction profiles are available on the continental side of the coast to improve our model. Synthetic seismograms were calculated by means of Haskell's layer matrix technique near the source and then propagated by ray theory into the far field. To avoid problems with triplications due to upper mantle discontinuities, only waves at distances between 30° and 90° were used. The ray theoretical seismograms were then convolved in the frequency domain with an attenuation operator ($T/Q=1$ was assumed), the free surface response at the receiver, the instrument response and the source time function. The source function used here was very simple due to the assumption of a point source implicit in the synthetic seismogram calculation. This assumption implies the same source function in all directions independent of the angle of radiation; such an assumption is probably adequate for the 1965 events which had small source areas. At that time also most WWSSN stations had 30 s long-period seismometers so that their time resolution was not enough to detect finiteness effects. For the 1971 event with a source duration of about 20 s some finiteness should be observable; yet the source time function is difficult to recover because of the reflected phases: a trade-off between signal duration and depth is always possible in the modelling so that no finiteness effects could be determined with certainty. For these reasons a point source model was used and source pulses were approximated by trapezoids. Other source functions, derived from dynamic source models, are equally possible but the synthetic seismograms would be undistinguishable from those calculated with trapezoids.

Let us consider first the *P*-waves observed for the 1965 March 28 intermediate depth, normal fault event. This earthquake has the right size to be well observed in most WWSSN stations; in Fig. 9 we present all the unsaturated *P*-wave observations that we could obtain. As is usual with events in South America we find a nest of stations in North America and

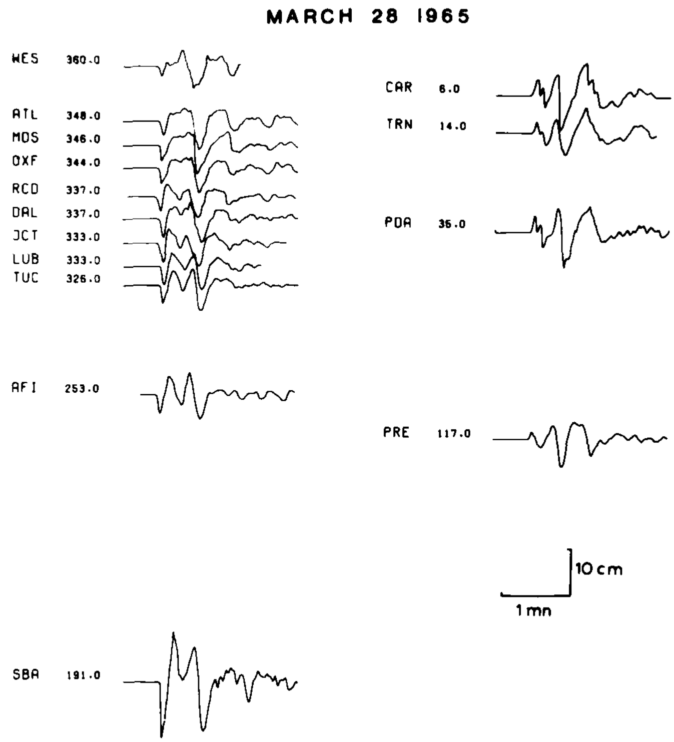

Figure 9. Observed *P*-waves for the 1965 March 28 event. The stations are distributed according to their azimuths, indicated by the numbers next to the station code. Vertical bar indicates 10 cm on a WWSSN long-period record with amplification 1500.

very few data from the other azimuths. The signals present a very systematic variation of shape with azimuths; we present them in two columns: to the left dilatational first motions corresponding roughly to radiation towards the coast of Chile; to the right a few compressions from Caribbean and African stations. The coherency and smooth variation of the signal registered in North America (WES to TUC) is striking. The double impulsions seen in these stations correspond to the direct *P* arrival and the surface reflected *sP*-wave. For most North American stations, especially near LUB and TUC *pP* is near a nodal plane. This is a very neat verification of the fault plane mechanism obtained previously. The time separation between the two negative peaks provides a strong control on the depth of this event. This can be done with good accuracy because of the short duration of the source pulse, estimated as 6 s. There is then none of the usual trade-off between signal duration and depth of the earthquake. The calculated depth depends obviously on the assumed velocity model, this may be quite important in a place like Chile where a high velocity slab is being subducted. Near the coast where the reflected phases propagate, there are no models for the local structure. For this reason we use, as previously mentioned, a simple crust of 33 km over a half space to model this event (Kulm *et al.* 1977). The mean *P*-wave velocity above the source in this model is 7.8 km s^{-1} only slightly lower than the upper mantle velocity of 8.1 km s^{-1}, usually quoted for the downgoing slab. We determined finally a focal depth of 72 km; the error from *P*-wave modelling is estimated as ± 3 km but it may increase significantly once the model uncertainties are defined. In Fig. 10 we show the synthetic *P*-wave group calculated for the same stations as in Fig. 9. The quality of the fit is quite acceptable; and we are able to explain very nicely the change of signal shape between the

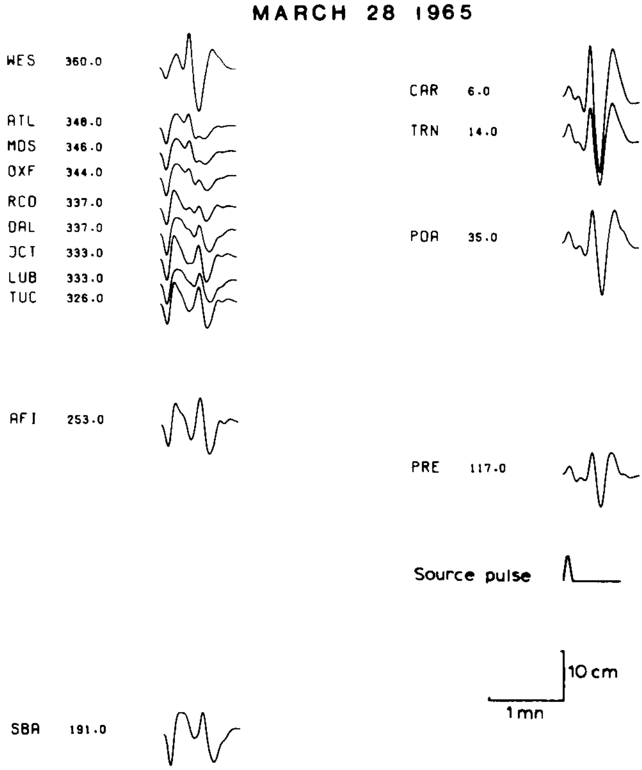

Figure 10. Theoretical P-waves for the 1965 March 28 event. The seismograms were calculated for the prefered fault plane solution (Fig. 3) and a depth of 72 km. At the bottom right we show the trapezoidal source time function used to generate the synthetics.

stations to the east and west by a simple reversal of sign of the P, pP group, while the sP phase keeps the same sign at all stations. We notice that the fit is quite good up to the negative peak of the sP-wave; yet we are not able, with the simple model, to explain the long duration of the second positive peak. We suspect that a second source impulsion arrives hidden behind the sP-wave, but we shall not pursue this search for source multiplicity here. The seismic moment was calculated averaging the moments calculated by fitting every station, we find $M_0 = 1 \times 10^{20}$ N m (10^{27} dyne cm). This is roughly one-half of the seismic moment found from surface waves, a further reason to believe that we are missing some source multiplicity in our models. The signal used in generating the synthetics was a simple trapezoid of duration 6 s, as shown in Fig. 10. Rise time is very poorly constrained from our data, we chose 2 s but we do not claim any precision for this value.

We studied next the larger, shallower thrust interplate event of 1971 July 9. This is definitely a larger event and many WWSSN stations were saturated by the P-waves. We have, however, assembled a good collection of data covering most of the azimuths. As is usual with shallow thrust event in Chile practically all stations with $\Delta > 30°$ (the only ones used in our study) display compressional first motion. Of the data shown on Fig. 11 only RAR is close to the steeply dipping nodal plane. Signals are relatively simple and we are able to model them very approximately with a simple half space model near the source. After several trials we obtain a source depth of 40 km and a source time function, shown in Fig. 12, which is a

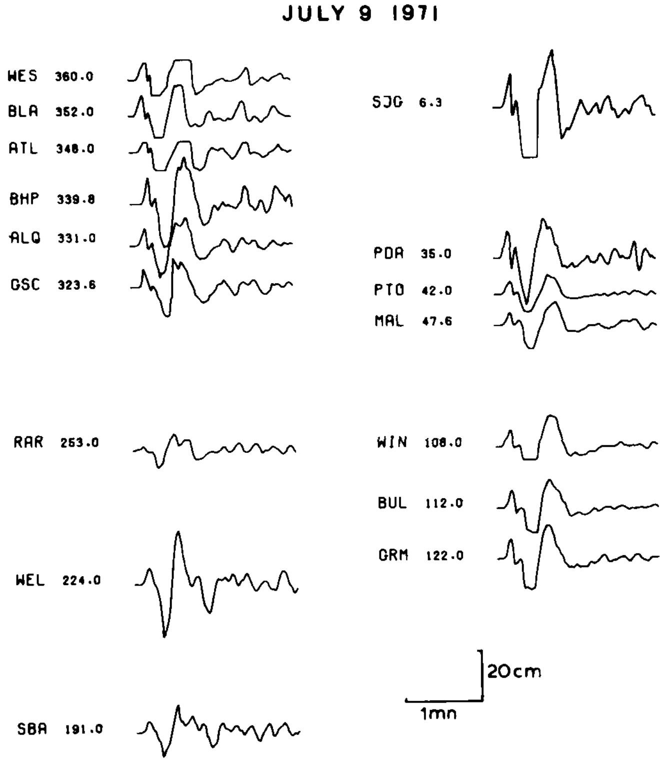

Figure 11. Observed *P*-waves for the 1971 July 9 earthquake. The stations are roughly distributed according to their azimuths, indicated by the numbers next to the station codes. Vertical line indicates 20 cm on a WWSSN long-period record of amplification 1500.

modified trapezoidal impulse: it is actually a double trapezoid. Total duration of the signal is close to 20 s and a rather long rise time of 4 s is found necessary to explain the weak *P* first motions. The second impulse in the source time signal was adopted in order to explain the small inflection observed in most stations about 9 s after the *P* onset. In fact we have not been able to model this weak pulse more precisely, probably because it comes from a different place on the fault plane than the main *P*-wave. The seismic moment estimated from the *P*-waves is entirely compatible with that found from the surface waves, i.e. $M_0 = 5.6 \times 10^{20}$ N m (5.6×10^{27} dyne cm).

Finally we studied the smaller, 1965 March 22 event. This event is much smaller than the other two so that its *P*-waves in long-period stations are very weak and very often contaminated by noise. In Fig. 13 we show the *P*-wave records that we have been able to recover; as in the 1971 event only compressions are available from distant stations. The western US station BOZ is very close to the steeply dipping nodal plane. Source duration is very short, less than 1 s, and therefore impossible to calculate from long period WWSSN records which respond to these short pulses with their impulse responses. We identify very clearly the *sP* phase in most stations; from these observations we find a focal depth close to 46 km, that is approximately the same as for the 1971 event. This event is clearly a small thrust earthquake at the interface between the South American and the Nazca plate. The seismic moment calculated from the *P*-waves is $M_0 = 4 \times 10^{18}$ N m (4×10^{25} dyne cm).

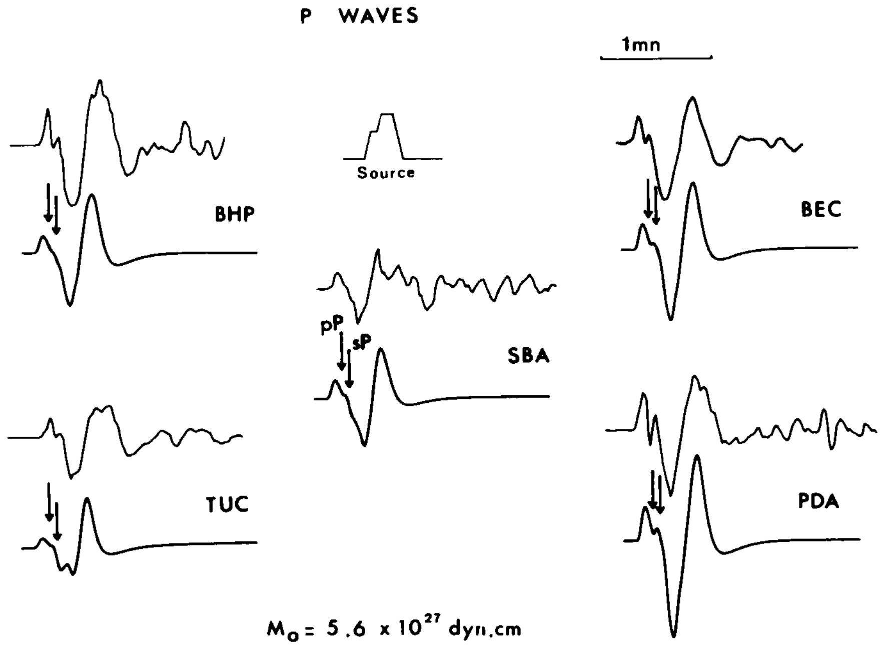

Figure 12. Comparison of observed versus synthetic *P*-waves of the 1971 July 9 earthquake for a few selected stations. *pP* and *sP* arrivals are indicated by the arrows. Source time function is the modified trapezoid shown in the centre.

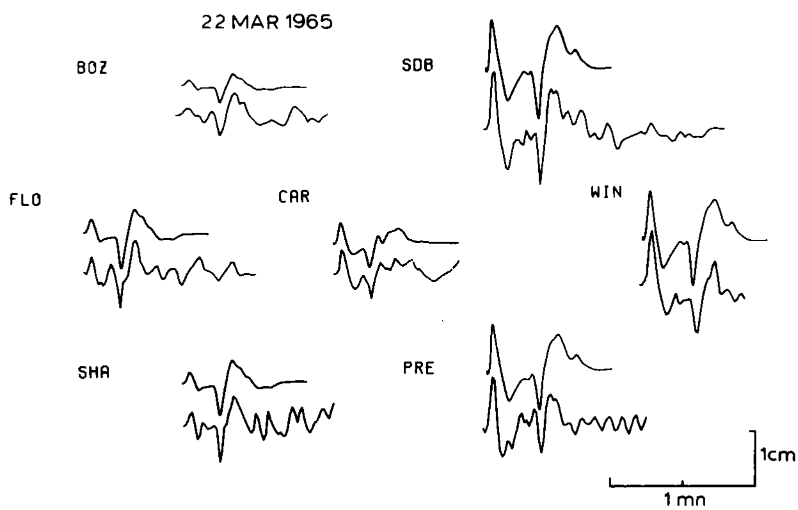

Figure 13. Comparison of observed versus synthetic *P*-waves for the smaller event of 1965 March 22. Synthetics are shown above the observed records for several WWSSN stations. A focal depth of 46 km was determined from the modelling.

Source parameters

The data we have just obtained allow us to make some rough estimates of the slip and stress drop for the larger events. In order to do this we need estimates of the source areas. For the 1971 event we have a rather large number of aftershocks recorded for about a month after the main event. As seen in Fig. 2 these aftershocks define a source area which is roughly circular in shape, with a radius of about $r = 40$ km. This very likely overestimates the actual source area because preliminary data from a local network, installed a few days after the main shock, showed a clear migration of the aftershocks towards Valparaiso. The main shock is seen at one end of the source area, this is a typical problem with hypocentral determinations in central Chile. The epicentres of large events located with worldwide data are usually further inland than their epicentres determined by the local networks in Chile (Barrientos 1980). This may further reduce the source area. From all those considerations we estimate the error in the source area to be close to a factor of two. We may check this value from the apparent width $\langle T \rangle$ of the source time function. For simple circular fault models, it has been found that this width is a very close estimate of the source diameter divided by the shear wave velocity:

$$\langle T \rangle = 2r/\beta.$$

From the source time function of Fig. 12 we find $\langle T \rangle = 16$ s; and for a shear wave velocity of $4.6\,\mathrm{km\,s^{-1}}$, appropriate to a source depth of 40 km, we find a source radius $r = 37$ km which is practically the same as found from the aftershock zone. Using the seismic moment $M_0 = 5.6 \times 10^{27}$ dyne cm and assuming a rigidity $\mu = 7 \times 10^{11}\,\mathrm{dyne\,cm^{-2}}$ we find the average slip on the fault $D = 1.6$ m. In order to calculate the stress drop we assume again a circular fault and we find $\Delta\sigma = 38$ bar. This value is subject to a very large uncertainty since it depends on a very rough estimate of the source area and on the value assumed for the rigidity.

Estimating source parameters for the 1965 events is a much more difficult task since we have no way to calculate the source area from the aftershock distribution. The 1965 March 28 had only two aftershocks located by ISC in 1965 May. We are thus forced to use the source time function obtained from *P*-wave modelling in order to estimate source slip and stress drop. We find from Fig. 10, $\langle T \rangle = 6$ s, which with $\beta = 5.6\,\mathrm{km\,s^{-1}}$ at 70 km depth yields a radius $r = 17$ km. Using the same procedure as for the 1971 event we obtain an average slip $D = 1.6$ m and a stress drop $\Delta\sigma = 91$ bar. The last figures are subject to an unknown uncertainty, but they are well within the typical values cited by Kanamori & Anderson (1975) for intraplate events.

No attempt was made to obtain fault parameters for the smaller 1965 event, since the source time function is too short to be determined from WWSSN long-period data.

Tectonic implications

The detailed study we have carried out confirms the existence of both thrust earthquakes at shallow depth and deeper intraplate events near the Chilean coast. The first type of earthquake has a typical thrust mechanism associated with interplate slipping, while the deeper ones have an almost down-dip extensional mechanism. The overall picture is shown in cross-section in Fig. 14. These earthquakes took place within the transverse valley province of Chile where subduction occurs at a shallow angle as shown by several authors (Stauder 1973; Barazangi & Isacks 1976; Hanus & Vanek 1968). This region is similar to the Lima–Huancayo area in central Peru which has been the subject of an animated controversy regarding the dip of the Benioff zone. In our case there is little doubt that subduction is shallow: well-located events cluster under western Argentina between 110–150 km depth, at about 500 km from the trench axis. The southern end of the aftershock zone (see Fig. 1) is located near 33°, the boundary between the transverse valley geological province and the

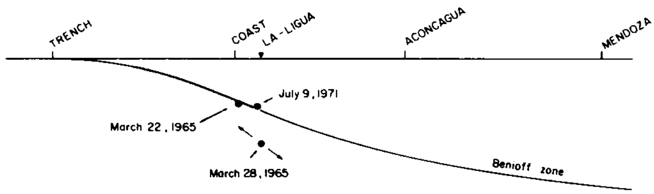

Figure 14. Schematic cross-section of the Aconcagua region in north-central Chile. A shallow dipping Benioff zone is shown, as proposed for this area by Barazangi & Isacks (1976). The 1971 July 9 and 1965 March 22 are underthrust events, while the deeper 1965 March 28 is an interplate extensional earthquake.

central valley region of south-central Chile. This line also marks the beginning of the southern line of active quaternary volcanoes. Barazangi & Isacks (1976) proposed that subduction south of 33° has a steeper dip, close to 30°. This Benioff zone is much more poorly defined since weak intermediate depth events occur rarely south of 33° S and their hypocentral determinations have large error margins.

We believe that there might be some connection between the pairing of intraplate extension and interplate thrust event in Chile and the recently discovered double Benioff zones in North Honshu (Hasegawa *et al.* 1978), Kurile (Stauder & Mualchin 1976; Vieth 1974) and Central Peru (Isacks & Barazangi 1977). Yet we have no clear evidence of intraplate reverse fault events in Chile. It is very likely that the two layers may not be resolved because of the poor precision of hypocentral determinations, due to the distribution of seismic stations and the lack of adequate models for the local structures. This leads systematically to large residuals at regional and local stations in the hypocentral determinations made by USGS and ISC. Another way to put in evidence the double layer would be through the fault-plane solutions of the larger events, since in the other places where the double layer exists the earthquakes in the upper layer have down-dip compressional mechanisms. The largest catalogue of fault plane solutions for Chile is that of Stauder (1973). In this catalogue we found the five normal fault intermediate depth coastal events at the left of Fig. 2. Most other events (14 in total) have underthrust mechanisms. No down-dip compression event that could be unequivocally associated with an upper layer of a double Benioff zone is found in Stauder's catalogue. There are, however, three shallow, slightly inland events that have horizontal pressure axes. Stauder himself proposed that these are shallow intra-continental plate events. The directions of their pressure axes are poorly constrained and it could also be that they occurred on the upper layer of a double Benioff zone. A decision is difficult at this point and we need some independent evidence to decide either way. The failure to identify the upper compressional events may also be simply attributed to smaller magnitude for such events, since only earthquakes of $M_s > 5.8$ allow unambiguous fault plane solutions. A study of recent events to complement Stauder's is being carried out; we hope this will help elucidate the presence of the upper compressional layer.

An alternative explanation of the pairing of extensional intraplate and interplate thrust events under north-central Chile was proposed by Abe (1972) for the 1966 and 1970 Lima, Peru earthquakes. Both Isacks & Molnar (1977) and Stauder (1973) showed that the downgoing slab under South America is under extension down to the intermediate depth gap in seismicity. We could then view the normal fault events as the uppermost spread of tensional stresses produced by the plate's own weight. Thrust earthquakes would occur at the interface between the oceanic and continental plate. The fact that thrust events occur only oceanwards from the Chilean coast indicates that the seismic interface between both plates is relatively narrow (only 200 km) and that deeper than about 40 km slip between the continental and oceanic lithosphere is taken up either by aseismic creep or by large-scale plastic deformation.

Conclusions

We have studied two large ($M_s \simeq 7.5$) earthquakes in the Aconcagua Province of Chile on 1965 March 28 and 1971 July 9. These events produced significant property and personal damage in their epicentral areas. They are of very different nature: the larger, shallower event of 1971 took place at the interface between the Nazca and South American plates and had the usual thrust mechanism associated with subduction. The 1965 event, which is deeper and has a normal fault mechanism, occurred inside the downgoing slab. It is very likely due to the tensional stresses that are know to prevail in the subducted plate under South America down to the intermediate depth seismicity gap. These two earthquakes occurred in one of the segments of the Peru–Chile trench that have been characterized by a shallow dipping Benioff zone.

An extensive analysis of the long-period body and surface waves was performed in order to refine their depth and fault plane solutions and to calculate their fault parameters. For the 1971 event the mechanism obtained from *P*-waves first motions and surface waves are in close agreement. Similarly, within the error of the determinations, the seismic moments found from *P*- and surface waves are identical and equal to 5.6×10^{27} dyne cm. The deeper, normal fault 1965 event is more complex; the seismic moments determined from *P*-waves and surface waves are 1×10^{27} dyne cm and 1.8×10^{27} dyne cm, respectively. *P*-wave first motions define only one steeply dipping nodal plane; Rayleigh and Love waves were then used to fix the shallow dipping plane. Modelling of *P*-waves confirms this mechanism and places the depth of this event at 72 km. From the *P*-waves we infer that this event is probably more complex than the simple model that we used, but we lack enough data to resolve this issue.

Both events had rather large stress drops. The 1971 event has a rather small area for a subduction zone thrust event of its moment; we found a stress drop of 38 bar. *P*-wave modelling and the coincidence of surface and body wave moments indicate that it probably represents a rapid, simple stress drop at the source. This event had a large aftershock series that lasted for several months. The 1965 event is intraplate and had only two minor aftershocks located in ISC in 1965 May. Having no information about the aftershock zone we interpret our *P*-wave modelling in terms of a simple circular fault model. The stress drop of 91 bar that we find is within the range of those found for other intraplate events.

The earthquakes of 1965 and 1971 are interesting also for the study of the seismicity of central Chile and the possible occurrence of a major shock in the Valparaiso–Santiago area (Kelleher 1972; McCann *et al.* 1979). The last great earthquake in this region dates from 1906 August 16. The return time for such great events is approximately 85 yr as determined from a historical record that is probably complete for the last 400 yr. McCann *et al.* (1979) classified the central Chile area as a category 2 seismic gap, i.e. as having a high seismic risk potential. They excluded the source area of the 1971 from the probable source area of the next great event in Valparaiso. A review of Montessus de Ballore's (1916) extensive study of the 1906 earthquake leaves no doubt that its source area included that of the 1971 event. Thus McCann *et al.*'s exclusion of the source area of the 1971 earthquake from the Valparaiso seismic gap is a plausible interpretation of the seismicity. Yet this area continues to have a high level of seismic activity and a relatively large event of $M_s = 6.7$ occurred on 1973 October 5 near the southern extreme of the 1971 aftershock zone, just north of Valparaiso. An alternative explanation of the 1971 and 1973 earthquakes is that they are foreshocks of a future great earthquake in this area, but this point requires additional study.

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