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ANNUAL PROGRESS REPORT

TO THE

NATIONAL AERONAUTICS AND SPACE ADMINISTRATION

ON

GRANT NGR 33-008-146

"Strain Buildup and Release, Earthquake Prediction and Selection of VLB Sites for the Margins of the North Pacific"

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SUMMARY

Field research supported by contract NGR 33-008-146 in 1975 was directed toward detailed microearthquake studies of the southern section of the San Andreas fault system in order to elucidate the role of known but little studied complications in the fault system which could affect the SAFE measurement. Data reduction from contract supported microearthquake studies in Baja California, Mexico, was completed and a report has been submitted to the Bulletin of the Seismological Society of America (a copy is attached as Appendix A).

A paper (Appendix B) comparing styles of deformation on the Alpine fault in New Zealand and on the San Andreas fault was submitted to the Journal of the Geological Society (London). Sections of these faults with comparable physical characteristics seem to deform in a similar manner, arguing that deformation style is controlled by certain fundamental relations and will continue in the same manner in the future.

A brief report discussing the interpretation of geology in eastern New Guinea was submitted to Geology (Appendix C). Deformation in the area is proposed to be related to collision of a series of island arcs with central New Guinea, which is part of the Australian continent. Among other unusual properties, eastern New Guinea contains the highest anomaly of the gravimetrically determined geoid based on GEM 6 and $1^0 \times 1^0$ surface gravity data (Marsh and Vincent, 1974).

Research Progress

A major part of the research effort involved a microearthquake study of faults located east of the San Andreas fault in the Los Angeles area during July and August, 1975. Data reduction is completed and a report detailing results is in preparation. Seismicity shows that deformation generally consistent with regional right-lateral shear presently occurs over a broad part of the western

United States. Some of the motion between the North American and Pacific plates may be absorbed by this regional shear. South of the Los Angeles area few earthquakes have been located east of the San Andreas suggesting that the plate boundary may change to a broad zone of deformation approximately north of Los Angeles.

The area studied is in the San Bernardino Mountains which form a continuation of the Transverse Ranges east of the San Andreas. Geology of the area is complex as to the south the San Andreas splits into two branches while the "big bend" lies to the north. In the big bend area the fault trace is oblique to the relative plate motion so a component of compression is generated. This area contains the anomalous aseismic uplift recently detected (Castle and others, 1976). Faults with a trend generally parallel to the San Andreas are mapped to the east of it. The function of these faults in tectonics of the region or even if they are presently active is uncertain. Preliminary interpretaitons of data gathered during the field program indicates that several faults are active and that the sense of motion is dominantly right-lateral strike-slip. Rates of motion on these faults areprobably small, but they obviously represent complications to a simple plate boundary.

A report detailing microearthquake studies in Baja California, Mexico is attached (Appendix A). Located microearthquakes substantially extend the known active portion of the San Miguel fault and show that it is clearly a major tectonic feature. Although we were unable to completely delineate the fault a simple extension of the known San Miguel trend would pass near the Otay Mountain site so that further study of this fault is suggested.

A paper entitled "Comparative tectonics of the San Andreas and Alpine fault systems" was submitted to the Journal of the Geological Society of London (Appendix B). Similar relations between structural elements of the fault systems are pointed out. The type of movement, that is by infrequent large earthquakes, by more frequent intermediate size events, or by aseismic creep, is related to structural features of the faults.

The largest positive anomaly determined by comparing the gravity field of a reference ellipsoid to the earth's actual gravity field as reflected in GEM 6 and $1^0 \times 1^0$ surface gravity data lies in New Guinea. A paper submitted to Geology (Appendix C) reinterprets the geology of eastern New Guinea, describing it in terms of collision between the leading edge of the northward moving Australian continent and an island arc. Another arc further west is colliding presently with the continent. The collision appears to be causing rapid uplift in eastern New Guinea as the former arc overrides part of the continental structure.

In July, this year, the trilateration networks set up in 1972 near the Quincy and Otay Mountain laser ranging sites will be reoccupied. These networks were set up to monitor any movements that may occur on local faults and thus effect the long baseline measurements.

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APPENDIX A

A Study of Microseismicity in northern Baja California, Mexico

by

Tracy L. Johnson, Juan Madrid and Theodore Koczynski

ABSTRACT

Five microearthquake instruments were operated for two months in 1974 in a small mobile array deployed at various sites near the Agua Blanca and San Miguel faults. An 80 km long section of the San Miguel fault zone is presently active seismically, producing the vast majority of recorded earthquakes. Very low activity was recorded on the Agua Blanca fault. Events were also located near normal faults forming the eastern edge of the Sierra Juarez suggesting that these faults are active. Hypocenters on the San Miguel fault range in depth from 0 to 20 km although two thirds are in the upper 10 km. A composite focal mechanism showing a mixture of right-lateral and dip slip, east side up, is similar to a solution obtained for the 1956 San Miguel earthquake which proved consistent with observed surface deformation.

Geologic evidence of recent deformation and diffuse but persistent seismicity indicate that motion is presently occurring on faults in northern Baja California, Mexico, and off the southern California coast. The common plate tectonic model of the area (e.g. Larson et al., 1968; Lomnitz et al., 1970) depicts present movement by an en echellon series of transform faults and spreading centers in the Gulf of California which eventually join the San Andreas fault system. Active faults in the continental borderland and in Baja California which lie west of the plate boundary in the gulf thus reflect internal deformation of the Pacific plate or the existence of a subplate. We conducted a microearthquake survey in northern Baja California, Mexico, to study the relation between faults in the borderland and the plate margin located in the Gulf of California.

The numerous earthquakes which have been located in northern Baja California by the California Institute of Technology (Hileman et al., 1973) appear to be scattered across the peninsula, suggesting a broad zone of deformation (Figure 1). Until recently however, a lack of local stations limited the accuracy of locations, so the scatter may be artificial. A field program, employing five portable seismometers operated as a small mobile array, was undertaken to locate earthquakes in northern Baja California more accurately to determine if seismicity is mainly related to known major faults in the area and to further define fault locations. Another part of the project was the installation of four quadrilateral geodetic figures across known faults in northern Baja California. These figures have dimensions of about a kilometer and, when resurveyed, will measure any motion on the faults. Field work was carried out in cooperation with Professors Cinna Lomnitz at the Universidad Nacional Autonoma in Mexico City and James Brune at the

University of California in San Diego.

Regional Features. Baja California is divided by a rugged NNW trending central mountain chain with an average elevation of 6 to 7,000 feet (2 km) and a few peaks reaching 10,000 feet (3 km). Toward the west the land slopes gently to the Pacific but in the east the mountains drop precipitiously to a low coastal plain. Northern Baja California thus resembles a block tilted about a horizontal NNW trending axis with the east side relatively raised, similar to the Sierra Nevada Mountains further north (Lindgren, 1888).

Rocks in Baja California can be divided into three groups (Allen, et al., 1960) defined by relations to extensive lower Cretaceous intrusions. 1) Pre-batholithic rocks are generally steeply dipping eugeosynclinal deposits, which are slightly metamorphosed except near intrusions. A few early Cretaceous fossils have been collected from pre-batholithic rocks but older units probably exist. Along the Pacific coast pre-batholithic rocks are dominant, but plutons become more numerous as the central mountains are approached and eventually the plutons coalesce to largely form the main ranges. 2) Batholithic rocks are commonly granodiorite or quartz diorite but scattered outcrops of gabbro are present (Castil et al., 1971). Plutons in the penninsular ranges in Mexico were intruded about 95 to 115 m.y. Detritus in the upper Cretaceous Rosario Group suggests uplift of the crustal ranges about 80 m.y. (Krummenacher et al., 1975). 3) Post batholithic rocks are divided into Tertiary marine sediments which outcrop mainly along the Pacific coast, and Tertiary to Recent volcanics and derived sedimentary rocks which are found on top of the Sierra Juarez and in the gulf coastal plain. Close association of Tertiary marine sediments with the present coastline suggests that a small amount of uplift has occurred on the Pacific coast but volcanic flows on the Sierra Juarez are truncated by the eastern escarpment, indicating recent uplift in the east (Allen et al., 1960).

Several active faults are known in northern Baja California (Figure 2). The Agua Blanca fault was described by Allen _et al._ (1960), who observed geomorphic features such as offset stream beds which outlined much of the trace of the fault and indicated recent strike-slip motions in a right-lateral sense. Over most of its length the Agua Blanca strikes about east-west, only turning to parallel the northwest trend of the San Andreas near the Pacific coast at Punta Banda (PB in Figure 3). A northwesterly continuation offshore is suggested by bathymetry and faulted sediments observed in profiler records (Moore, 1969). In the east the fault is lost in Valle de la Trinidad and no continuation into the gulf is known (Allen _et al._, 1960; Gastile _et al._, 1971; Henyey and Bischoff, 1973; Moore, 1973). Allen _et al._ tentatively suggest about 20 km of right-lateral strike-slip motion has occurred on this fault since the Cretaceous.

The San Miguel fault was definitely located when a series of moderate earthquakes (largest $M_L = 6.8$) in 1956 resulted in ground breakage over about 20 km (Shor and Roberts, 1958). Displacement observed after the earthquakes was variable in amount but consistently showed right lateral motion with a dip slip component, east side up. A study of microearthquake activity along the San Miguel fault (Reyes _et al._, 1975) detected large numbers of events with short S-P intervals indicating continued activity on the fault. Linear topographic features and aligned springs suggest a northwest continuation of the fault for another 60 km before its trace is lost. Toward the southeast, the San Miguel approaches the eastern end of the Agua Blanca fault in the Sierra Juarez. While this range provides good rock exposure, access is difficult and neither fault has been traced through the mountains. Sediments in the coastal plain hinder detection of possible continuations into the Gulf of either the Agua Blance or San Miguel faults, but present information suggests that no continuations exist.

Eastern escarpments of the Sierra Juarez and San Pedro Martir mark the locations of a series of normal faults striking sub-parallel to the axis of the gulf which form a third major fault system in northern Baja California. Scarps suggest recent motion on some of these faults (Gastil et al., 1971). Although normal faults are ubiquitous in rift areas, these are interesting as they strike approximately parallel with the direction of spreading in the Gulf of California.

Field Program. Five portable smoked drum microearthquake instruments, designed at Lamont-Doherty, were used in the survey. Geophones were built by Mark Products and had a three hz free period. Noise levels, primarily due to wind or other natural sources, were such that no filtering of the signal was required and displacement amplifications of about 3 to $6 \cdot 10^5$ at 20 hz were usually possible. Response of the system is limited by geophone sensitivity at low frequencies and by pen motor response (50 hz) at high frequencies. A Sprengnether crystal oscillator in each unit provided time control. Second marks were written on the record and WWV time signals were recorded at the beginning and end of most records to determine clock drift. With a microscope the P time of local events could be read to .01 sec. Considering the various sources of error in the system it is felt that sharp arrivals could be timed to $\pm$ .05 sec. Uncertainty in S wave arrival times is greater due to emergent beginnings.

P and S wave arrival times were used in a computer program to locate hypocenters. Approximately 200 events were located. Those (81) with RMS error less than 0.2 sec and five or more arrival times are plotted in Figure 3. Velocity structure (layer 1 V_p = 5.73 km/sec thickness = 10.5 km, layer 2 V_p = 6.38 km/sec thickness = 15.5 km, half space V_p = 8.00 and V_p/V_s = 1.74) was taken from a surface wave study by Thatcher and Brune (1973). Calculated uncertainties in epicentral locations determined by the program for events within an array were about two km. Actual local velocities probably vary which introduces uncertainty in epicenter and especially depth. No known explosions were recorded during the field program so a local velocity study was impossible. By assuming that earthquakes located near faults actually occur on the mapped trace, however, the uncertainty in locations can be estimated. Most events located by a network centered on the San Miguel fault lie within a few km of the mapped fault. When located by an array about 60 km to the west near Punta Banda, events apparently occurring on the San Miguel fault are consistently to the west of the mapped fault by 5 to 10 km, suggesting that actual velocities are a bit higher than those of the crustal model for this path. Two events apparently located on the Cerro Prieto fault near the mouth of the Colorado river, about 100 km from the San Miguel array, are about 5 km east of the fault. Thick sediments of the Colorado delta may lower the average P velocity sufficiently to match the model.

A total of four braced quadrilateral figures were established across the Agua Blanca and San Miguel faults. A Laser Systems and Electronics Ranger type electro-optical distance measuring instrument was used. Manufacturer's specifications are a resolution of one mm and a nominal accuracy of $\pm$ (5 mm + 2 ppm). Corrections were applied for barometric pressure and temperature measured at the ends of the survey lines. Resurveying in a few years will detect if any significant creep occurs on the faults.

<u>Observations</u>. Microearthquake instruments were operated in a network centered on the western part of the Agua Blance fault for three weeks (see Figure 3 and Table 1). During this period many events were located near the San Miguel fault but only one small event occurred on the Agua Blanca fault. During the entire two months recording period only three events were located near the Agua Blanca

compared to about 20 events per day on stations located near the San Miguel. Such a low level of seismicity is suprising since geomorphic features clearly suggest recent motion and the number of earthquakes located over the past 40 years near the Agua Blance is smaller but of the same order as those located near the San Miguel fault (Figure 1).

Activity along the San Miguel fault dominates our seismicity map of northern Baja California (Figure 3). Epicenters form a narrow band extending about 50 km northwest of the 1956 ground breakage and suggest a 30 km extension southeastward through the Sierra Juarez. Ground breakage associated with the 1956 earthquakes thus occurred over about one quarter of the presently active zone defined by microearthquakes. Near Ojos Negros (ON in Figure 3), north of the 1956 earthquakes, microearthquakes suggest that the fault passes through Valle de San Raphael where its trace is hidden by soil instead of following faults mapped on the northeast side of the valley. Further northwest, microearthquake activity diminished. Reyes _et al._ also recorded little or no activity and very little seismicity was recorded from this area by C.I.T. in the last 40 years. Strong geologic evidence for recent faulting following a northwest trend near the international border in the Tijuana area is also lacking, but thorough mapping is not completed. Further study to determine the northwest extent of the San Miguel fault system is important since a simple extension would pass near Tijuana and San Diego (e.g. Moore, 1972). If such a fault exists it would increase the probability that damaging earthquakes could occur in this presently seismically inactive area.

Epicenters, which are more scattered than those on the northwest branch, extend 30 km southeast of the section of the San Miguel fault defined by surface breakage. Interpretations of seismicity near the southeastern part of the fault is complicated by the proximity of NNW trending normal faults which appear to be active. Unfortunately earthquake first motion data from individual events were not extensive enough to provide focal mechanisms for distinguishing between faults. Seismicity is consistent, however, with an extension of the San Miguel fault through the Sierra Juarez, no evidence indicating a continuation into the Gulf of California of either the San Miguel or Agua Blance faults was found.

A number of microearthquakes north of the San Miguel fault lie near mapped normal faults and probably result from continued tilting of the central block, although no focal mechanisms were obtained from these events so this suggestion is not proven. A series of earthquakes (largest M_L = 4.5) occurred in 1968 north of Ensenada distinctly off the Agua Blanca or San Miguel faults, but in an area where normal faults are mapped. These events suggest that active normal faults may also exist on the west coast.

A few events were located in the area between the Agua Blanca and San Miguel faults. Some of these microearthquakes probably represent mislocations due to uncertain velocity structure but others may be associated with nearby mapped faults. Difficult access to the area and the small number of events hindered more extensive study of the seismicity. Gold was mined in the 1890's in this area, raising the possibility that some events are explosions from continued mining. The single event located south of the Agua Blance fault is near several mines indicated on the geologic map of Gastil et al. Bearing in mind the limited time sample, the microearthquakes we located are consistent with the idea that most motion is occurring on a few fault systems rather than throughout the whole peninsula.

Depths of the 43 events located on or near the San Miguel fault between stations SR and SC fall between 0 and 20 km. Dividing this range into 5 km thick zones, 28 events (66%) were equally divided between the top two layers, 9 events (21%) fell between 11-15 km, and 6 (14%) of the events were in the lowest layer. The trial depth assumed in the computer program was 10 km. Formal uncertainty in depth was usually calculated as about 2 km so events located near 10 km tended to remain at the trial depth. Of the six events in the lowest layer only two events which occurred within about 12 hours and only a few km apart were located between 18 and 20 km depth. Reyes et al. located some events in the same general area and found depths of 8 to 14 km. Within the uncertainty of location essentially all our events occurred at depths of 15 km or shallower and two thirds were shallower than 10 km.

Hodgson and Stevens (recomputed in Wickens and Hodgson, 1967) determined a focal mechanism for the main San Miguel earthquake which was consistent with observed surface deformation. First motions from microearthquakes were used to produce a composite focal mechanism which proved to be consistent with that of the 1956 event. Nodal planes determined for the 1956 earthquake are plotted in Figure 4 with first motion readings from five microearthquakes located on the San Miguel trend between stations SR and SC. Minor rotation of the planes will make all readings consistent aside from two compressions in the far northwest quadrant. One of these is a refracted arrival and thus untrustworthy in view of the simple crustal model. The other is inconsistent and demonstrates the possibility of rotations of 10° to 20° of the nodal planes from event to event. Events located at other points along the fault appear grossly consistent with the composite mechanism although rotations of the nodal planes are definitely suggested. Insufficient data were recorded to produce a reliable focal mechanism at other points along the San Miguel or other faults.

Discussion. Seismicity during March and April, 1974, in northern Baja California mainly occurs on a few reasonably well defined fault zones. Although the Agua Blanca fault was essentially aseismic during this period, geologic and earlier seismic evidence suggest that it should be considered active. The San Miguel fault is clearly a presently active major feature. It is difficult to interpret the San Miguel in terms of a regional pattern since its extent, and thus its relation to other faults, is unknown.

Seismicity suggests present activity on normal faults striking subparallel to the axis of the Gulf and located on both its east and west sides. Active normal faulting on the eastern side of the gulf in Sonora is shown by surface displacements observed after a major earthquake in 1887 (epicenter at 30°48'24" North, 109°5'55" West, Aguilera, 1920. The article is a translation summarizing Aguilera's 1888 report.). Normal faulting with the west side down was observed over a distance of 56 km with predominantly NNW trend. Vertical displacements of 1 to 4 m were commonly observed and 8 m was observed in one area. It is not known if any strike-slip motion occurred.

Typical morphology of a section normal to the axis of a rift zone is a broad arch, commonly several hundred km wide, containing the rift as a central down-dropped section (e.g. Lowell and Genik, 1972). Opening direction is about perpendicular to the strike of normal faults forming the graben. Arching is usually considered to be related to thermal expansion due to perturbations in thermal gradient related to the rifting process (e.g. Sclater, et al., 1971; Dewey and Burke, 1974). Topography around the Gulf of California although complex, especially in Sonora where basin and range style deformation also occurs, is consistent with the arch like form usually found around rift zones, even though the spreading direction if subparallel with the axis of the gulf.

Perhaps thermal effects of the short ridge segments which march up the Gulf overlap sufficiently to produce a broad central thermal anomaly capable of producing uplift of the surrounding continental crust and thus normal faults striking subparallel to the opening direction and producing a complex plate boundary.

ACKNOWLEDGEMENTS

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TABLE I

Location of Seismic Stations

Name	Initials	North Latitude	West Longitude	Date of Occupation
Punta Banda	PB	31°41.1'	116°38.1'	March 6-12
San Francisquito	SF	31°41.4'	116°29.5'	March 7-28
Santa Tomas	ST	31°34.4'	116°34.4'	March 8-14
Ejido Uruapan	EU	31°36.5'	116°25.6'	March 9-14
Cerro Chocolate	CC	31°32.5'	116°24.9'	March 9-10
Southeast Santo Tomas	SS	31°31.6'	116°17.1'	March 12-26
Agua Blanca	AB	31°24.7'	116°16.6'	March 11-20
San Vicente	SV	31°21.3'	116°04.1'	March 14-26
Ojos Negros	ON	31°55.4'	116°21.2'	March 14-25
San Rafael	SR	31°51.5'	116°09.7'	March 23 - April 30
El Alamo	EA	31°37.1'	116°00.1'	March 25 - April 20
Santa Catarina	SC	31°39.8'	115°48.7'	March 29 - April 20
Valle Trinidad East	VE	31°22.5'	115°28.7'	March 30 - April 16
Valle Trinidad South	VS	31°18.6'	115°45.2'	March 30 - April 8
Guadalupe	GA	32°20.5'	116°29.4'	April 18-21

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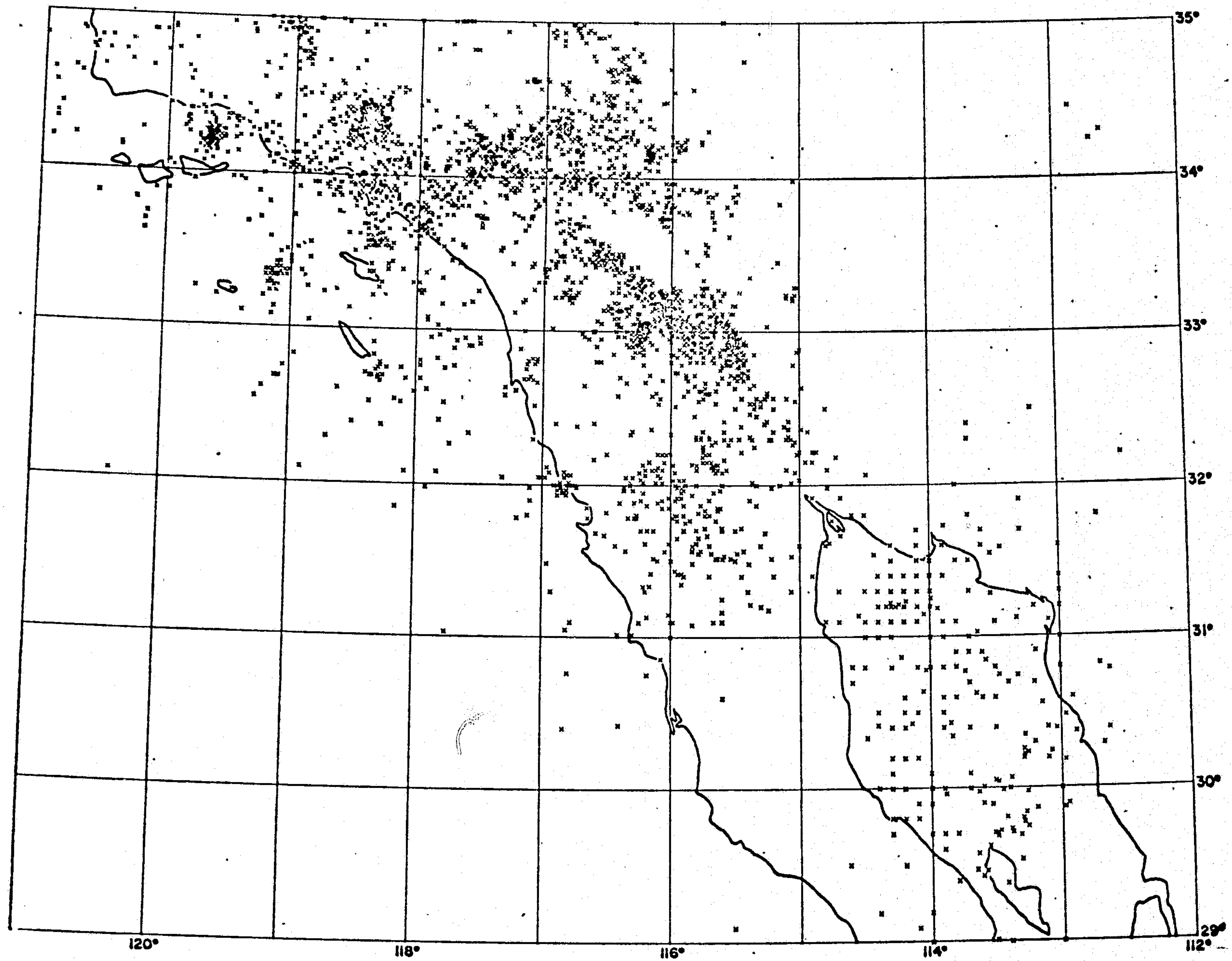
FIGURE CAPTIONS

Figure 1: All available earthquake epicenters in the map area from January 1961 to February 1975. Locations were computed by California Institute of Technology or by U.S. Geological Survey, National Earthquake Information Service.

Figure 2: Geologic map of the southern California area showing general fault trends. Data are from various articles cited in the text and the Geologic Map of California.

Figure 3: Microearthquakes located during this survey. Letter codes indicate stations occupied at different times (see Table 1). Surface deformation associated with the San Miguel earthquake is shown by a heavy line near SC. Agua Blance fault runs from PB to VE. Other faults are from Gastil et al. (1971).

Figure 4: Upper hemisphere equal area plot of first motions from five microearthquakes located on the San Miguel trend between stations SR and SC. Nodal planes were determined for the 1956 San Miguel earthquake (Wickens and Hodgson, 1967). Minor rotation of the planes satisfies all but two readings, one of which is a refracted arrival.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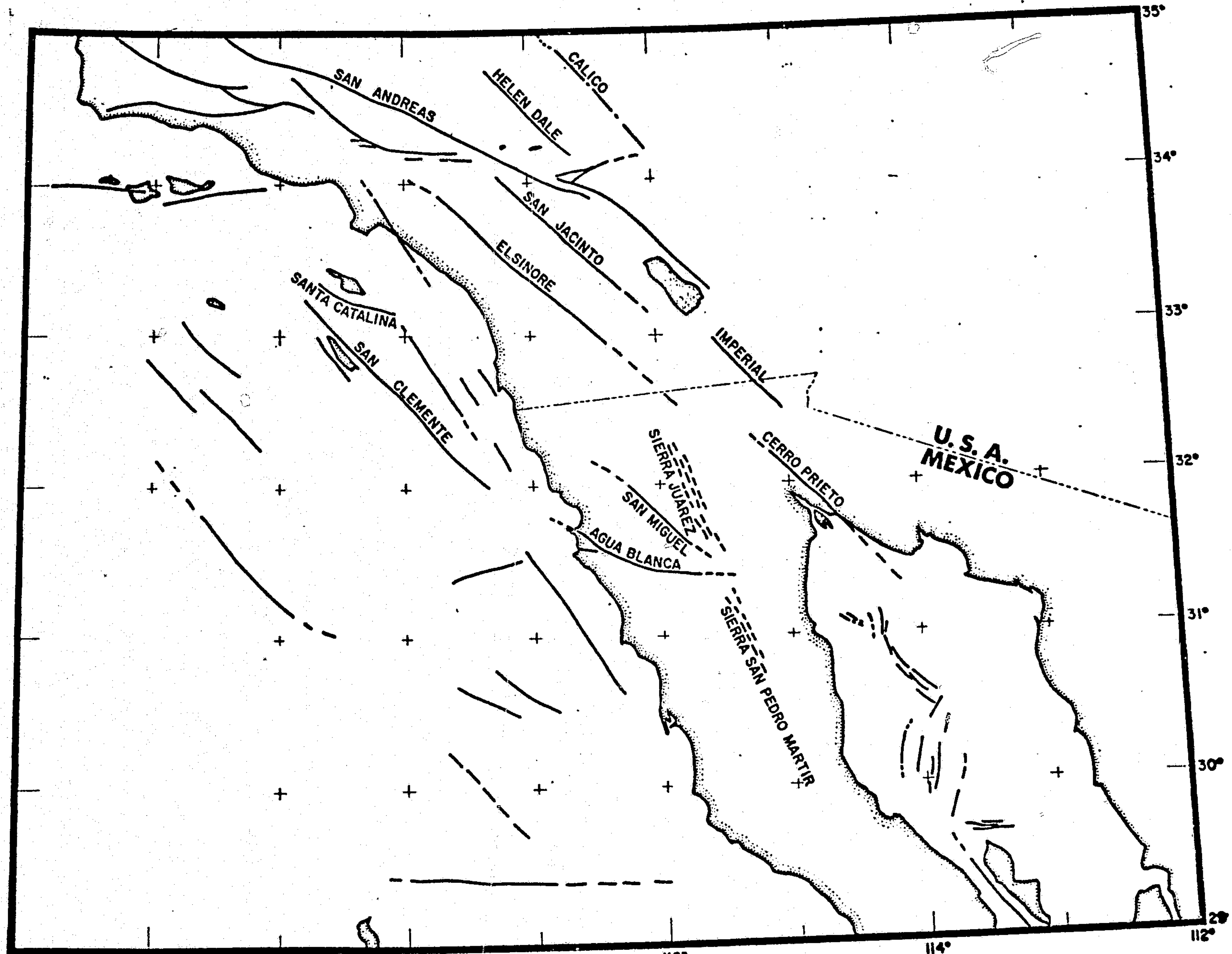

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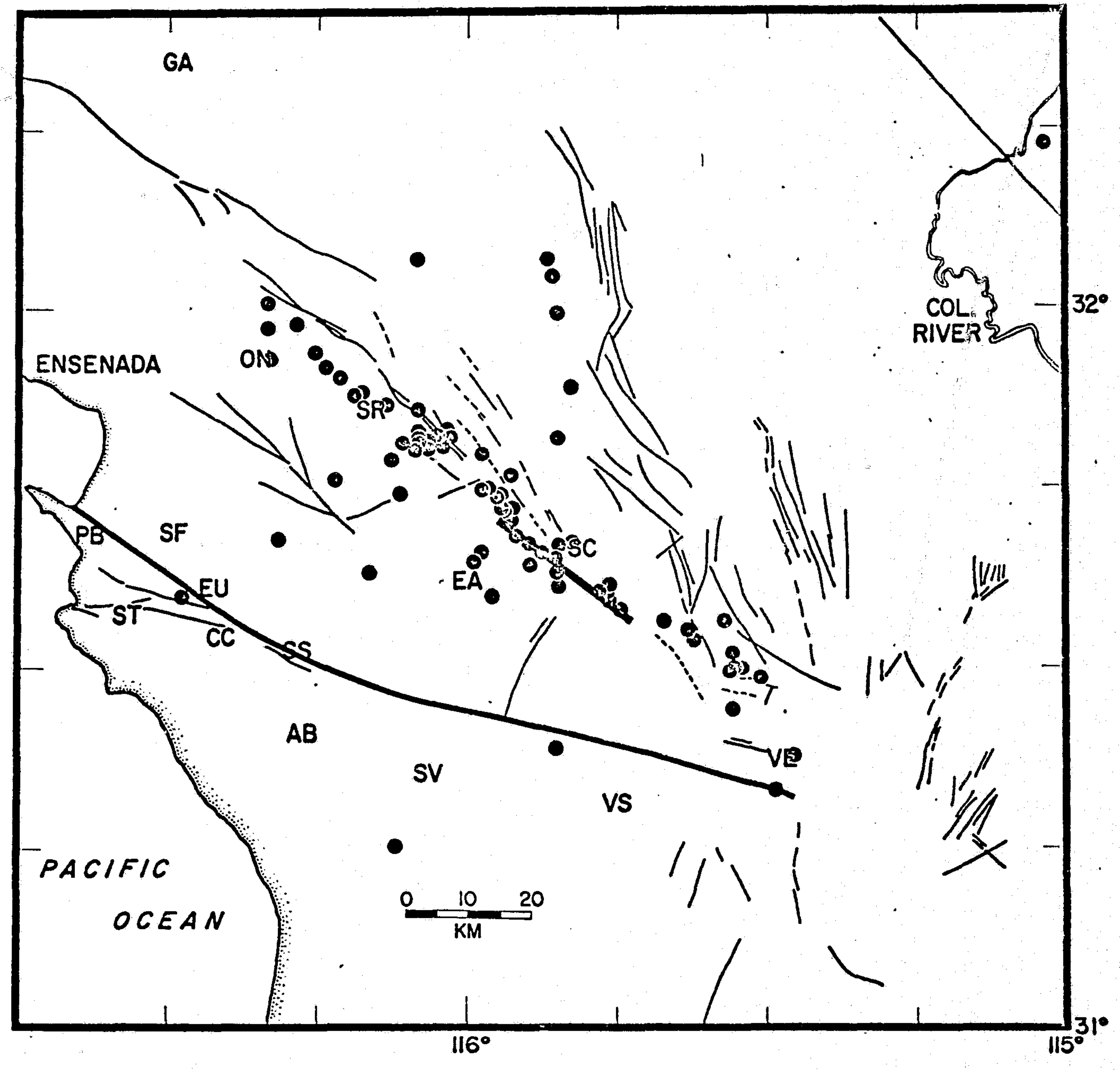
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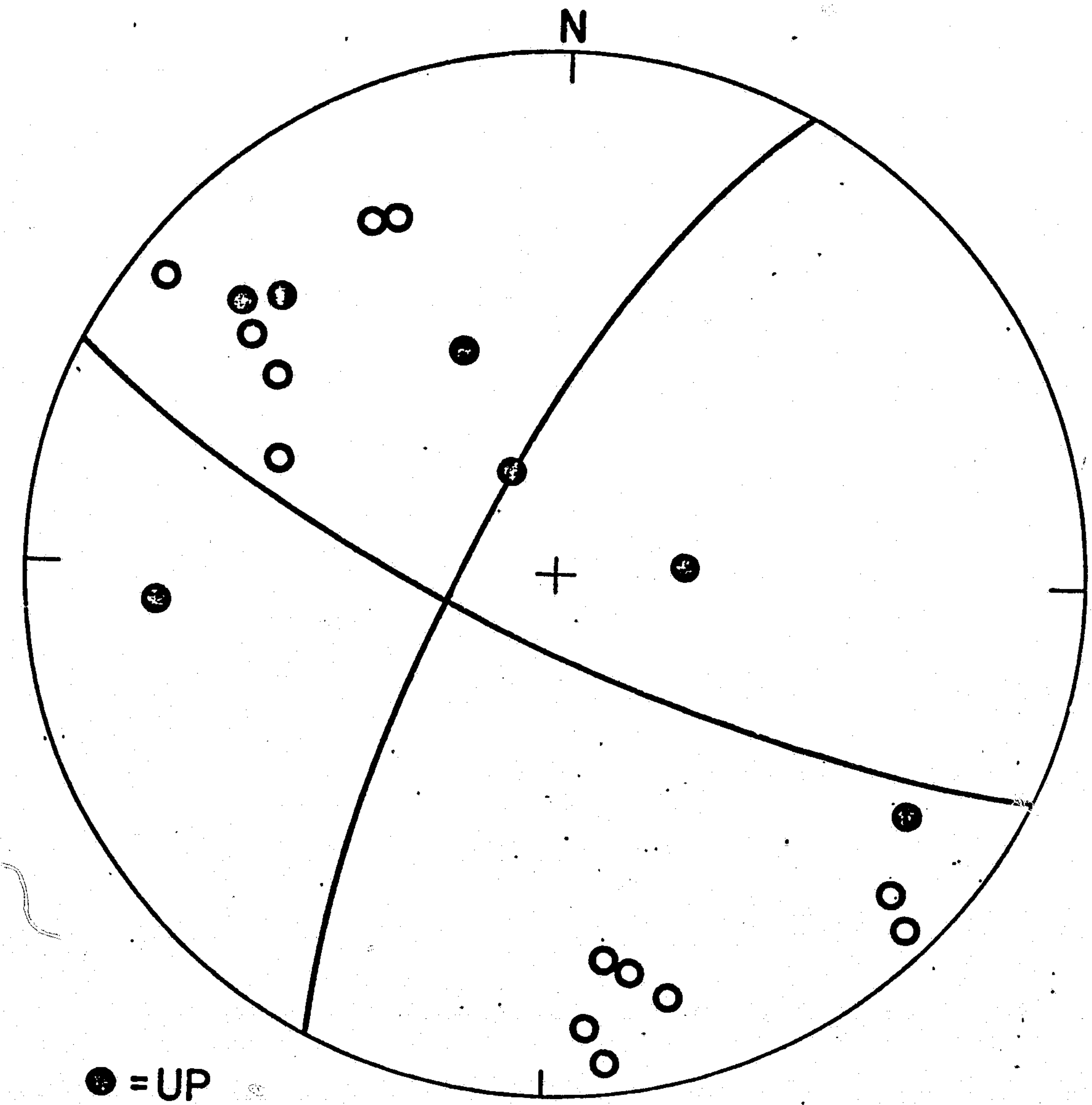
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APPENDIX B

COMPARATIVE TECTONICS: THE SAN ANDREAS AND ALPINE FAULT SYSTEMS

C.H. Scholz

Summary

Three distinctive modes of slip, or "seismic style", can be recognized for different segments of these major transform fault systems. These are: 1) slip during great ($M \geq 8$) earthquakes separated by long periods of quiescence; 2) slip during more frequent large ($6.5 \leq M \leq 7.5$) earthquakes, and 3) aseismic slip (creep). These different seismic styles occur in distinctive tectonic settings that argue that they represent fundamental variations in the mechanical properties of the faults. Style 1 occurs along sections of the fault oriented such that a component of tectonic convergence occurs, leading to a high effective normal stress and hence high frictional strength. Style 2 occurs along sections of faults that strike close to the regional slip vector, and hence have a lower normal stress. Aseismic slip only occurs along sections of faults in central California where anomalously high pore pressures reduce the effective normal stress to a very low value and the fault is entirely within the stable sliding frictional field.

When the two fault systems are compared within the reference frame of their regional slip vectors, their major tectonic elements are seen to occupy similar positions. This argues that a genetic relationship exists between the development of these complexities.

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In the ideal case, a transform fault lies along that part of a plate boundary parallel to the local vector of relative motion between the two plates. A transform fault thus should lie on a great circle about the pole of relative rotation of the plates. Oceanic transform faults in general obey these geometric rules, but transform faults that transect continental lithosphere often exhibit complexities that result in marked departures from ideality. For example, in the two cases that we shall discuss here, the San Andreas fault system of California and the Alpine fault system of New Zealand, major sections of the faults strike obliquely to the slip vector, resulting in continental convergence. Other sections of these fault systems consist of a number of subparallel active faults that together contribute to plate motion distributed over a several hundred km wide plate boundary.

Furthermore, different sections of these faults exhibit distinctively different mechanical properties that result in the several modes of strain release that we will call "seismic styles". Three seismic styles can be recognized for the San Andreas fault system: aseismic slip, slip by infrequent great earthquakes separated by long periods of quiescence, and slip by more frequent large earthquakes. The aseismic slip, sometimes referred to by the misnomer "creep", is accompanied by very frequent small to moderate earthquakes ($M < 6$) that do not significantly contribute to fault slip. For the purpose of this discussion we shall define great earthquakes as those of $M \geq 8$ that produce meters of slip over hundreds of kilometers of fault length, and large earthquakes as those events in the magnitude range $6.5 \leq M \leq 7.5$ that produce tens of cm of slip over tens of kms of fault length. There is clearly a continuum of earthquakes spanning the entire magnitude range, but these three size ranges seem to typify these three seismic styles, and hence these definitions are not entirely arbitrary and are convenient for this discussion.

One must ask the question whether these seismic styles are permanent features of those sections of the fault zone on which they have been observed during the scant historic record. That is, do those regions in which slip has occurred during great earthquakes always rupture by great earthquakes? Allen (1968) has argued that this is the case by showing that the sections of the faults that exhibit different seismic styles are also in distinctly different geologic environments. Thus he notes that those areas of the San Andreas system that have had great earthquakes are those sections where the fault undergoes a major bend which may serve as a "locking" device; that those sections that slip aseismically have abundant serpentinite bodies along the fault zone; and that those areas characterized by more frequent large earthquakes are those in which several subparallel splay faults exist.

In this paper we shall describe two fault systems, the San Andreas fault system of California and the Alpine fault system of New Zealand, in order to evaluate the tectonic environments of zones of differing seismic style and to attempt to understand the way in which these faults have developed the complex geometries that they now possess. The physical basis for the different seismic styles will then be interpreted in the light of laboratory studies of rock friction, with results that differ in some major respects from Allen's interpretation.

1. The San Andreas fault system: southern part

The two fault systems we are discussing are boundaries between major lithospheric plates, hence the motion on them must reflect the motion of the plates. The San Andreas system is a transform fault of the ridge-ridge type, shown schematically in Figure 1a. It takes up the motion between the northern end of the East Pacific rise spreading in the Gulf of California, and the Gorda and Juan de Fuca ridges to the north. The Alpine fault, on the other hand, is a trench-trench transform fault (Figure 1b) lying between the Hikurangi trench to the north, along which the Pacific plate is being subducted beneath the Indian plate, and the Puyseger trench to the south, which has oppositely directed subduction. Large right-lateral displacements have occurred on both faults: as much as 400 km on the San Andreas (Dickenson and Grantz, 1968) and 450 km on the Alpine fault (Wellman, 1955). Rates of motion from marine magnetic data are similar for the two faults; 6 cm/yr for the San Andreas fault (Larson et al., 1968) and 5 cm/yr for the Alpine fault (Hayes and Talwani, 1972), but geologic data indicate rates slower by about a factor of two (Dickenson and Grantz, 1968, Suggate, 1963). Whichever data set one favors, the rates of motion on these two fault systems are comparable, and since the historic record from the two areas are of similar length, a direct comparison of the seismic histories of the two fault systems is possible.

A. The big bend

A fault map of southern California is shown in Figure 2. Those sections of faults that are known to have slipped during earthquakes are shown as thick traces. Great earthquakes ($M \geq 8$) are shown with their dates enclosed in a box, events of $M \geq 7$ with their dates underlined and all other earthquakes of $M \geq 6$, for which the record is complete from about 1900, are shown as closed circles with their dates.

The 350 km long segment of the fault from Chalome in the north to Cajon Pass to the south consists of a relatively simple single strand. This section was apparently ruptured completely during the great earthquake of 1857, with displacements in places of as much as 10 m (Wood, 1955, Allen, 1968, 1975).

In more recent times, this section of the fault has been extremely quiet, showing almost no seismic activity even when measured at the microearthquake level (Brune and Allen, 1967), and no evidence for aseismic slip.

The southern two thirds of the Chalome-Cajon Pass sector is oblique by as much as 30° to the Pacific-America plate slip vector, which is very nearly parallel to the San Andreas fault in the vicinity of Cholame. The sense of this obliqueness requires convergence that is largely accomodated by crustal shortening in the Transverse Ranges.

The San Andreas fault in its big bend sector sharply separates the flat lying, nearly underformed Mojave block to the north from the Transverse Ranges to the south. The Transverse Ranges, as their name implies, are a series of folded and uplifted crustal blocks that strike east-west, oblique to the main structural grain of California. The convergence implied by the San Andreas fault north of the Transverse Ranges is evidently taken up by shortening in this province by folding and faulting on a complex system of nearly east-west striking right-lateral strike-slip and thrust faults. The destructive 1971 San Fernando earthquake occurred on one of these faults and produced both thrust and strike slip motion (Whitcomb et al., 1973). Not all crustal shortening takes place in the Transverse ranges, however. The 1952 Kern County earthquake (Figure 2) that occurred in the area to the north of the stable Mojave block, was a thrust event with a component of left-lateral motion (Gutenberg, 1955). The Buena Vista Hills thrust also lies in this area and is known to be slipping aseismically (Nason, 1971).

B. The southern faults

South of Cajon Pass, the San Andreas fault swings to a trend more nearly parallel to the regional slip vector and splits into several distinct traces. One of these, identified on Figure 2 as the Banning-Mission Creek trace, must be considered on geologic grounds to be the primary trace of the fault system, since it is for that fault that a 250 km offset can be demonstrated (Gastil, 1968). The northern part of the Banning-Mission Creek fault strikes nearly east-west. In this area field evidence suggests that it is acting as a thrust fault, dipping about 65° to the north (C.R. Allen, personal communication, see Figure 5). Thrusting on this fault may be the prime mechanism of uplift of the San Bernardino Mountains that lie to the north. The 1948 Desert Hot Springs earthquake (Figure 2) may have been a thrust event: its aftershocks dipped about 60° to the north (Richter, 1958). Aside from the 1948 earthquake, the Banning-Mission Creek fault has been almost totally quiet seismically during the historic period. This may suggest that the seismic style of this fault may be similar to that of the San Andreas in the Chalome-Cajon Pass sector, i.e. infrequent great earthquakes separated by long periods of quiescence. Geologically the two fault segments are similar in that they both have a major fraction of their lengths oblique, in a convergent sense, from the regional slip vector.

The fault that has been seismically most noticeable during the historic period is the San Jacinto fault, a right-lateral strike-slip fault that strikes nearly parallel to the slip vector and bifurcates from the Banning-Mission Creek fault to the south of Cajon Pass. This fault has been broken over almost its entire length in a series of large ($6.5 \leq M \leq 7.5$) earthquakes since 1890 (Thatcher et al., 1975). The behavior of this fault, which shows no evidence of aseismic slip, contrasts sharply with that of the San Andreas fault in the big bend region, and typifies the second style of slippage: seismic slip brought about by relatively frequent large earthquakes.

In spite of the high seismicity associated with the San Jacinto fault in the last century, it is clear that motion on this fault does not constitute the majority, or even a major fraction, of the total slip that must occur across the entire San Andreas fault system. The average rate of slip for the San Jacinto fault as a whole during this recent active period of 1890-1973 has been obtained by summing the moments of the large earthquakes that occurred during that period and is only 0.8 cm/yr (Thatcher et al., 1975). The rate of motion for the San Andreas fault system is estimated at 6 cm/yr from magnetic lineations at the mouth of the Gulf of California (Larsen et al., 1968) and at about 3 cm/yr from geologic (Dickenson and Grantz, 1968) and geodetic (Savage and Burford, 1973) measurements. Whichever estimate is correct, it is clear that the San Jacinto fault only contributes a small fraction of this motion. Furthermore, geologic observations indicate that the average slip rate of the San Jacinto fault has been only about 0.3 cm/yr, averaged over the past 2 M.Y. (Sharp, 1967) and 10,000 years (Clark et al., 1972). The activity since 1890 thus appears to be an unusually high spate, analogous to the sequence of major earthquakes that ruptured the Anatolian fault since 1939, but which reoccurs only at intervals of about 250 years (Ambraseys, 1970).

The Imperial fault and the southern extension of the San Jacinto fault in the Colorado Delta have also been ruptured along much of their lengths in two sequences of large earthquakes, those of 1934 and 1940, and an earlier pair of earthquakes in 1915. The seismic styles of these faults thus appears to be of the San Jacinto type.

In addition to the Banning-Mission creek and San Jacinto faults, several other faults in this region may contribute to the overall plate motion. The Agua Blanca and San Miguel faults in northern Baja California are two active strike-slip faults. The San Miguel fault was ruptured by the large earthquake of 1956, but it cannot be traced as far north as the international border. The Agua Blanca fault may extend north along a series of faults that have been identified by bathymetric lineaments and seismicity on the continental shelf off the coast of southern California (Allen et al., 1965).

The Elsinore fault is another major fault that may contribute to the plate motion. It has had no major earthquakes associated with it during the historic period, and has a very low level of microearthquake activity (Langenkamp and Combs, 1974).

The plate boundary defined by the San Andreas fault system south of its big bend is one of shear distributed over a series of subparallel faults covering a region several hundred kilometers wide. None of these faults evidently move by aseismic fault slip. Large earthquakes can usually be associated with faults of known Holocene activity (Allen, 1975) in this area, but small and microearthquakes appear to occur scattered over the entire area (Allen et al., 1965). This suggests regional high stresses for this area, and contrasts markedly with the minor seismicity of central California, which is almost entirely restricted to the major faults.

2. The San Andreas fault system: northern part

A. Aseismic slip

The third major style of faulting occurs on the San Andreas fault from a point about 40 km north of Cholame (Figure 3) to San Juan Battista. There slippage occurs by aseismic fault creep on the San Andreas fault and on two faults, the Hayward and Calaveras, that splay to the east. The rate of aseismic slip have been measured instrumentally in recent years and the rates, in cm/yr, are shown in Figure 3. Measurements of offsets in features such as curbs and fences indicate that this slip has been occurring continuously at least since the early part of this century (Brown and Wallace, 1968; Rogers and Nason, 1968). Savage and Burford (1973) have compared these slip rates to geodetic measurements and argue that they reflect the total plate motion, although in the author's view this is controversial (see Scholz and Fitch, 1969).

The aseismic slip is primarily of the type known in friction studies as episodic stable sliding in that most of the motion occurs in episodes during which a few mm of slip occurs within a few days with little slip between episodes. Except for a few isolated localities, creep-meter measurements have indicated that only one section of fault some 10 to 20 km long, indicated as A in Figure 3, slips by continuous stable sliding (Yamashita and Burford, 1973; Nason et al., 1974).

Aseismic slip in this region is accompanied by a high level of moderate to small earthquakes ($M < 6$). These earthquakes appear to cause slip on patches of the fault within the depth range 4 to 10 kms (Wesson and Ellsworth, 1973) but the summation of their moments indicates that they only contribute a small fraction of the slip rate that is produced aseismically. Minor seismicity in this region, in contrast to that of southern California, is almost entirely limited to the major mappable faults, and in fact delinate them clearly (Lee et al., 1972).

There is strong evidence to indicate that much of the length of the Hayward fault was ruptured in the two large earthquakes of 1836 and 1868 (Tocher, 1959). This fault is now slipping aseismically, and this should warn that in at least some areas the seismic style may not be a permanent characteristic of a given section of fault. The role of the Hayward and Calaveras faults are problematical. There are no clear continuations of these faults north of San Francisco Bay, and although seismicity occurs associated with several faults in that area, activity seems to gradually diminish to the north (Bolt and Miller, 1971).

B. Parkfield - Chalome sector

A short segment about 40 km in length between the northern end of the 1857 fault break at Chalome and the region of aseismic slip appears to exhibit a transitional seismic style. This segment has slipped during two moderate sized earthquakes in 1922 and 1966, and hence appears to behave in the San Jacinto style. The 1966 earthquake, however, was unusual in that it was followed by an anomalously large amount of aftercreep. It appears that slip during the earthquake was primarily at depth, and that the upper few kilometers of the fault began to slip aseismically after the earthquake at a rate that decreased exponentially with time (Scholz et al., 1969). The total amount of aftercreep was about 30 cm, similar to that inferred as the slip at depth during the earthquake. The upper part of the fault thus appears to slip primarily aseismically, while the lower part slips seismically.

C. San Juan Battista north

Near where the Hayward-Calaveras fault branches off, the San Andreas fault undergoes a bend to the west. The rate of aseismic slip decreases north of that point, and is no longer observed north of the town of San Juan Battista. The point where aseismic slip stops marks the southern end of the fault break of the 1906 earthquake. The 1906 earthquake is the second great earthquake that has occurred on the San Andreas fault. It ruptured the fault over a distance of about 500 km from the northern end of the fault at Cape Mendicino to San Juan Battista, and produced an average slip of nearly 4 m (Tocher, 1959). Since 1906 this section of the fault has been seismically almost totally quiet and has shown no evidence of aseismic slip.

The section of the San Andreas fault north of San Juan Battista thus has behaved similarly to that of the 1857 fault break in the big bend area to the south. As Allen (1968) has pointed out, both areas are where the fault zone is a single trace and where major bends exist that imply convergence along the fault and may serve to "lock" the fault.

3. The Alpine fault system

The historic earthquake record for the South Island of New Zealand is complete for large events from about 1830 (Eiby, 1968) and hence is of similar length as that of California. For much of this period, however, the rugged terrain actually affected by earthquakes had a population density much less than California, so that information on early earthquakes is generally sparser. Observations of fault breakage are much less complete in New Zealand, largely because of the thick bush that covers most of the area traversed by the Alpine fault system. For example, the Inangahua earthquake of 1968 was a shallow event of magnitude 7.1, and, although it had a 45 km long aftershock zone (Adams and Lowry, 1971), primary surface breakage could only be identified at two isolated localities where the fault crossed cleared land (Lensen and Otway, 1971). In cases of this type, we infer the extent of faulting from the aftershock distribution or from the extent of major secondary effects, principally landslides.

A. Regional Setting

In the past there has been some controversy over the tectonic role of the Alpine fault. This has arisen because although there is substantial evidence of right lateral motion of as much as 450 km on this fault (Wellman, 1955) there is also abundant evidence for vertical motion, amounting to as much as 20 km since mid-Miocene and accounting for the uplift of the Southern Alps (Suggate, 1963). This controversy has been resolved by recent microearthquake studies

(Scholz et al., 1972; Arabasz and Robinson, 1976). These studies show that the regional slip vector is parallel to the Hope fault (Figure 4) and thus oblique to the Alpine fault, requiring for it oblique right lateral slip with a thrusting component.

B. The Marlborough-West Nelson faults

North of Arthur's Pass, a series of subparallel faults splay off easterly from the Alpine fault. These faults, of which the most prominent are the Hope, Awatere and Wairau, are known collectively as the Marlborough-West Nelson faults. Large earthquakes occur frequently in this region, many of which are probably associated with these faults. The Amuri earthquake of 1888 produced right-lateral strike-slip motion of as much as several meters along the Hope fault (McKay, 1890), and an earthquake in 1848 is reported to have broken a 100 km length of the Awatere fault (Lensen, 1970). The 1929 Arthur's Pass earthquake did not occur on the Hope fault, but on an unmapped parallel fault some 10 km to the south (Speight, 1933). The extent of faulting during that earthquake is inferred as 50 km from the landslide distribution (Rynn and Scholz, 1976), a criterion that appears to be reliable in this terrain.

None of these earthquakes can be classified as great events. The frequent occurrence of earthquakes in the magnitude range $6.5 < M < 7.5$, and the lack of any evidence of aseismic slip on these faults (Lensen, 1970) strongly indicates that their style of slip is akin to the San Jacinto type: seismic slip by relatively frequent large events.

Since all of the Marlborough-West Nelson faults must be reckoned active on geologic and seismological grounds, this part of the Alpine fault zone is similar to the San Andreas system south of the big bend; i.e. the plate motion occurs as shear distributed over this entire group of strike-slip faults. A further similarity is that microearthquakes occur scattered over this region and are not restricted to the major faults (Scholz et al., 1972; Arabasz and Robinson, 1976).

The earthquakes of 1929 and 1968 that occurred west of this fault system (Figure 4) were high angle reverse faulting events on northerly striking faults (Lensen and Otway, 1971; Fyfe, 1929). Their mechanisms are consistent with the regional NW-SE compression direction and their tectonic role is analogous to that of the large southern California earthquakes of 1952 and 1971.

C. Alpine fault

South of the intersection of the Hope fault with the Alpine fault at Arthur's Pass, the Alpine fault defines the plate boundary as a single, relatively simple strand. It is along this part of the fault that evidence for thrusting is most prevalent (Figure 6).

No large or even moderate size earthquakes have occurred within the historic record that can be assigned to the Alpine fault in this section (Evison, 1971). This quiescence extends to the microearthquake level (Scholz et al., 1972). This long period of total quiescence is very similar to that observed on those sections of the San Andreas fault that have had great earthquakes. Although of course, we have no evidence that great earthquakes have ever occurred on the Alpine fault, we must assign the entire Indian-Pacific plate motion to this fault. Using the best estimated rate of 5 cm/yr, we can calculate that just within the historic period over 7 meters of plate motion has accumulated in the form of strain that must be released by slip on this fault. In the absence of any evidence for aseismic slip, we must conclude that this section of the Alpine fault is ripe for a great earthquake and that it shares the seismic style of the "big bend" area of the San Andreas fault.

4. Conclusions

A. Regional comparisons

In the foregoing we have noted a variety of similarities of the tectonic elements of the San Andreas and Alpine fault systems and have discussed similarities of the seismic style of these elements. We can only rationally compare the two fault systems, however, in the reference frame of the plate motions that take place across them, since it is those motions that determine the state of stress that acts on these faults. We do this if Figure 7 in which we show maps of the northern part of the Alpine fault system and the southern part of the San Andreas fault system, drawn to the same scale but where the maps have been rotated so that the slip vectors for the two areas coincide.

This projection shows that within the reference frame of the plate motions, the tectonic elements in these two areas occupy the same positions. The main trends of the Alpine fault and the San Andreas fault are oblique to the slip vector. This implies downslip (in the direction opposite that of the slip vector) convergence that is accomplished by uplift of the Southern Alps in New Zealand and of the Transverse Ranges in California. The mechanism of uplift is somewhat different, being produced by thrusting on the Alpine fault in New Zealand but in California the uplift is distributed over the Transverse Range faults.

Further downslip, the plate boundary is a broad belt of shear distributed over a series of strike-slip faults nearly parallel to the slip vector. In both cases the fault with the greatest geologically recorded offset, the Wairau and Banning-Mission Creek faults, occupy the same position. The other faults are younger and have much smaller total displacements.

In New Zealand the Alpine fault was evidently originally parallel to the slip vector and acted as a pure transcurrent fault. In mid- to late Miocene, a minor migration of the nearby Indian-Pacific plate of rotation caused a major change in the local slip vector, resulting in convergence on the Alpine fault (Scholz et al., 1972). The result was the Kaikura Orogeny (Suggate, 1963) which was marked by uplift of the Southern Alps and formation of the Marlborough-West Nelson faults. The formation of these faults was accompanied by a southward migration of the Hikurangi trench (Rynn and Scholz, 1976). The role of these faults can be seen to take up some of the motion on the Alpine fault and thus minimize the amount of continental convergence.

The big bend of the San Andreas fault also appears to have developed in the Miocene, the time of the beginning of Transform Range deformation. The bend is thought to have developed as a response to expansion in the Basin and Range Province of Nevada and Utah, that resulted in the westerly translation of that portion of the San Andreas fault north of the big bend (Scholz, et al., 1971). The San Jacinto and Elsinor faults presumably plays a similar role as the Marlborough-West Nelson faults in New Zealand: that of distributing the shear over a broader area and thus minimizing the continental convergence required.

B. Significance of seismic styles

We have noticed that those sections of these fault systems that are characterized by very infrequent great earthquakes, are those of which a major portion of the fault length lies oblique in a convergent sense to the regional slip vector. These are the big bend area of the San Andreas, the area of the 1906 fault break in northern California, the main section of the Alpine fault fronting the Alps, and possibly the Banning-Mission Creek trace of the San Andreas fault.

In contrast, all sections of these fault zones that are nearly parallel the slip vector with the exception of those in central California that slip aseismically, slip in more frequent large events.

The cause of these two distinctive seismic styles can be understood simply in terms of a difference in effective normal stress, $\bar{\sigma}_n$, acting on these fault segments. The effective normal stress $\bar{\sigma}_n = \sigma_n - p$ where σ_n is the applied normal stress and p is the pore pressure. We know from rock friction studies that the shear stress τ_1 required to produce stick-slip, and hence an earthquake, increases strongly with $\bar{\sigma}_n$, as $\tau_1 = \mu_o \bar{\sigma}_n$ at low normal stress and $\tau_1 = s_o + \mu \bar{\sigma}_n$ at high normal stress where s_o, μ_o, and μ are constants (Brace and Byerlee, 1966; Byerlee, 1967). We also know that the stress drop, $\Delta\tau$, and hence the displacement during stick-slip increases proportionally with τ_1 (Byerlee, 1970; Scholz _et al_., 1972).

The effective normal stress across a fault parallel to the slip vector is likely to be close to $\bar{\sigma}_n = (\sigma_1 - p)$ where σ_1 is the lithostatic load. If we consider then, that there is a tectonic driving stress σ oriented parallel to the slip vector at a transform plate margin, then the normal stress on any part of this plate boundary that strikes at an angle ϕ from the slip vector in a convergent sense will have a higher effective normal stress of $\bar{\sigma}_n = (\sin^2 \phi\sigma + \sigma_1 - p)$. We thus expect that those regions will be associated with earthquakes of greater stress drop and displacement and consequent longer recurrence intervals, as observed.

The difference between these two seismic styles thus is adequately explained by a difference in the effective normal stress operating across the faults. A corrolary is that transform faults striking parallel to the slip vector will be associated with large, but not great, earthquakes. This generalization appears to hold true for oceanic fracture zones, but it is not possible to apply it to other major continental transcurrent faults such as the North Anatolian fault of Turkey or the Phillipine fault, since the slip vectors for those faults cannot be determined independently from marine magnetic data.

Allen (1968) has suggested that aseismic slip prevails in central California because of the common occurrence of serpentinitic rocks along the fault zone in that region. This concept is supported by laboratory studies that show that serpentine rich rock tends to slid stably without stick-slip (Byerlee and Brace, 1968). It appears, however, that this is not sufficient to explain aseismic fault creep, since it would imply that oceanic fracture zones should slip aseismically. Kanamori and Stewart (1976) have shown that slip on the Gibbs fracture zone of the North Atlantic occurs by frequent large earthquakes that can totally account for the known slip rate of the North Atlantic of 2 cm/yr. Brune (1968) earlier reached a similar conclusions for the Romanche fracture zone. Thus oceanic fracture zones, that must contain abundant serpentinitic rock, exhibit the San Jacinto type seismic style.

Frictional sliding of rock in the laboratory occurs by continuous stable sliding at low effective normal stresses and high temperature, and gives way to stick-slip at higher normal stresses and lower temperatures (Byerlee and Brace, 1968). Episodic stable sliding, similar to that observed in most places in central California, has also been observed in the laboratory and is known to be transitional between continuous stable sliding and stick-slip (Scholz et al., 1972).

Since there is no evidence for anomalously high temperatures along the San Andreas fault in central California (Brune and Henyey, 1969), we must look for very low values of effective normal stress to explain aseismic slip. The low values of $\bar{\sigma}_n$ are evidently caused by anomalously high pore pressures in this region. Berry (1973) has shown that near lithostatic pore pressures exist in this region, and Irwin and Barnes (1975) have shown a strong geographical correlation between those sections of the faults that slip aseismically and the geologic requirements for the presence of high pore pressures.

This argument, then, though not directly corroborated by deep measurements in the fault zone, suggests that aseismic slip occurs in central California because the effective normal stress is below that necessary to cause stick-slip due to the presence of anomalously high pore pressures. This implies in turn that the shear stress required to produce slip on these sections of the fault are considerably smaller than elsewhere. This appears to explain why micro-earthquakes appear to be almost totally concentrated on the faults in central California, because the fault is much weaker there and the stress cannot build up in adjoining areas. This contrasts with southern California and New Zealand, where the occurrence of microearthquakes scattered over a broad region suggests regionally high stress.

The explanation for aseismic slip in central California is a local one, requiring unusual geologic conditions for the generation of anomalously high pore pressures. This is not suprising since the aseismic mode of fault slip seems to be quite rare: no other occurrence of it has been observed except on a few short segments of the North Anatolian fault (Ambrayseys, 1970; some of these observations may be afterslip following large earthquakes). Indeed the success, on a global scale, of gap theory (e.g. Kelleher et al., 1973) and of predicting rates of plate motion using seismic moments (Davies and Brune, 1971) precludes aseismic slip being a common phenomena.

It appears, then, that of the transform fault systems studied here, the normal mode of slip is of relatively frequent large earthquakes. Great earthquakes occur only where the effective normal stress on the fault is unusually high because the fault lies oblique to the regional slip vector. Fault creep only occurs where the effective normal stress is very low, in the case considered because of unusually high pore pressures.

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FIGURE CAPTIONS

Figure 1 Schematic diagram of two types of transform faults, a) the ridge-ridge type, b) the trench-trench type.

Figure 2 Fault map of southern California showing the major historic seismicity and documented fault slippage. Data mainly from Allen et al., 1965 and Thatcher et al., 1975.

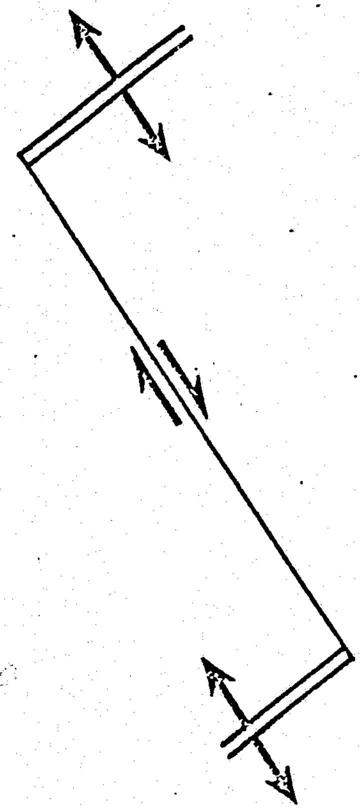
Figure 3 Fault map of central California showing major seismicity and aseismic slip. Rates of aseismic slip are from Savage and Burford (1973), seismicity from Tocher (1959) and Bolt and Miller (1971).

Figure 4 Map of the Alpine fault system, South Island, New Zealand, and historic seismicity. Data primarily from Eiby (1968) and Lensen (1970).

Figure 5 Photograph looking southwest along the San Andreas fault. The San Bernardino Mountains are to the left, the town of San Bernardino to the right. The fault shows a component of thrusting in this area. Photo courtesy of C.R. Allen.

Figure 6 The Alpine fault looking northeast along the escarpment of the Southern Alps. The town of Fox Glacier is at the bottom of the photograph.

Figure 7 Fault maps of South Island, New Zealand, and southern California. The two have been rotated so that the slip vectors are parallel for the two regions. Historic large earthquakes are indicated, with dates.



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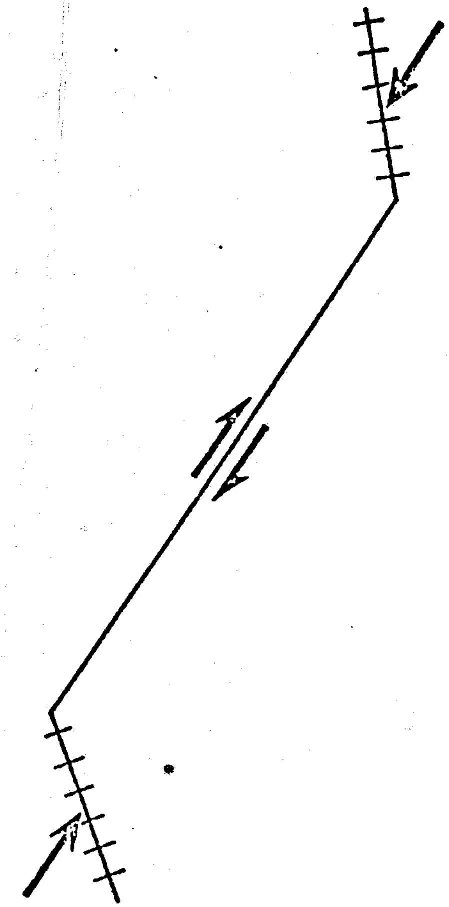

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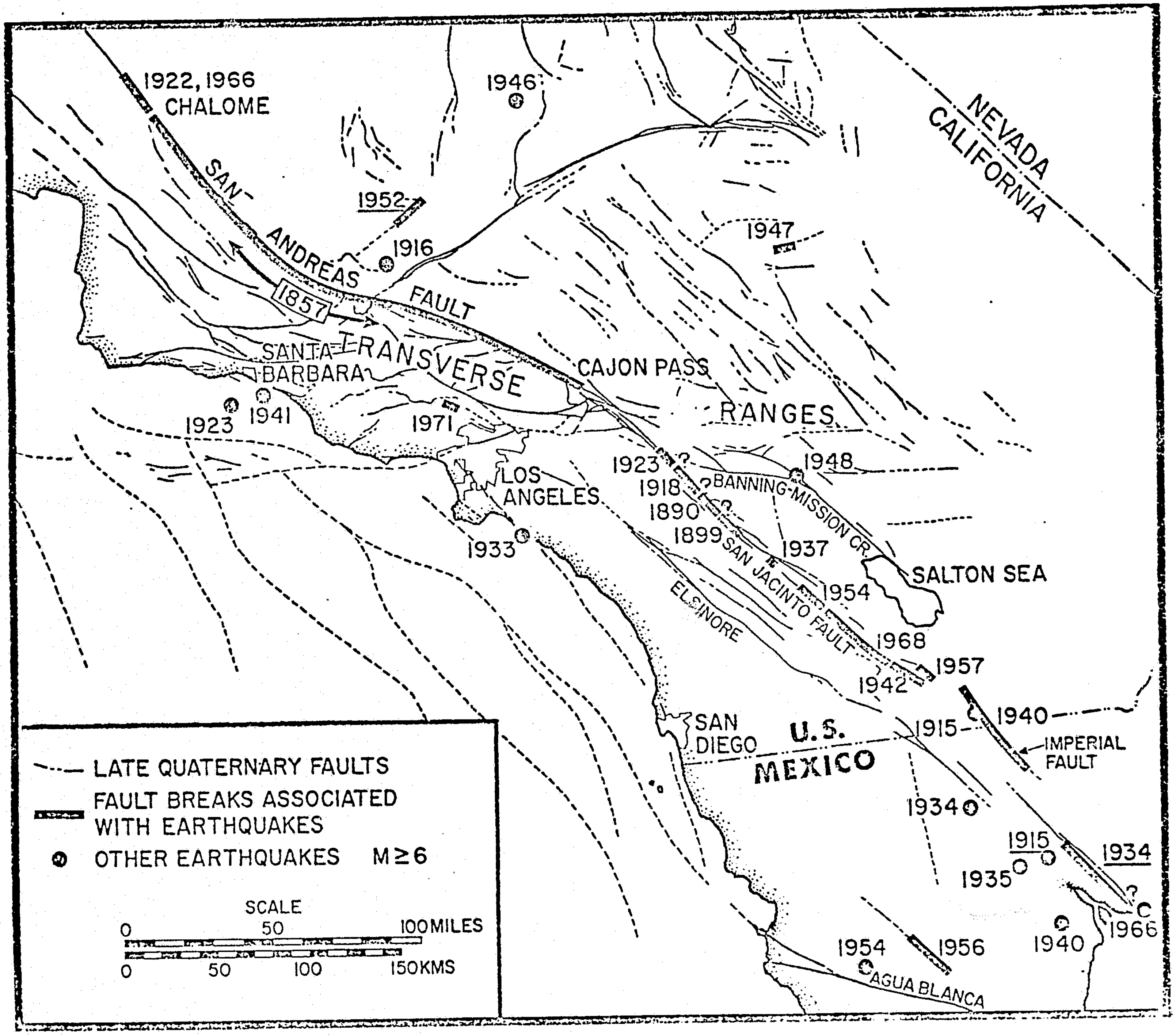

1922, 1966
CHALOME
1946
NEVADA
CALIFORNIA
SAN ANDREAS FAULT
1952
1916
1947
1857
SANTA BARBARA
TRANSVERSE
CAJON PASS
RANGES
1923
1941
1971
LOS ANGELES
1923
1918
1890
1899
1948
BANNING-MISSION CR.
1933
1937
SAN JACINTO FAULT
SALTON SEA
ELSINORE
1954
1968
1957
1942
1940
SAN DIEGO
U.S.
MEXICO
1915
IMPERIAL FAULT
1934
1915
1934
1935
1966
1954
1956
1940
AGUA BLANCA
LATE QUATERNARY FAULTS
FAULT BREAKS ASSOCIATED WITH EARTHQUAKES
OTHER EARTHQUAKES M≥6
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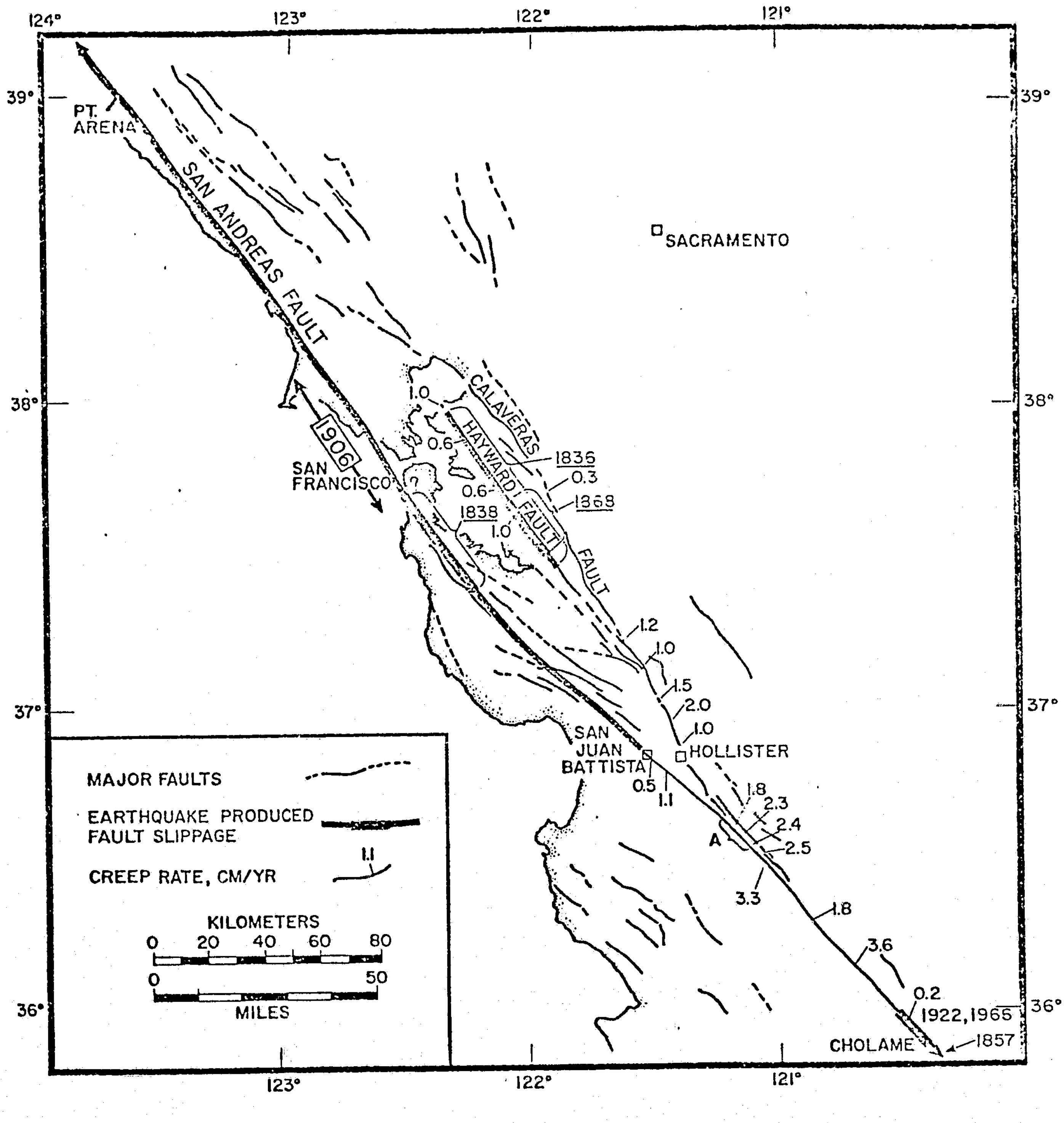
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1906
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CALAVERAS
HAYWARD FAULT
FAULT
1.0
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0.6
1836
0.3
1868
1838
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SAN JUAN BATTISTA
HOLLISTER
0.5
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A
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1922, 1966
CHOLAME
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MAJOR FAULTS
EARTHQUAKE PRODUCED FAULT SLIPPAGE
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1.1
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0 20 40 60 80
0 50
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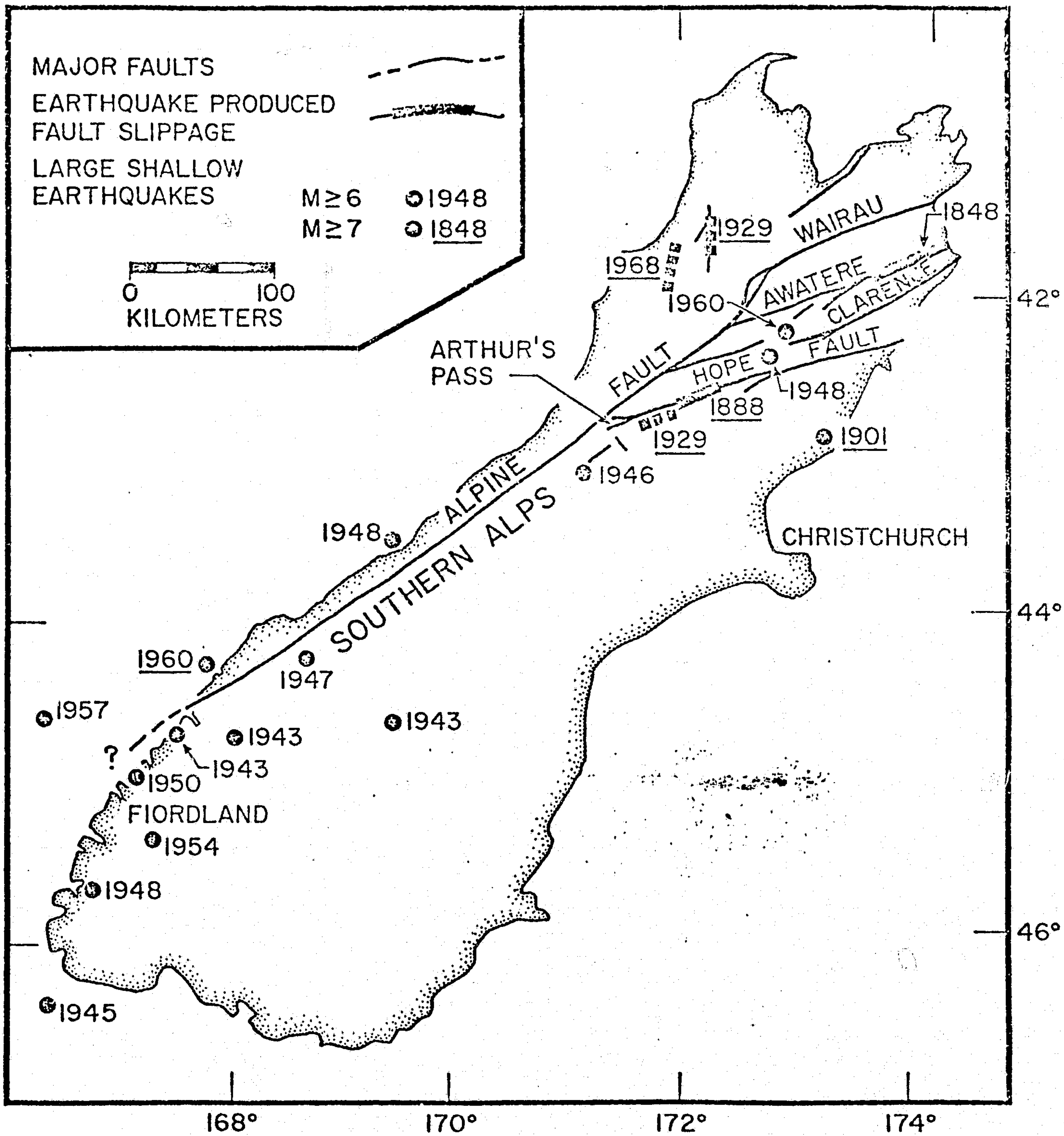
FIG 3


MAJOR FAULTS
EARTHQUAKE PRODUCED FAULT SLIPPAGE
LARGE SHALLOW EARTHQUAKES
M≥6 1948
M≥7 1848
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KILOMETERS
WAIRAU
AWATERE
CLARENCE
HOPE
FAULT
FAULT
ALPINE
SOUTHERN ALPS
ARTHUR'S PASS
CHRISTCHURCH
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1929
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1960
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174°

FIG 4

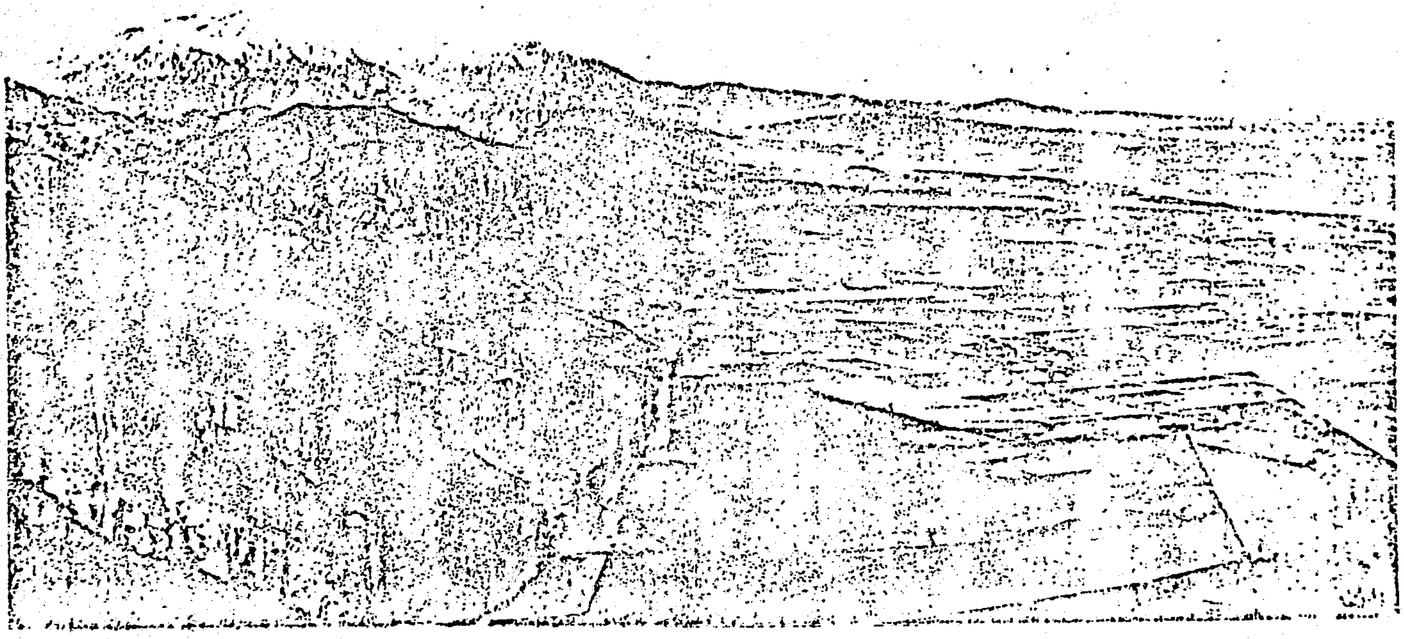

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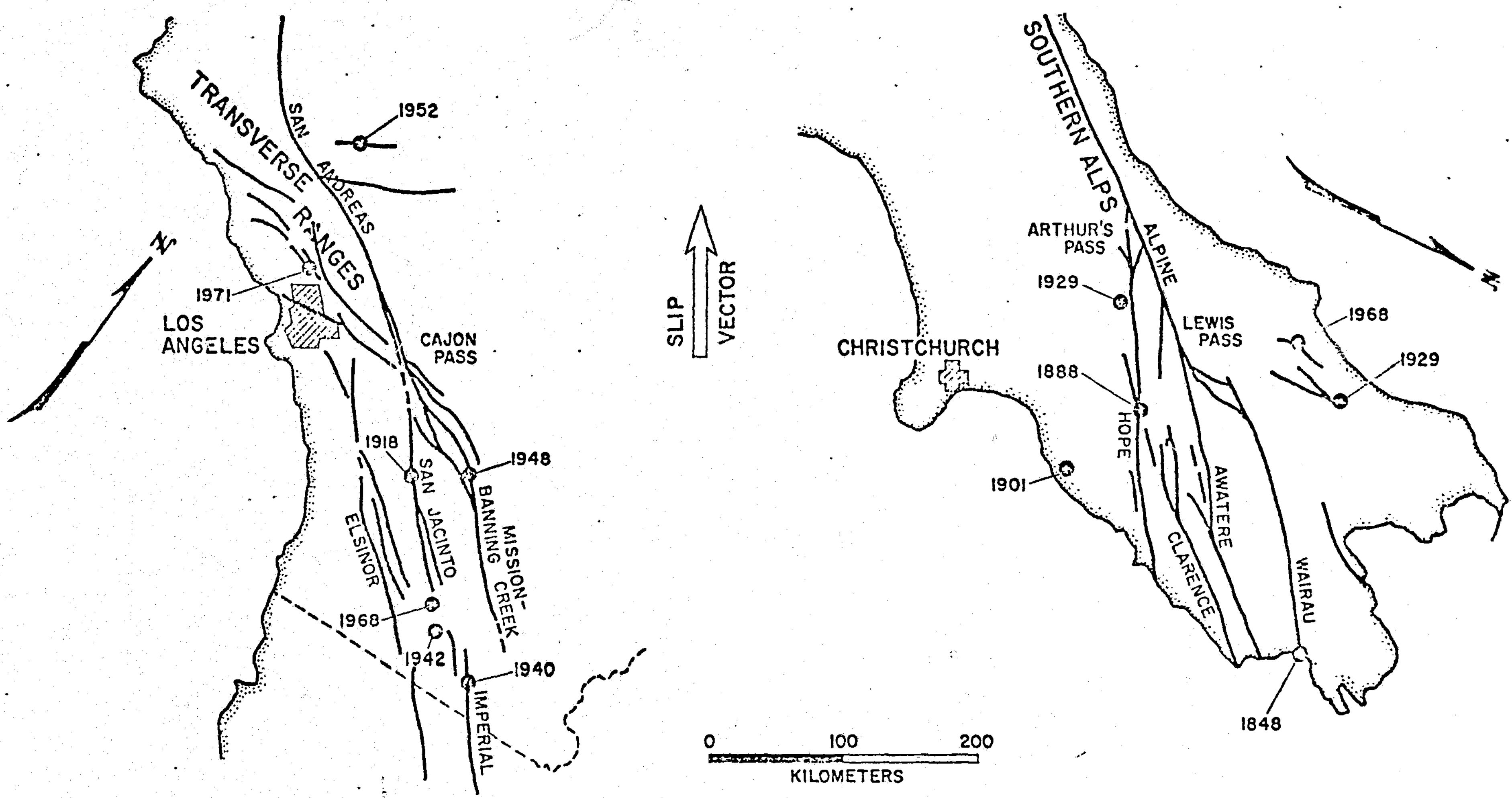
Fig. 3



CALIFORNIA

NEW ZEALAND

APPENDIX C

A NON-OBDUCTION MODEL FOR THE EMPLACEMENT OF THE PAPUAN OPHIOLITE BELT

TRACY L. JOHNSON *Lamont-Doherty Geological Observatory of Columbia University, Palisades, New York 10964*

ABSTRACT

Eastern Papua, New Guinea is interpreted as a now inactive island arc which collided with the Australia continent. The proposed arc was created in the upper Paleocene (?) on oceanic crust of Cretaceous or Jurassic age lying northeast of Australia. Rifting, which eventually created the Coral Sea, started near the Australian continental margin also in the Paleocene (?) and probably was related to the iniation of subduction. The Owen Stanley metamorphic rocks are interpreted as sediments deposited on oceanic crust and accreted onto the island arc during subduction. The Papuan ophiolite complex is interpreted as oceanic crust which was emplaced as part of the upper plate during an early stage of subduction.

INTRODUCTION

The ophiolite complex on eastern Papua, New Guinea is commonly used as an example of obduction -- the overthrusting of parts of oceanic crust onto continental edges (Coleman, 1971). In the obduction model, the Papuan ultramafic belt was thrust over the Owen Stanley metamorphic belt (Fig. 1) which is assumed to be continental crust (for example, Coleman, 1971; Davies and Smith, 1971). Since eastern Papua is a "type locality" for obduction it is important to be sure of the geologic relations. I suggest a reinterpretation of the geology in which the Owen Stanley metamorphic rocks are not "continental" but represent sediments initially deposited on oceanic crust and accreted onto an island arc as the oceanic crust they were deposited on was subducted. If this model proves reasonable then the Papuan ophiolites can be interpreted as part of the upper plate in a subduction zone under which the Owen Stanley belt was subducted. Milsom (1973) considered the possibility that the Papuan ophiolites were part of an island arc but he implicitly assumed that the Owen Stanley metamorphic rocks were continental and so was unable to produce a model which he considered to be satisfactory.

MODEL

Geologic development I propose is as follows: Cretaceous or older oceanic crust initially lay off the Australian continent (see Fig. 1 for limits of Paleozoic basement). In the Paleocene (?) a spreading ridge formed close to the continental edge and subsequently migrated away from Australia until at least middle Eocene, creating oceanic crust, part of which presently forms the Coral Sea. This oceanic crust was consumed by eastward subduction beneath an island arc lying north and east of Australia. Sediments were deformed, metamorphosed, and accreted onto the island arc during subduction. Eventually the ridge itself was subducted as the arc approached Australia. The arc neared Australia in the Miocene; its western part was slowly uplifted as it collided with the continent from about Middle Miocene through Pliocene time. Subduction of the Coral Sea floor slowed and finally ceased and motion of the surrounding plates was reorganized. In this model orogenic events in eastern Papua were not necessarily directly related to those in central New Guinea as the shape and nature of the plate boundary of which the proposed island arc was a part is undetermined.

GEOLOGIC EVIDENCE

Present seismic data are consistent with the model. Crustal thickness of eastern Papua is about 20 to 25 km (Finlayson and others, 1976). A similar crustal thickness was deduced by Grow (1973) in the central Aleutians, which are an island arc formed in the Tertiary on oceanic crust.

If eastern Papua is formed of continental material rifted from Australia there should be similarities in the pre-rift geology of formerly continuous areas. The Coral Sea basin originated in the lower Tertiary (Andrews and others, 1975). The pre-Tertiary histories of southwestern New Guinea, which undoubtedly is part of Australia, and eastern Papua do not suggest close similarity. Specifically, basement in southwestern New Guinea is granite of probable Permian age (Thompson and Fisher, 1965). Similar granitic rocks have not been found in eastern Papua (Bain, 1973). Also in southwestern New Guinea Cretaceous and Jurassic marine sediments show continental affinities while Thompson and Fisher (1965) conclude that at least back to Cretaceous time the Owen Stanley metamorphic rocks were beyond the limit of continentally-derived sediments and were separated from Australia by deep sea.

A key part of the model is the interpretation that the Owen Stanley metamorphic rocks were deposited on oceanic crust and were deformed and metamorphosed while being subducted and accreted to the island arc. Not surprisingly, information about the nature of the basement on which the Owen Stanley sediments were deposited appears to be lacking. Owen Stanley rocks themselves are "composed essentially of regionally metamorphosed graywacke sediments and limestone, and locally metamorphosed igneous rocks" (Thompson and Fisher, 1965, p. 125). "Where the original sediment can be recognized it is most commonly tuffaceous lithic labile sandstone, siltstone, or mudstone, with quartz, plagioclase, acid volcanic clasts and some carbonaceous material" (Davies and Smith, 1971, p. 3300). "Where dignostic textures or fossils are preserved, the limestone appears to be of deep rather than shallow-water origin" (*ibid*, p. 3302). "Most of the intrusives are derivatives of granodiorite magma exposed at various levels of emplacement" (Thompson and Fisher, 1965, p. 126). Slightly metamorphosed upper Cretaceous submarine basalts of oceanic tholeiite type are also present (Smith, 1975). Owen Stanley rocks are most commonly metamorphosed to greenschist facies although blue schist facies rocks occur in a belt a few kilometers wide along the Owen Stanley fault, which forms the contact between ophiolites and metamorphic rocks (Davies and Smith, 1971). "The metamorphic foliation is tightly folded in places but elsewhere dips consistently as though uniformly tilted or folded onto broad anticlinal structures. For instance, along most of the northeastern front of the (Owen Stanley) range the foliation dips consistently to the northeast" (*ibid*, p. 3301). As described, the composition of the Owen Stanley rocks seems compatible with an origin as arc derived graywacke and pelagic sediment deposited on oceanic crust. The structural relationship in which ophiolites are in thrust contact with underlying blue schist facies rocks which in turn grade into structurally lower green schist facies rocks is similar to the relationship observed in northern California between the Franciscan and the ophiolite portion of the overlying Great Valley sequence (for example, Platt, 1975). Internal structure of the Owen Stanley metamorphic rocks is not well known but the model predicts similarities to structures found in the Franciscan.

The proposed model allows a relatively straightforward explanation for the nature and timing of major features of the geologic history of Papua and the Coral Sea. JOIDES hole 287 in the middle of the Coral Sea dates the oceanic crust there as about 51 my or lower Eocene (Andrews and and others, 1975), so spreading was initiated sometime earlier (late

Paleocene?). Middle to upper Eocene submarine basalts present on the southeastern tip of the Papuan Peninsula suggest that spreading may have continued to that time at least (Davies and Smith, 1971). Geophysical knowledge of the Coral Sea basin is not complete enough to allow identification of magnetic anomalies there but magnetic lineations present appear to run approximately east-west and are not obviously symmetric (Falvey, 1972). Although not a strong point considering its age, topography associated with an extinct east-west spreading ridge is not apparent in the Coral Sea (Ewing and others, 1970). A simple explanation for the asymmetric magnetic anomaly pattern and lack of a central ridge is that the ridge, the northern half of the sea floor generated by it, and an undetermined amount of the southern half of the sea floor, a fragment of which now forms the Coral Sea, were consumed by subduction. JOIDES hole 209 located at the eastern edge of the Queensland plateau (Burns and others, 1973) as well as southwestern Papuan geology (for example Thompson and Fisher, 1965) clearly show that the Australian margin of the Coral Sea has been quiescent since the middle Eocene in sharp contrast to the complex tectonism occurring throughout the Tertiary in eastern Papua. If future studies confirm that former portions of Coral Sea oceanic crust are now missing, then subduction by an arc located northeast of Australia is a very simple solution.

Ophiolites, probably representing Cretaceous or possibly Jurassic oceanic crust, are generally considered to have been thrust from the east over the Owen Stanley metamorphic belt in the early Tertiary (for example Davies and Smith, 1971) although Rod (1974) has questioned the existence of substantial low angle thrusting on the Owen Stanley fault (Fig. 2). The Papuan ophiolites are interpreted as part of the initial sea floor of the overthrust plate on which a new island arc was formed. Davies and Smith present evidence indicating emplacement of the ophiolites over the Owen Stanley belt by lower Eocene time. Major thrusting in eastern Papua is thus roughly contemporaneous with the initiation of rifting forming the Coral Sea. It seems reasonable that factors responsible for the initiation of rifting in the Coral Sea area could also cause subduction to begin elsewhere, indeed since forces associated with down-going slabs seem to dominate plate motion (Forsyth and Uyeda, 1975), rifting could have been caused by the initiation of subduction. Continued subduction and accretion of successively younger slices of supercrustal material concurrent with uplift and erosion will produce a regional zonation in age and metamorphic grade (for example Platt, 1975). The age of rocks in the Owen Stanley belt is uncertain but metamorphic grade generally decreases from east to west (Davies and Smith, 1971) as predicted by the subduction model. There is a definite zonation in the spacial distribution of sediment types being deposited across an island arc, ranging from deep water turbidites associated with the trench through relatively shallow marine sediments on the frontal arc to sediments and intrusives associated with the volcanic chain. Spacial relations of this type have been obscured in eastern Papua by extensive erosion in the Owen Stanley ranges and by burial under Quaternary volcanics and sediments. If a very generalized section extending northeast from Port Moresby is constructed from outcrops of Eocene age rocks, however, the spacial relations described above are observed. Near Port Moresby, Eocene rocks are cherts and dark shales indicative of deep-water deposition (Thompson and Fisher, 1965). Bioclastic limestone containing clasts of schist is mapped in approximately the center of the peninsula (outcrop at 8°20'S., 147°E.) (Davies and Smith, 1971). Finally, on the northeast (8°S., 147°30'E.) there are Eocene age quartz diorite intrusives and probably associated andesitic volcanics and limestones (Davies and Smith, 1971).

A major sedimentary unit (Aure Trough, unit B in Figure 1) lies west of the Owen Stanley metamorphic belt. About 11 km of lower Miocene to Pliocene sediments were shed from the east into a trough that is today quite narrow (for example Davies and Smith, 1971). Sediments in the axis of the trough were probably deposited by turbidity currents. Clasts of reef limestone are present and the graywacke is chemically equivalent to andesite (Davies and Smith, 1971). The sedimentary rocks are tightly folded and faulted along axes which generally parallel exposures of Owen Stanley metamorphic rocks (Thompson and Fisher, 1965). In the model proposed here, as the island arc approached Australia the intervening sea became narrower and extensive uplift of the island arc began, leading to rapid deposition of arc derived sediments into a narrow trough. Sediments were deformed during the final stages of the collision (Graham and others, 1975 proposed a similar model for other areas).

Clearly any model, enumerated in a single paragraph, which purports to explain 50 my of the history of a thousand km long peninsula is bound to be oversimplified in some respects. For instance, no mention is made of opening in the Woodlark Basin (Milsom, 1970) as it is felt that this has no direct bearing on the obduction problem. However, the proposed model seems to explain the distribution of rock types, their ages, and the age of deformation reasonably well using a straightforward mechanism.

ACKNOWLEDGEMENTS

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Lamont-Doherty Geological Observatory Contribution Number 0000.

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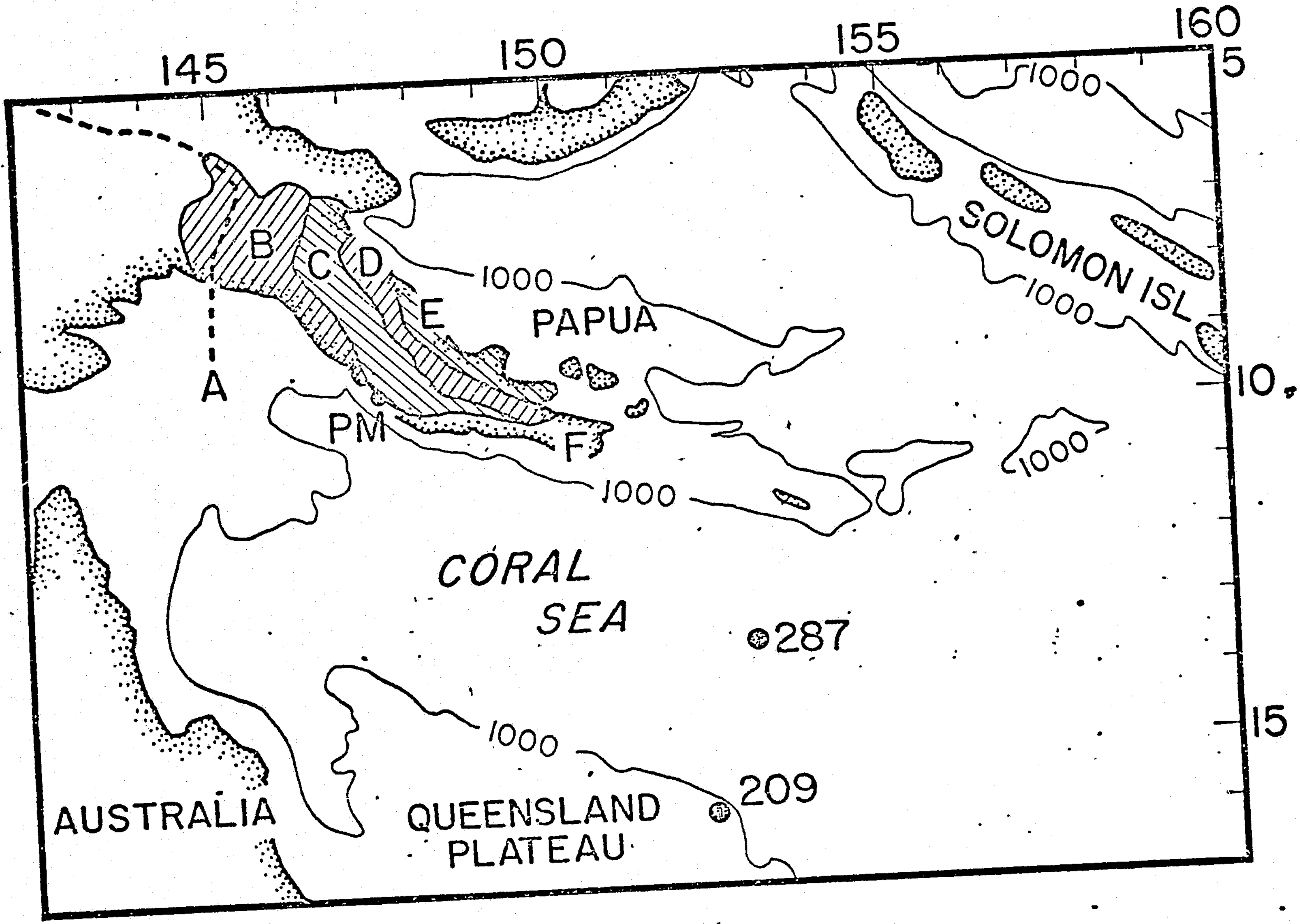
FIGURE CAPTIONS

Figure 1: Sketch map of the Papua-Coral Sea area. Bathymetric contour is 1000 fathoms (Mammerickx and others, 1971). Location of JOIDES holes 209 and 287 is plotted. Single letters label geologic features (Bain, 1973). A is the eastern limit of Paleozoic rocks in New Guinea, B the location of the Aure Trough, C is the Owen Stanley metamorphic unit, D the Papuan ophiolite belt, E the Cape Vogel basin, and F is the area containing Eocene tholeiitic submarine basalt. Port Moresby is near letters PM.

Figure 2: Generalized pre-collision section indicating the proposed geologic relations in eastern Papua. Initials represent: T = trench, OSM = Owen Stanley metamorphic belt, OPH = Papuan ophiolites, V = volcanoes. Numbers 1 through 4 schematically represent oldest through youngest slices of accreted supercrustal rock (after Platt, 1975). Lines separating the slices represent old thrust surfaces which could be occasionally reactivated.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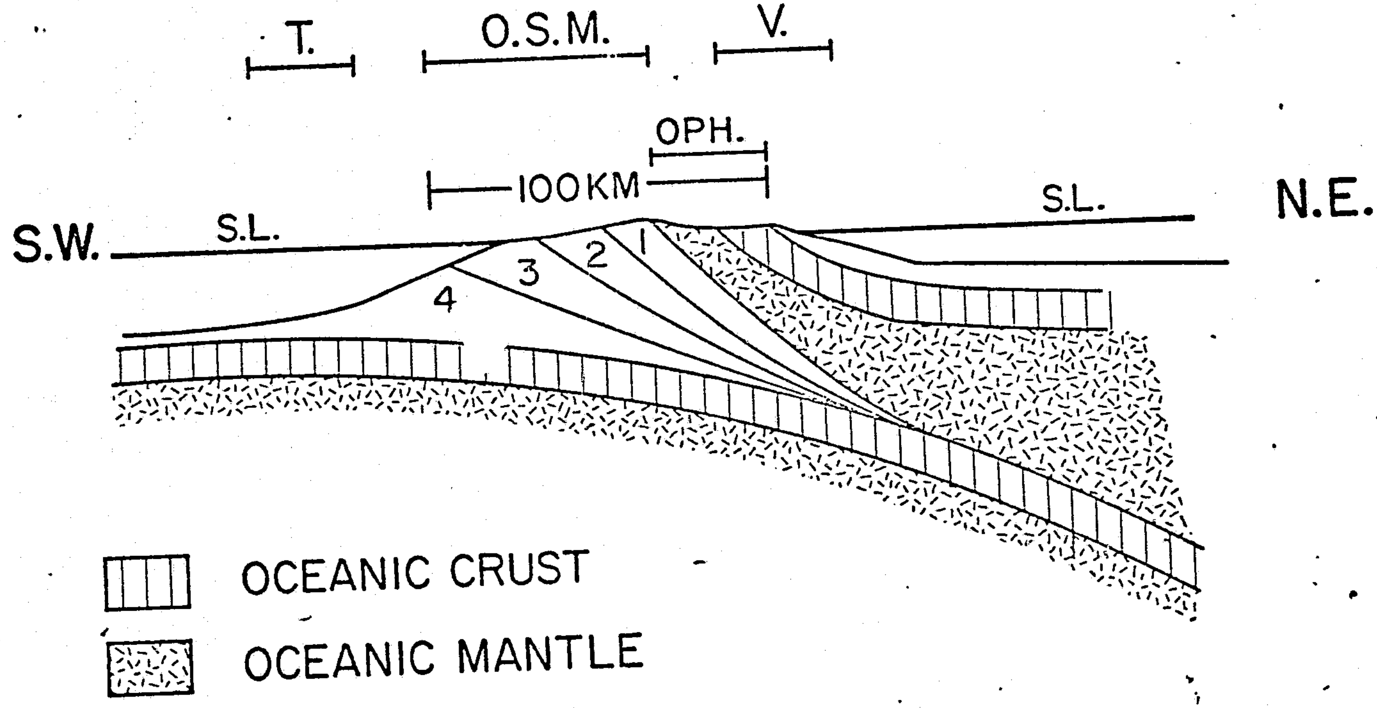

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OCEANIC CRUST
OCEANIC MANTLE

FIG. 2