SOURCE PARAMETERS OF THE 1933 LONG BEACH EARTHQUAKE

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Abstract

Regional seismographic network and teleseismic data for the 1933 $(M_1 = 6.3)$ Long Beach earthquake sequence have been analyzed. Both the teleseismic focal mechanism of the main shock and the distribution of the aftershocks are consistent with the event having occurred on the Newport-Inglewood fault. The focal mechanism had a strike of 315°, dip of 80° to the northeast, and rake of -170° . Relocation of the foreshockmain shock-aftershock sequence using modern events as fixed reference events, shows that the rupture initiated near the Huntington Beach-Newport Beach City boundary and extended unilaterally to the northwest to a distance of 13 to 16 km. The centroidal depth was 10 \pm 2 km. The total source duration was 5 sec, and the seismic moment was $5 * 10^{25}$ dyne-cm, which corresponds to an energy magnitude of $M_w =$ 6.4. The source radius is estimated to have been 6.6 to 7.9 km, which corresponds to a Brune stress drop of 44 to 76 bars. Both the spatial distribution of aftershocks and inversion for the source time function suggest that the earthquake may have consisted of at least two subevents. When the slip estimate from the seismic moment of 85 to 120 cm is compared with the long-term geological slip rate of 0.1 to 1.0 mm / yr along the Newport-Inglewood fault, the 1933 earthquake has a repeat time on the order of a few thousand years.

INTRODUCTION

The 1933 Long Beach earthquake was the second largest earthquake (after the 1971 San Fernando earthquake) to strike the Los Angeles area in the twentieth century. It caused extensive damage in the greater Los Angeles area and in particular, throughout the southern part of the Los Angeles basin (Fig. 1). Many hundreds of people were injured and 120 people died (Wood, 1933). This widespread damage was in part caused by the lack of a seismic safety element in building codes (Richter, 1958). The Richter magnitude scale had not been devised at this time, and Wood (1933) described the main shock not as a great earthquake but as "a fairly strong, moderately large local shock." Richter (1935) assigned a magnitude of about 6.2 to the main shock by comparing the recorded amplitudes at Tinemaha and Haiwee in eastern California with the amplitudes of a $M_L = 5.5$ aftershock. Later, Richter (1958) refers to the more commonly known value of $M_L = 6.3$ of the 1933 earthquake.

In the late 1920s, the Caltech seismographic network with seven stations was installed to monitor earthquake activity in southern California (Wood, 1933). The 1933 Long Beach earthquake was the first damaging large earthquake to occur within this network and was well recorded. In addition, the main shock was recorded by three strong motion accelerographs, one in Long Beach and two in downtown Los Angeles (Heck, 1933; Heck and Neumann, 1933). The earth-

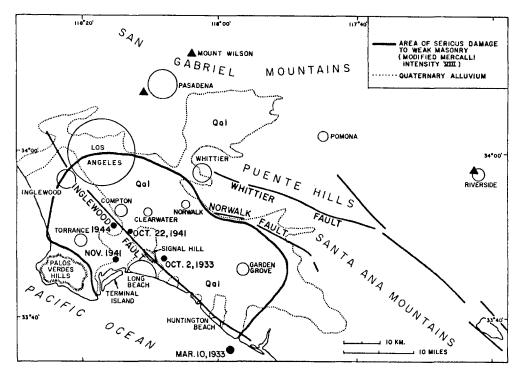


Fig. 1. Map of the Los Angeles and Orange County areas shaken by the 1933 Long Beach earthquake. Important epicenters (solid circles) and the Modified Mercalli intensity line of VIII are also shown. Dotted curve outlines the Quaternary alluvium. Solid triangles indicate the location of seismograph stations. Adopted from Richter (1958) with modifications from Barrows (1974).

quake was also recorded at teleseismic distances by long-period seismographs in operation throughout the world and, in particular, in the United States and many European countries.

The 1933 Long Beach earthquake has not been studied in as much detail as some other similar damaging earthquakes, such as the 1971 San Fernando earthquake. This paper reanalyzes some of these old regional and teleseismic data using modern techniques to enhance our understanding of this sequence and to confirm or negate previous findings. Newer and more advanced techniques make it possible to resolve more of the temporal and spatial details of the earthquake faulting and to determine or refine the existing estimates of source parameters such as focal mechanism, seismic moment, and centroidal depth for the main shock.

At the time of the earthquake, epicenters were calculated by hand. Because the volume of data was large, many of the aftershocks that occurred during the first few weeks were arbitrarily assigned identical epicenters (Fig. 2; Wood, 1933; C. F. Richter, unpublished data, 1933). Because one reliable felt report was available from south Long Beach, most of the aftershocks were assigned this location (Wood, 1933). It is therefore impossible to evaluate the spatial extent of the aftershock zone or possible spatial clustering of the aftershocks from these locations. The first objective of this study has been to remedy this problem by rereading the old seismograms and relocating the aftershocks.

The second major objective of this study is to determine the teleseismic focal mechanism of the main shock. The source mechanism of the 1933 earthquake

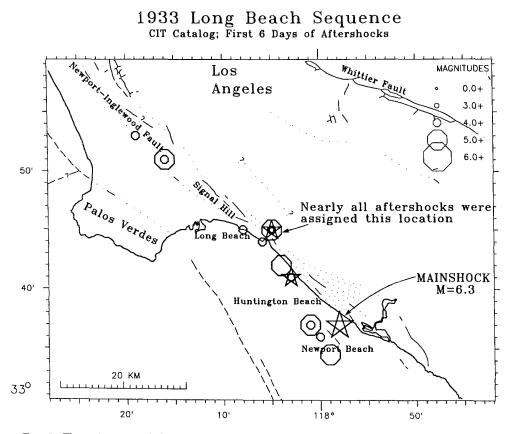


FIG. 2. The epicenters of the main shock and first 6 days of aftershocks as published in the Caltech Earthquake Catalog (Hileman *et al.*, 1973). Events with $M \ge 5.3$ are shown as stars and events with M < 5.3 are shown as open circles. Most of the events are assigned one location in southern Long Beach.

was first studied by Benioff (1938), who used the elastic rebound theory to show that the source was not a point source but had a finite spatial extent. He also reinterpreted the azimuthal distribution of tombstones, as determined by Clements (1933), to show that the faulting was strike-slip. Barrows (1974) reviewed the geological and seismological aspects of the earthquake and argued that the event occurred on the Newport-Inglewood fault (NIF). In a more recent study, Woodward-Clyde (1979) determined a strike-slip focal mechanism that was not consistent with the earthquake being caused by the Newport-Inglewood fault, unless some complex effects of the three-dimensional velocity structure of the Los Angeles Basin were included.

Recent studies of the seismotectonics of the Los Angeles basin (Hauksson, 1987, 1990) and compressional structures such as folds (Davis *et al.*, 1989) suggest that thrust faulting in the Los Angeles basin is more common than previously thought. Recently, Suppe (1989) suggested that the NIF has a significant thrust faulting component. Therefore it is important to reevaluate the data from the 1933 Long Beach earthquake to investigate whether it had a significant thrust faulting component.

The focal mechanism is also used to determine other source parameters, such as seismic moment and stress drop. The seismic moment estimated by Thatcher

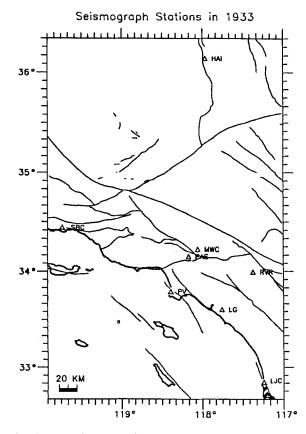


FIG. 3. The Caltech seismographic network that began operating in the late 1920s. The stations PV and LG were temporary sites were one portable station was deployed to record the 1933 aftershocks.

and Hanks (1973) using a single station determination at Pasadena (PAS) and by Woodward-Clyde (1979) using teleseismic waveforms are reevaluated using the focal mechanism determined in this study. The average stress drop is determined using the seismic moment and the source radius derived from the source time function.

EARTHQUAKE LOCATIONS

Seismographic Network

The Caltech seismographic network in operation at the time of the 1933 earthquake is shown in Figure 3. Richter (unpublished data, 1933) used data from these stations to determine the epicenters of the earthquake and some of the aftershocks. Each station was equipped with both vertical and horizontal seismometers and photographic recording. Two stations, PV and LG, were temporary sites occupied with a portable station for a few days at each site (C. F. Richter, unpublished notes, 1933).

Time Corrections

To relocate the aftershocks, the P and S arrival times for 94 events were reread from the photographic records. An additional 12 events, with four arrival times each, that C. F. Richter had read himself and listed in his unpublished data were also included in the dataset. The P and S arrival times measured off the records were corrected to a common time base by adding time corrections read from radio records for each station. These radio records are the same size as a seismogram and have the same time ticks generated by the local clock, but they are attached to a short-wave radio that received signals in morse code. Richter determined detailed time corrections for two of the stations, Mount Wilson (MWC) and Riverside (RVR). For these stations, his time corrections were used, which are documented in his "Long Beach" notebook (C. F. Richter, unpublished data, 1933). Time corrections were made by finding several (about eight per day) readily identifiable points in the radio records on all the stations. These were measured on all the stations and marked so that their identity could be checked later if there were any doubt. Time at Pasadena was used as the common time base, and the time corrections for other stations were found by subtracting the other station's time from Pasadena time.

The 1933 earthquakes were located using the P and S arrival times from the original photographic paper records and adding the above determined time corrections. Six of the seven available stations were read. The portable stations, PV and LG, provided 11 and 5 arrival times, respectively. Tinemaha in north-eastern California is very distant and had excess static on its radio record, so it was not included. Haiwee was also very distant and the clock drift was so large that these readings were subject to additional timing error. The Haiwee readings were used with the others but were nearly always given a low weight. This left five generally useful stations with a maximum of 10 P and S arrival times per event. The Santa Barbara (SBC) records also seemed to be subject to additional timing errors caused by irregular drum speed, so readings were in some cases given a lower weight.

Reference Events

Two seismographic networks with closely spaced stations (the University of Southern California, Los Angeles Basin seismographic network and the USGS and Caltech seismographic network) have been in operation in and adjacent to the Los Angeles basin since the early 1970s. Nine recent local earthquakes that had been located by the dense networks were chosen as reference events (Fig. 4). The VELEST code (Roecker and Ellsworth, 1978) was used to invert simultaneously for the hypocentral parameters, velocity models, and station delays. The reference events were treated as blasts with fixed epicenters in the VELEST code to calculate absolute station delays. Both P and S arrival times of 96 events from 1933 were included in the inversion.

Only a limited set of arrival times was available, with most stations located more than one depth distance away from the aftershock zone. A simple velocity model consisting of a layer over a half-space was therefore chosen with a 32-km-thick layer of 6 km/sec over a half-space of 7.8 km/sec. This velocity model was kept fixed in the VELEST inversion with a large damping factor. This velocity model and the station delays determined by VELEST were used as input for HYPOINVERSE (Klein, 1985) to calculate final locations for the 106 earthquakes, which included the 12 events with four arrival times each from C. F. Richter. The depths of all the events were allowed to vary to see whether or not the depth distribution of the aftershocks could be resolved.

Station delays were determined for both the old and new stations. Data from both the 1933 and the reference events were used in the inversion. For the five

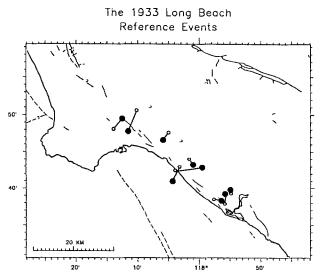


FIG. 4. The true epicenters (from the modern seismographic network) of nine reference events (1978 to 1989) are shown as open circles on this map. These events were relocated using the 1933 network with the new delays and the simplified velocity model. These redetermined epicenters, shown as solid circles, indicate the mislocations that may be present in the 1933 epicenters.

stations still in operation, absolute station delays were calculated. For the three stations no longer in operation, when the reference events were recorded, the station delay is only a relative delay, referred to the velocity model. For all three stations (LJC, PV, and LG), a modern station is located near the location of the discontinued old station, so the travel time to it serves as an additional constraint on the travel time to the nearby old station.

The quality of the station delays can be tested by relocating the reference events. The true (modern network) epicenters of the reference events are shown as open circles in Figure 4. Using only arrivals at the 1933 stations, the simplified velocity model, and the station corrections, the relocated epicenters of the reference events are shown as solid circles in Figure 4. There does not appear to be a systematic mislocation error because the events move in different directions. The north-south mislocations however, appear to be smaller than the east-west mislocations. The largest mislocation vector is approximately 7 km, although most of them are in the range of 2 to 4 km. The relocations of the reference events suggest that, even for the sparse station distribution available in 1933, the data can still be used to obtain locations without systematic mislocations with average mislocations caused by scatter in the data in the range of 2 to 4 km. This scatter in the data, to a large extent caused by the sparse distribution of stations, cannot be easily reduced.

The depths of the reference events changed less than 2 to 5 km for seven of the nine events. The depths of two events that had the largest change in epicentral location changed 7 and 8 km, respectively. Therefore, although the depths are poorly resolved compared with depths determined with data from modern seismographic networks, it is possible to determine if they are all shallow (depths less than 10 km) or all deep (depths greater than 10 km).

Final Locations

The relocated hypocenters of the 1933 foreshock-main shock-aftershock sequence are shown in Figure 5. The first six days of aftershocks of M > 4 are

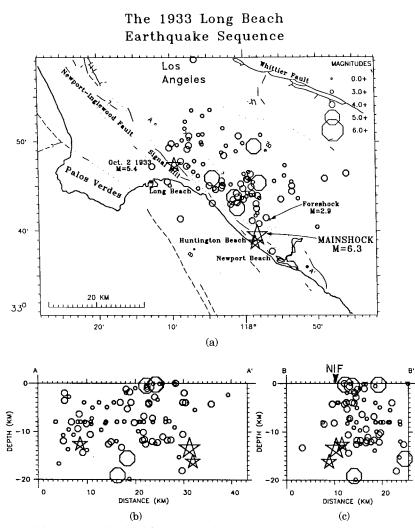


FIG. 5. (a) Map showing the final locations of the 1933 foreshock-main shock-aftershock sequence recorded during the first 6 days of aftershock activity. (b) A cross section A-A' along strike of the Newport-Inglewood fault. (c) A cross section B-B' normal to the strike of the Newport-Inglewood fault.

shown with several smaller events in the M = 3 to 4 range also included. The smaller events were included because C. F. Richter had already read the arrival times. The main shock location at the southernmost extent of the aftershock zone suggests unilateral rupture, beginning in Huntington Beach and propagating to the northwest toward Long Beach.

The main shock was preceded by a M = 2.9 foreshock (mentioned as a magnitude 4 foreshock by Richter, 1958) by about 1.5 days. The foreshock was located within a few kilometers of the main shock. The offset in location between the foreshock and main shock is within the range of uncertainty in the data. This pattern is consistently observed for foreshocks in California (Jones, 1984).

Although a very simple velocity model is used and only a few of the aftershocks have data from the nearby portable stations, the overall pattern in the

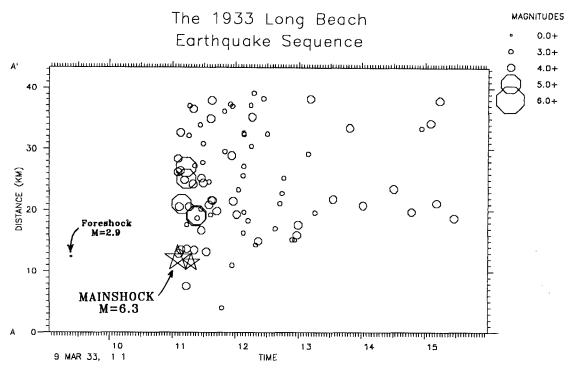


Fig. 6. Distance versus time for the first 6 days of aftershock activity. The distance is measured along the line A-A' in Figure 5.

spatial distribution appears to be real. Some of the details, however, are artifacts. As noted above, the distribution of aftershocks is better constrained in the north-south direction than in the east-west direction (Fig. 4). The east-west scatter may be caused by the difficulties with determining time corrections for the Santa Barbara (SBC) station and the sparse distribution of stations as discussed above.

The cross section A-A' shows the depth distribution of the main shock and aftershocks along the fault (Fig. 5b). The main shock hypocenter is located at the southeast end near the bottom of the distribution. The aftershocks are scattered in the depth range of 0 to 20 km as opposed to being clustered near the surface or near the lower end of the seismogenic zone.

The cross section B-B' is taken normal to the fault (Fig. 5c). It is difficult to determine a dip for the distribution of aftershocks. Although the depths are not constrained, if the $M \ge 5.3$ aftershocks that are represented by stars are excluded, the systematic location of epicenters to the east of the surface trace of the NIF can be interpreted as being either vertical or having a steep dip to the east. The $M \ge 5.3$ events may be systematically mislocated differently than the smaller events, because their seismograms are usually clipped and no S arrival times are available.

Distribution in Space and Time

This space-time diagram taken parallel to the strike of the fault shows that the initial aftershock zone was about 13 to 16 km long and extended within hours to 20 to 25 km length (Fig. 6). The initial length of the aftershock zone is

Origin	Time	Latitude	Longitude	Depth	Magnitude	Focal Mechanisms			Mo
Day	(UT)	(N)	(W)	(km)	M _L	Ddir	Dip	Rake	(dyne-cm)
11 March 1933	0154	33°39.54′	117°58.30′	13	6.3	45°	80°	-170°	$5*10^{25}$

 TABLE 1

 Location and Focal Mechanism of the 1933 Long Beach Earthquake

an approximate measure of the length of the main shock rupture. The distribution of aftershocks relative to the foreshock and main shock suggests that the rupture was unilateral and extended from the main shock hypocenter toward the northwest along the Newport-Inglewood fault. Two clusters of aftershocks can be seen. The first is 7 to 9 km northwest of the main shock and the second 13 to 16 km. This distribution of aftershocks can be interpreted as two asperities, one 6 to 8 km long between main shock and first cluster, and the other 3 to 4 km long between the first and second cluster. These asperities may accommodate most of the slip during the main shock rupture.

The aftershock zone extended from the Huntington Beach-Newport Beach City boundary into Long Beach. Arrival times by Richter (unpublished notes, 1933) were used to relocate the 2 October ($M_L = 5.4$) aftershock that defines the northernmost extent of the aftershock zone. At the time of this earthquake, it was classified as a nonaftershock or an independent main shock by Benioff (1951), because it was followed by its own aftershock sequence and was located too far away from the 1933 ($M_L = 6.3$) Long Beach earthquake and its aftershocks. The relocated aftershocks in Figure 5a show that the 2 October event occurred near the northwestern edge of the aftershock zone but outside the rupture area of the main shock. It thus extended the aftershock zone and should be classified as a late large aftershock as Richter (1958) suggested.

Source Parameters of the Main Shock

The main shock hypocenter is located near the Huntington and Newport Beach City boundary (Table 1; Fig. 5). This location is similar to the original location by Wood (1933) who located it offshore 4 to 6 km to the south of the new location. The main shock hypocentral depth determined with HYPOINVERSE and based on the six local arrival times was 13 km.

Focal Mechanism

The focal mechanism of the main shock was determined using teleseismic waveforms in the distance range of 26° to 86° and the waveform modeling technique and computer programs by Nábělek (1984, 1985). The instrument response of the Galitzin seismographs operated at most of the European stations, from McComb and West (1931) and Kanamori (1988), are listed in Table 2. The instrument response was unknown at Buffalo (BUF) but was assumed to be a Galitzin as is shown in Table 2.

The same velocity model consisting of a layer over a half-space was used here to get results comparable with the aftershock locations. A time function with a 5 sec duration (with a 2 sec rise time and a 3 sec decay time) was found by trial and error to provide the best fit to the data. The waveforms were then fit by forward modeling by stepping through ranges of azimuth, dip, rake, depth, and seismic moment. The best model fit gave the smallest normalized root-meansquare error in the fit between the observed and calculated waveforms.

_			. .	Pendulum Period (sec)			Weight		
Station	Azimuth (°)	Distribution (°)	Instrument Type	Z	N	E Comp.	Р	SV	SH
SIT	338	26.0	Wenner	12			0.3		
BUF	61	31.8	Galitzin*	12			1.0		
KEW	35	78.9	Galitzin	25	25	25	0.3	0.0	0.5
DEB	31	81.0	Galitzin	12	25	25	1.0	0.0	0.8
STR	32	84.7	Galitzin	12			0.3		
STU	32	85.2	Galitzin	12	12	12	0.3	0.0	0.3

TABLE 2
STATION PARAMETERS

*Assumed instrument type.

The best fitting focal mechanism shows right-lateral strike-slip motion with a minor normal component (Fig. 7). The best constrained nodal plane has a strike of 315° , dip of 80° to the northeast, and rake of -170° . The strike and dip of this nodal plane are similar to the strike and dip of the NIF (Barrows, 1974; Wright, 1990).

The azimuth and dip of this plane are constrained to within 10° and 5° , respectively, by the nodal waveforms recorded by the European stations (KEW, DEB, STR, AND STU). The *P* waveform at Buffalo, New York (BUF), provides additional constraints on the dip of this plane, although a small fraction of the first *P*-wave pulse is lost in a blank time mark on the original record. The auxiliary nodal plane is less well constrained. Some tradeoff in terms of depth versus rake exists, so that a shallower centroidal depth requires a larger normal

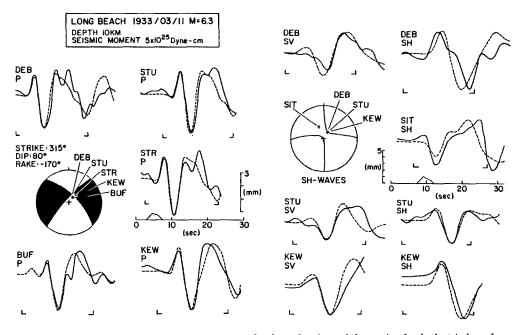


FIG. 7. (Left) The lower-hemisphere P-wave focal mechanism of the main shock that is based on modeling of teleseismic waveforms shows strike-slip faulting. P, SH, and SV waveforms are drawn with solid lines and calculated synthetic seismograms are drawn with dashed lines. The brackets indicate the part of the waveform used in the inversion. (Right) The corresponding SH-wave focal mechanism. The best fitting average source time function is also shown.

component in the mechanism. Because of the limited azimuthal distribution of the data in the distance range of 30° to 90° , the *SH* wave form Sitka, Alaska (SIT), at a distance of 26° was also included with a low weight, because it may be contaminated by upper mantle reflections.

Woodward-Clyde (1979) determined a focal mechanism for the main shock based on first motions and forward modeling of teleseismic waveforms. They constrained the nodal planes with first motion polarities of upgoing phases such as pP and sS and first arrivals recorded at regional and teleseismic distances. They had 6 out of 18 inconsistent first motion polarities. They found an almost pure right-lateral focal mechanism with one nodal plane striking N22°W. This is 23° away from the strike of the NIF. They proposed that dipping layers in the Los Angeles basin sediments caused a change in the take-off azimuth by as much as 18° to explain the difference in strike between the nodal plane and the NIF. The focal mechanism of this study differs from the Woodward-Clyde focal mechanism because it is based on fitting the whole teleseismic waveforms and not individual first motion polarities. This study shows that a strike of N45°W fits the data best and the explanation of dipping layers is not needed.

Centroidal Depth

The centroidal depth of 10 ± 2 km is fairly deep for a southern California strike-slip earthquake. The simple velocity structure used here may bias the centroidal depth to slightly greater depths than the real depth. The hypocentral depth of 13 km and the centroidal depth of 10 km are similar, although both should be considered as an order of magnitude estimates based on only a few data points. No reliable surface rupture was reported (Wood, 1933). This also suggests that the hypocenter of the main shock was probably fairly deep.

Seismic Moment and Magnitude

The best fitting seismic moment determined from the teleseismic records was $5 * 10^{25}$ dyne-cm. The weight of the different components in the analysis are listed in Table 2. Copies of the original seismograms from DEB were available and, because the data are of high quality, the *P* wave was assigned a weight of 1.0. The *SH* waves from DEB and *P* and *SH* waves from other European stations were assigned a weight of 0.3 if they were included in the analysis. These low weights were chosen to account for the redundancy in data from the European stations. The *SV* waves were in general given a weight of zero.

Thatcher and Hanks (1973) estimated a seismic moment from the Pasadena records and found a value of 2×10^{25} dyne-cm. Woodward-Clyde (1979) found a seismic moment of 4 to 6×10^{25} dyne-cm using the same data set as was used in this study, which is similar to the seismic moment of 5×10^{25} dyne-cm found in this study. The seismic moment is equivalent to $M_W = 6.4$ using the Hanks and Kanamori (1979) moment magnitude relationship. This magnitude is similar to the published local magnitude of $M_L = 6.3$ by Richter (1958) and the $M_S = 6.25$ published by Gutenberg and Richter (1954). The 1933 Long Beach earthquake is thus the second largest, with the 1971 ($M_W = 6.6$) San Fernando earthquake being the largest, to have hit the greater Los Angeles area in this century.

Source Time Function

The source time function that fit the teleseismic data best has a duration of 5 sec. This is consistent with observations at the time of the earthquake by Wood

(1933) who reported strong shaking of 5 to 10 sec duration in Pasadena. In addition, the accelerograms recorded at distances of 27, 48, and 55 km show strong ground shaking of 4 to 6 sec duration (Heck, 1933; Heck and Neumann, 1933; Heck, 1934). The peak accelerations measured range from 0.2 to 0.4 g. This, however, is approximately the saturation level for these instruments. The 5 sec source duration is also consistent with the relationship between source duration versus seismic moment reported for a world-wide set of earthquakes by Ekstrom *et al.* (1990).

The method of Nábělek (1984, 1985) was then used to invert for the shape of the source time function and the centroidal depth using the above results as an initial model. The depth changed less than 0.5 km, while the source time function tended to split into two triangular shaped time functions, the first with a 3 sec duration and the second with a 2 sec duration and half the amplitude of the first.

Because the number of waveforms is limited, this result must be considered to be tentative at best. The two time functions do not improve the visual fit to the waveforms, although they reduce the overall root-mean-square error of the fit. In most cases only the first 20 sec of the waveform are used in the inversion, so the effect of possible instrument back swing, which may occur beyond 15 sec, should be minimized.

The two source time functions and the distribution of aftershocks are consistent with the interpretation that the main shock consisted of at least two subevents. Such multiple subevents are commonly reported for moderate-sized or large earthquakes in California. For instance, Pacheco and Nábělek (1988) used a similar approach to identify large subevents or asperities within the main shock rupture of three moderate-sized California earthquakes that occurred in 1986. This is also similar to the interpretations of Mendoza and Hartzell (1988), who argued that aftershocks occur outside areas of large slip, i.e., asperity, during the main shock and represent the continuation of slip in the outer regions of the main rupture or the activation of subsidiary faults.

Stress Drop

To estimate the stress drop of the 1933 earthquake, the approach of Pacheco and Nábělek (1988) is used. The source radius (r) is estimated from the characteristic time (τ_s) . That is, the time it takes to release half the seismic moment (M_0) as determined from the source time function. If the rupture is circular and propagates at a velocity of:

$$v_r=0.75\beta,$$

where, an average S-wave velocity $\beta = 3.5$ km/sec is assumed. The source radius is

$$r = v_r \tau_s$$

Using the first and second equation and the characteristic time of 2.5 to 3 sec, the source radius is 6.6 to 7.9 km. The Brune (1970) stress drop is

$$\Delta\sigma=\frac{7M_0}{16r^3},$$

where M_0 is the seismic moment of $5 * 10^{25}$ dyne-cm. For a source radius of 6.6 to 7.9 km, which matches the 13 to 16 km length of the aftershock zone, the

average stress drop is 44 to 76 bars. This stress drop falls within the 30 to 100 bars range of stress drops observed for most California earthquakes, which have occurred on both high and low slip rate faults (Hanks, 1979).

This stress drop is higher than found by Pacheco and Nábělek (1988) who determined a stress drop of 27 to 41 bars for the 1986 North Palm Springs earthquake, which showed strike-slip faulting on the San Andreas fault. Kanamori and Allen (1986) argued that earthquakes with larger repeat time have shorter fault lengths or higher stress drops than events with small repeat time. The 1933 Long beach earthquake falls in the range of events with repeat time of 300 to 2000 years as grouped by Kanamori and Allen (1986).

Slip

The average slip in the 1933 earthquake can be determined from the relation

$$M_0 = \mu \pi r^2 D,$$

where M_0 is seismic moment, μ is crustal rigidity, r is source radius, and D is slip. The average slip ranges from 85 to 120 cm for source radii of 6.6 to 7.9 km. If the long term geological slip rate along the NIF is 0.1 to 1.0 mm/yr (Ziony and Yerkes, 1985), the average return time of the 1933 Long Beach earthquake along the same segment of the NIF is on the order of several millennia.

DISCUSSION

Faulting Along the Newport-Inglewood Fault

Both the focal mechanism of the 1933 main shock and the spatial distribution of aftershocks indicate that the earthquake occurred on the NIF. The NIF forms a major basement boundary between the metamorphic basement terrane of the Continental Borderland to the west and the slightly metamorphosed sediments and plutonic and volcanic rocks to the east (Wright, 1990). It also is a part of a system of plate boundary faults within the San Andreas fault system that accommodates motion between the Pacific and North-American plates. The total geological right-lateral offset along the NIF since middle-Miocene time is 3 km (Yeats, 1973). A series of low lying hills, en-echelon fault strands, and numerous oil fields located adjacent to the NIF make the subsurface and surface trace very prominent. This prominent surface expression may be a manifestation of the basement boundary rather than being primarily caused by the right-lateral offset (Fig. 8).

Previous investigators have often described the fault as a classic example of folds and faults along a deep-seated, strike-slip fault (e.g., Wilcox *et al.*, 1973). From Long Beach to the southeast until it heads offshore near Newport Beach, the NIF appears as a near continuous single strand. To the northwest of Long Beach, it consists of several en-echelon fault strands. Scattered background seismicity is observed along the whole length of the NIF (Hauksson, 1987). Recently, however, more detailed data from oil fields in the Los Angeles basin show a more complex picture suggesting that most geological structures adjacent to the NIF are not secondary features resulting from wrench faulting (Wright, 1990) but are rather primary structures resulting from north-south compression of the basin (Hauksson, 1990). The north-south compression therefore causes both strike-slip and thrust faulting in the Los Angeles basin. This

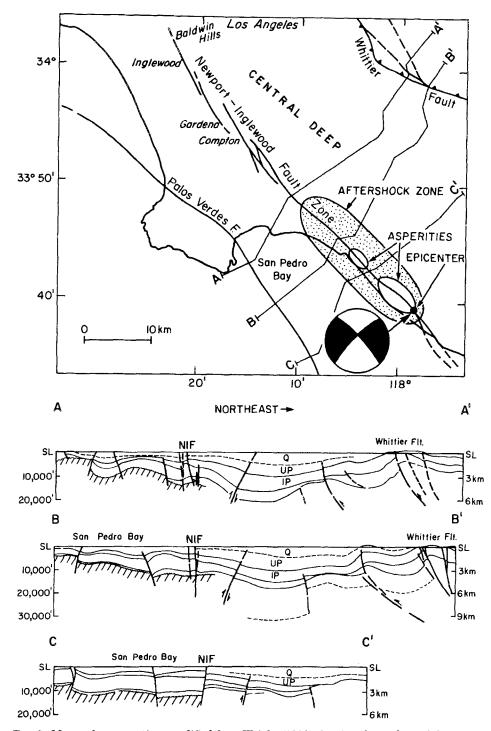


FIG. 8. Map and cross sections modified from Wright (1990) showing the geology of the area and three cross sections across the Newport-Inglewood fault (NIF). The dark shaded area shows the location of the region of high aftershock activity. The two blank areas within the shading show the location of the two asperities. The cross section (A-A') crosses Long Beach and Signal Hill. The cross section (B-B') crosses Seal Beach. The cross section (C-C') crosses Sunset Beach. Both cross section B-B' and C-C' cross the NIF within the aftershock zone. Q: late Quaternary alluvium; uP: upper Pliocene sediments; IP: lower Pliocene sediments.

observation was explained with a slip partitioning model by Hauksson (1990), where strike-slip faulting on vertical faults and thrust faulting on gently dipping faults replace a system of oblique faulting. The almost pure strike-slip mechanism of the 1933 earthquake and the pure thrust faulting mechanism for the 1987 ($M_L = 5.9$) Whittier Narrows earthquake are consistent with this slip partitioning model.

The three cross sections in Figure 8 (Wright, 1990) show the trace of the NIF as mapped from oil well data. As seen in these cross sections, the NIF dips steeply to the northeast near Long Beach (line A-A') and rotates to vertical near Seal Beach (B-B'). At Sunset Beach (C-C') the dip is steeply to the southwest. The two southernmost cross sections show that a small normal component with the northeast side down exists along the NIF. Given the uncertainty in the seismological and the geological data, the difference in dip of the NIF and the nodal plane of the focal mechanism is not significant. The existence of a small normal component in the mechanism and in the geological cross sections suggests that the southwestern block of the Los Angeles basin is still subsiding.

State of Stress

The state of stress in the greater Los Angeles basin was determined by inverting data from focal mechanisms of local earthquakes (Hauksson, 1990). The stress state near the aftershock zone of the 1933 earthquake is consistent with strike-slip faulting with the intermediate principal stress being vertical. The maximum principal stress is horizontal and forms a high angle, 70° to 90°, with the NIF suggesting that the NIF is weak. Similar refractions of the stress field in central California are observed near the San Andreas fault (Mount and Suppe, 1987). The NIF thus appears to show rheological behavior similar to the San Andreas fault in central California.

What Happened in the 1933 Earthquake?

To mitigate the earthquake hazards in Long Beach and surrounding areas, it is important to identify what segment of the NIF slipped in the 1933 Long Beach earthquake. The location of the main shock hypocenter shows that the rupture initiated along a fairly smooth section of the fault near the Huntington Beach and Newport Beach City boundaries. The main shock ruptured the fault toward the northwest and can be interpreted as having consisted of two subevents or as having ruptured at least two asperities (Fig. 8). The distribution of aftershocks and the source time function suggest that the first asperity had a diameter of 6 to 8 km and its rupture released approximately 75 per cent of the seismic moment. The rupture of the second asperity that is offset 2 to 3 km further to the northwest and had a diameter of 3 to 4 km accounts for the remaining moment release. Thus, a 13 to 16 km long segment of the NIF accounted for the bulk of the moment release in the 1933 earthquake. Aftershock activity grew rapidly in two locations, the first being between the two asperities and the second to the northwest of the end of the main shock rupture, extended the aftershock zone almost across the City of Long Beach.

The name of the 1933 earthquake, "The Long Beach Earthquake" was adopted because of the extensive damage in the City of Long Beach (Wood, 1933). This damage was mostly caused by soft near-surface ground conditions and possibly by directivity effects in the radiation pattern. This name is somewhat misleading because the 1933 earthquake probably did not release any significant shear stress stored along the NIF within the City of Long Beach. The main shock caused 85 to 120 cm of slip at depth along its rupture surface to the south of Long Beach. Using the magnitude-seismic moment scaling relations of Thatcher and Hanks (1973), the largest aftershock on 2 October located in the City of Long Beach slipped less than 10 per cent of the total main shock slip. The earthquake hazards in Long Beach thus should be considered to be similar to elsewhere along the NIF to the northwest of the 1933 segment.

Earthquake Hazards Implications

Sibson (1989) presented a model of faulting which suggests that large earthquakes on strike-slip faults preferentially stop at dilational jogs, because no mechanism is available for rapid transfer of stress from one fault strand to the next. By this model, an earthquake along the NIF starting in Long Beach and propagating northward would quickly run into a dilational jog and hence stop, because the strike of the fault is changing to a more northerly direction. In contrast, an earthquake rupture starting in the Baldwin Hills would see compressional bends as long as it continued to rupture to the southeast.

The next earthquake that will cause significant damage in Long Beach could occur on either the NIF or on concealed faults offshore, collectively called the Torrance-Wilmington thrust belt (Hauksson, 1990). If this future earthquake were to occur on the NIF, it will probably not start in Long Beach or further to the south if Sibson's (1989) model is correct. It would more likely start to the northwest, possibly as far away as the Baldwin Hills. Such a southeast-directed rupture would probably propagate through Inglewood, Gardena, Compton, and Long Beach, then terminate near the south edge of Long Beach. Similar to effects of the 1933 event, a southeast-propagating rupture would also cause an amplification of strong ground motions in the Long Beach area because of soil types and the directivity effect.

Ziony *et al.* (1985) evaluated seismological and geological effects of a postulated M = 6.5 earthquake on the NIF with rupture initiating just to the north of Signal Hill and extending 30 km to the north across the Baldwin Hills. Their modeling predicted amplification of strong ground motions near the northern end of the NIF caused by the directivity effects of the rupture propagation. They also predicted surface rupture along 16 km of exposed late Quaternary faults, tectonic elevation changes, and liquefaction-related ground failure. Many of these same earthquake hazards would also be caused by an earthquake starting to the north, although the spatial distribution of the effects would be different.

CONCLUSIONS

The 1933 ($M_W = 6.4$) Long Beach earthquake showed right-lateral motion along the NIF with a small normal component. Slip along the NIF thus contributes to the relative plate motion between the Pacific and North-American plates. The normal component indicates continued subsidence of the southwest corner of the Los Angeles basin. The absence of a thrust component is consistent with the slip partitioning model of the seismotectonics of the Los Angeles basin by Hauksson (1990). The main shock was a factor of 2 smaller in seismic moment than the 1971 ($M_W = 6.6$) San Fernando earthquake, not a factor of 5 smaller as previously thought. The distribution of aftershocks and the source time function suggest that the main shock may have consisted of two subevents. It released strain along a 13 to 16 km long segment of the NIF from Newport Beach to the southern edge of Long Beach. Accumulated strain along the section of the NIF within the City of Long Beach thus may not have been released, even though this segment of the fault experienced a relatively high level of aftershock activity.

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