PAGEOPH, Vol. 124, Nos. 1/2 (1986) 0033-4533/86/020141-17$1.50+0.20/0

Seismic Measurements of the Internal Properties of Fault Zones

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Abstract—The internal properties within and adjacent to fault zones are reviewed, principally on the basis of laboratory, borehole, and seismic refraction and reflection data. The deformation of rocks by faulting ranges from intragrain microcracking to severe alteration. Saturated microcracked and mildly fractured rocks do not exhibit a significant reduction in velocity, but, from borehole measurements, densely fractured rocks do show significantly reduced velocities, the amount of reduction generally proportional to the fracture density. Highly fractured rock and thick fault gouge along the creeping portion of the San Andreas fault are evidenced by a pronounced seismic low-velocity zone (LVZ), which is either very thin or absent along locked portions of the fault. Thus there is a correlation between fault slip behavior and seismic velocity structure within the fault zone; high pore pressure within the pronounced LVZ may be conducive to fault creep. Deep seismic reflection data indicate that crustal faults sometimes extend through the entire crust. Models of these data and geologic evidence are consistent with a composition of deep faults consisting of highly foliated, seismically anisotropic mylonites.

Key words: Seismic refraction, mylonites, fractures, seismic reflection, low velocity zone, microcracks.

Introduction

The deformation of rocks due to faulting causes changes in lithology, pore pressure, and seismic velocity, all of which are evident from geophysical measurements. The internal properties of fault zones may provide information on the manner in which deformation takes place, the slip behavior of faults, and variations in the depth of seismogenic zones. Fault zones are likely regions for successful earthquake-prediction measurements, since they are expected to exhibit premonitory phenomena. Thus, an understanding of the internal properties of fault zones contributes to an understanding of the faulting process.

In this paper we review seismic measurements of fault zones and adjacent rocks. We discuss principally the in-situ compressional-wave velocity (V_p) structure of fault zones, since this parameter has been the most commonly measured; where data are available, we discuss the shear-wave velocity (V_s) properties, and we relate both

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these velocity determinations to laboratory measurements.

We proceed from the effects of intragrain microcracks and rock fractures associated with minor faults to fault gouge along major faults and, finally, to the seismic evidence of the internal constitution of faults in the lower crust. We present no new data, but we critically reëvaluate previously published results. We present evidence that, in the San Andreas fault of California, the internal properties of the continuously creeping sections of the fault differ markedly from the locked sections of the fault that fail catastrophically in great earthquakes. In the final section we use recent seismic reflection data to evaluate SIBSON's (1977) conceptual model that deep faults consist of mylonite zones.

Seismic velocities of fractured rocks

Microcracks

The effect of microcracks on the wave velocity of crystalline rocks has been studied in the laboratory and in boreholes. BIRCH (1960, 1961) showed in laboratory measurements that dry crystalline rocks' seismic velocities at low confining pressure are much lower than their intrinsic velocities (uncracked, high pressure) and that the closure of microcracks with increased confining pressure causes a rapid increase in P-wave velocity. DORTMAN and MAGID (1969) and NUR and SIMMONS (1969) later demonstrated that the seismic velocity of saturated granitic rocks is higher than that of unsaturated (dry) rocks, and that the rate of velocity increase with pressure is diminished, owing to the low compressibility of water. The important effect of water on the velocity of cracked rocks was also emphasized by SIMMONS and NUR (1968), who found higher velocities in situ at two wet boreholes than in laboratory measurements of dry samples. Water is available in most cases in situ, and WANG and SIMMONS (1978) found the in situ velocities in gabbroic rocks at a 5.3 km depth in the Michigan Basin to be about equal to the velocity of saturated laboratory samples at the appropriate pressure.

Macrofractures

A measurable velocity reduction is expected theoretically and observed experimentally in densely cracked and fractured media. O'CONNELL and BUDIANSKY (1974) considered the effects of cracks on elastic moduli for the case in which crack interaction is important. Their theoretical model predicts that for saturated cracks an increase in spatial crack density decreases both V_p and V_s and increases V_p/V_s. The lower velocity of densely cracked and fractured rocks has also been measured in situ. STIERMAN and KOVACH (1979) reported the results of logging a 600 m borehole in Stone Canyon, California, 1.2 km from the active trace of the San Andreas fault.

The well penetrates wet quartz diorite, ranging in condition from extremely weathered and friable to hard and unweathered. Their logs showed velocities that are 30 to 500 percent lower than the velocities commonly measured in saturated granite. They concluded that the low velocities are due to the fact that the macrofractures in the well are held open by tectonic shear stress associated with the San Andreas fault. They contend that their conclusion is supported by the occurrence of microearthquakes near the borehole, indicating that the rocks are near failure.

The relationship between macrofracture density and V_p and V_s was further investigated by MOOS and ZOBACK (1983), who reported sonic logs from four deep boreholes (0.6 to 1.2 km) drilled in granitic rocks. In wells with a relatively low density of macroscopic fractures the velocity in situ is similar to that of a saturated core measured in the laboratory (about 5.9 km/sec.). With dense fracturing, however, they found that V_p and V_s are very much reduced and V_p/V_s increased relative to laboratory measurements. Significantly, of the two boreholes with low fracture density the hole with the higher density exhibited a higher sonic velocity. This observation demonstrates that velocity is not directly proportional to fracture density, but that other factors, such as rock composition, play important roles as well. MOOS and ZOBACK (1983) also found lower velocities in boreholes from unfractured rocks adjacent to the macrofractures, which they interpreted as indicating chemical or mechanical alteration of these rocks. In those intervals of the borehole that are highly fractured it was not possible to measure the S-wave velocity, because of the attenuation of the waveform.

CHRISTENSEN (1984) and CHRISTENSEN and WANG (1985) reported laboratory measurements of the effects of pore pressure on basalts and on Berea sandstone. They demonstrated that, at a given confining pressure, the shear-wave velocity is more sensitive to increased pore pressure than is the compressional-wave velocity, thereby causing increases in V_p/V_s and Poisson's ratio. Thus, high pore-pressure zones may potentially be identified on the basis of Poisson's ratio, in the crust, where both shear and compressional waves can be measured.

Besides depending on crack density and degree of saturation, velocity variations in fractured rocks will depend on the preferred fracture orientation. If oriented fractures remain open, seismic velocities will be anisotropic, with lower P-wave velocities perpendicular to the cracks (CRAMPIN, 1978; HUDSON, 1981); if fractures are completely saturated, theory predicts a 4θ dependence, with slow P waves parallel and perpendicular to the fractures. CRAMPIN (1984a and 1984b) discussed in detail the key theoretical and observational aspects of the propagation of waves in anisotropic media. Velocity anisotropy for P waves around a borehole drilled in crystalline rock in a geothermal field was reported by LEARY and HENYEY (1985). They interpreted the anisotropy as arising from the existence of fractures in the basement having an orientation subparallel to the latest episode of local faulting. BAMFORD and NUNN (1979) presented additional examples of seismic methods of locating fracture zones, and CRAMPIN et al. (1980) estimated crack parameters by using synthetic seismograms to model shear-wave polarization anomalies.

Fault zones: the San Andreas

Whereas cracked and fractured rocks are found in association with minor faults and adjacent to major faults, the near-surface portion of a major fault zone is often observed to be filled with 10 to 100 m of fault gouge due to severe alteration and hydration of the rocks. This is the case within the San Andreas fault zone, which in some locations is filled with severely altered rocks containing a significant amount of clay minerals. Since it is known from laboratory studies that fault gouge has a lower velocity than its protoliths (WU, 1978; WU et al., 1975; WANG et al., 1978), seismic methods are effective in delineating the extent of alteration within the fault zone.

Along the linear stretch of the San Andreas fault system (Fig. 1) south of Hollister, California, a substantial seismic low-velocity zone (LVZ) has been determined (HEALY and PEAKE, 1975; AKI and LEE, 1976; FENG and MCEVILLY, 1983). In the model of HEALY and PEAKE (1975) the LVZ is 3 km wide and 12 km deep, with velocities ranging from 3.0 to 4.5 km/sec, compared to 5.2 to 6.5 km/sec outside the fault zone. The detailed nature of the velocity distribution within this zone is indeterminate. The low average velocities determined for this zone may actually represent the velocity throughout the zone. Alternatively, it is possible that very high-resolution refraction measurements (e.g., SJOGREN, 1984) would reveal that the LVZ medium with an average velocity of 4.0 to 4.5 km/sec actually consists of narrow zones with velocities as low as 3.0 km/sec alternating with largely unfractured rocks with velocities approaching 6.0 km/sec. This model is appealing, since gouge zones at the surface are always much thinner than the 3 km LVZ we have discussed.

Some discrepancies exist in the depths to which fault-zone LVZ's penetrate the crust. Along the portion of the San Andreas fault system discussed above, AKI and LEE (1976) found no evidence of low velocities at depths below 5 km, whereas HEALY and PEAKE (1975) report them at depths of 12 km at the same location, and FENG and MCEVILLY (1983) extend the LVZ to the Moho.

From seismic refraction measurements there is general agreement that along the Calaveras fault zone (Fig. 1) an LVZ exists within the fault, with velocities of 3.0 to 4.0 km/sec extending to a depth of 6 km (MAYER-ROSA, 1973; MOONEY and LUETGERT, 1982; BLÜMLING et al., 1985; MOONEY and COLBURN, 1985). Fan profiles recorded across the fault indicate that the seismic low-velocity zone is 1 to 2 km wide at the near-surface (BLÜMLING et al., 1985). At this same location, however, THURBER (1983) found no evidence of low velocities in an inversion of earthquake travel times (Fig. 2). To resolve the discrepancy between the results of THURBER (1983) and the seismic refraction results, CORMIER and SPUDICH (1984) calculated synthetic seismograms for earthquakes from a model of the Calaveras fault which included a deep wedge-shaped LVZ (Fig. 3). Their calculations successfully modeled

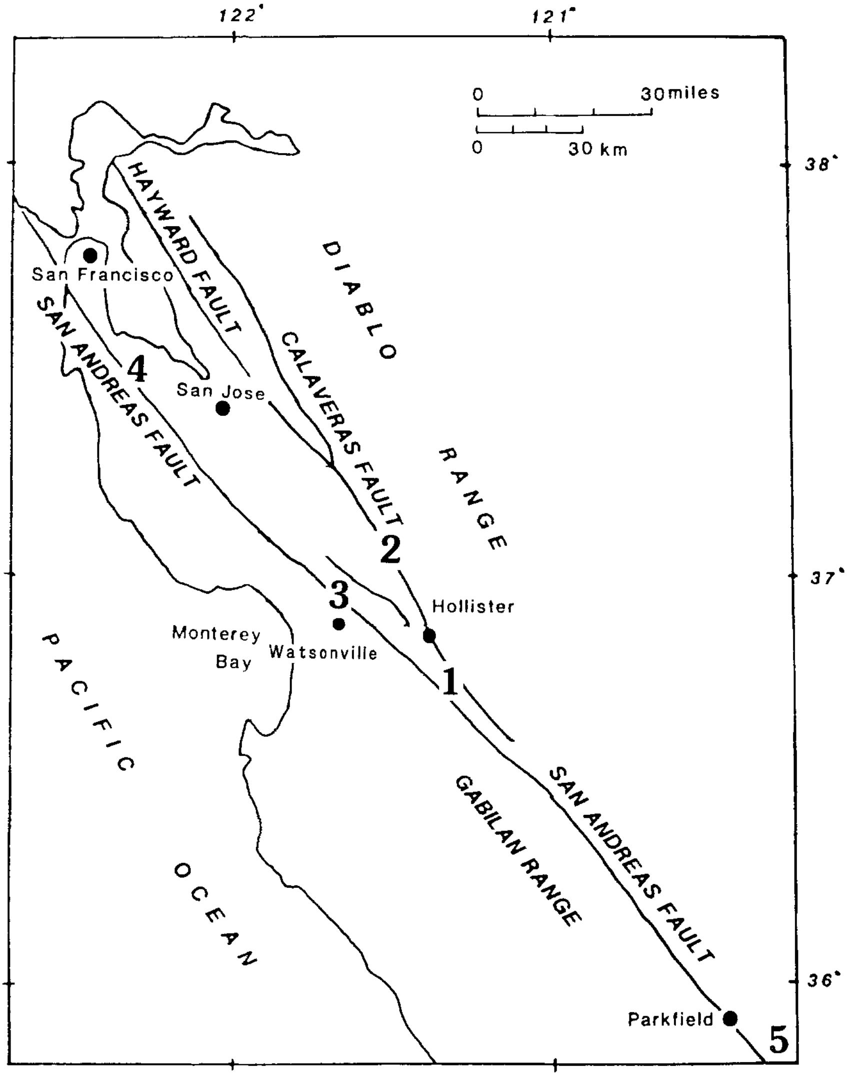

Figure 1

Location map of studies of fault-zone structure in central California. The San Andreas fault is seismically active (creeping) from Parkfield to Watsonville. The Calaveras fault is active from Hollister to the latitude of San Francisco. Numbers next to faults refer to seismic studies discussed in text. 1: AKI and LEE (1976), HEALY and PEAKE (1975), FENG and MCEVILLY (1983). 2: MAYER-ROSA (1973), MOONEY and LUETGERT (1982), THURBER (1983), CORMIER and SPUDICH (1984), MOONEY and COLBURN (1985), BLÜMLING et al. (1985). 3: THURBER (1983), MOONEY and COLBURN (1985). 4: BOKEN and MOONEY (1982). 5: TREHU and WHEELER (1986), MEREU (1986).

the seismic energy trapped and focused within the wedge, leading to locally high amplitudes directly over the fault (Fig. 3). Thus, their model contradicts Thurber's

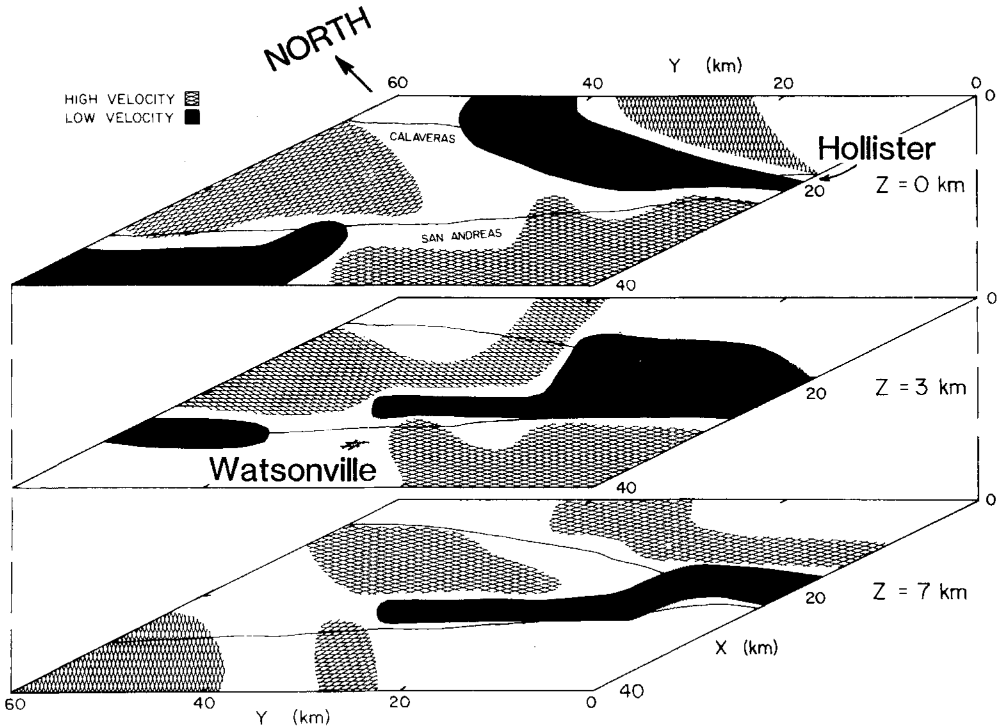

Figure 2

Seismic-velocity inversion results for an area in central California, from THURBER (1983). The Calaveras and San Andreas faults are labeled, as are Hollister and Watsonville. Areas of high and low velocity are indicated at three depths: 0, 3, and 7 km. At 7 km a low-velocity zone follows the San Andreas fault as far north as Watsonville; such a zone is absent along the Calaveras fault at that depth. Along the San Andreas fault the low-velocity zone is consistent with other seismic studies; the lack of one on the Calaveras fault is inconsistent with these studies. See text for discussion.

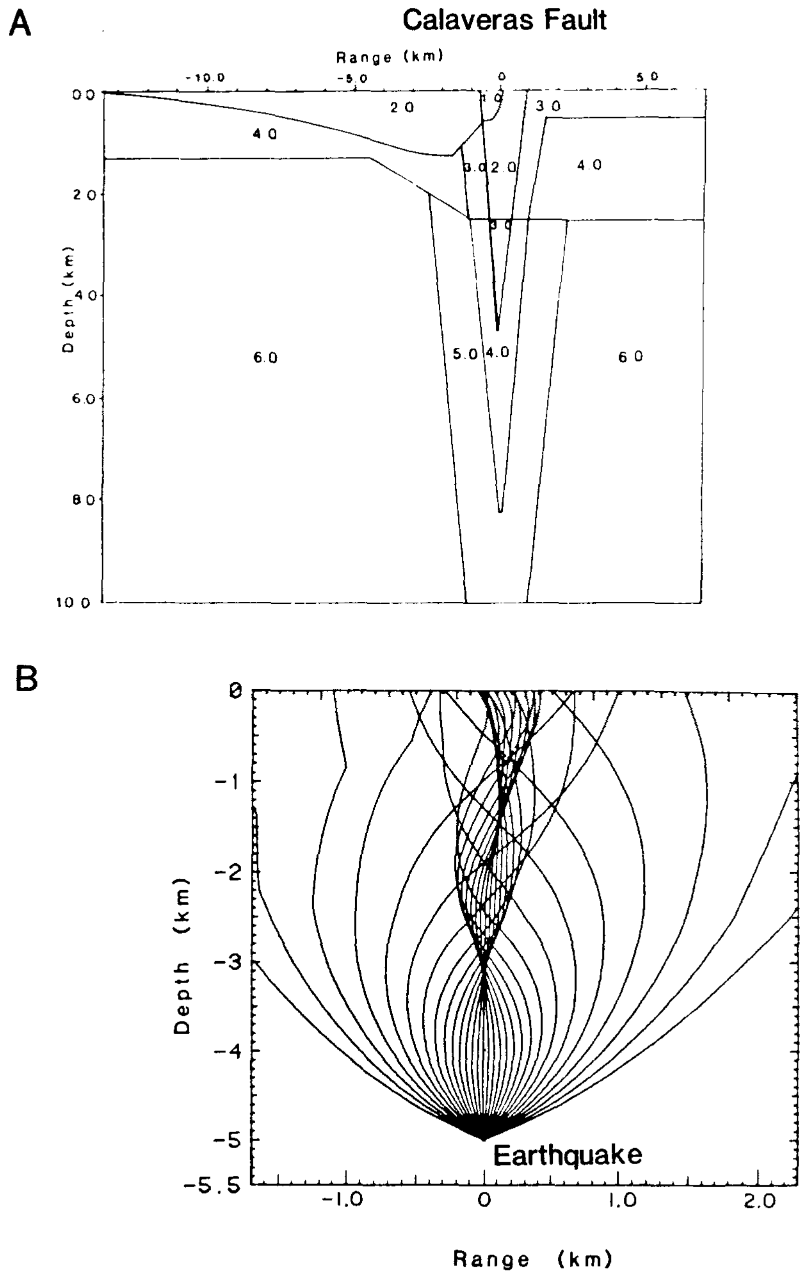

Figure 3

(A) Velocity model across the Calaveras fault, used by CROMIER and SPUDICH (1984) to model earthquake seismograms recorded over the fault; the fault model consists of a pronounced low-velocity zone to a depth of 10 km. (b) Diagram of ray paths propagating from a depth of 5 km through model in (A); note ray multipathing and high ray density at the fault, correctly predicting local site amplification. This modelling supports the existence of a substantial low-velocity zone along the Calaveras fault.

conclusion that an LVZ is lacking along this portion of the Calaveras fault. An important aspect of their study was that it indicated the potential for deriving the properties of the fault zones from dense array recordings of microearthquakes.

The discrepancies among the seismic interpretations discussed above are due largely to the resolutions inherent in the methods and to the varying spatial densities in the data. The studies of AKI and LEE (1976) and THURBER (1983) were based on the inversion of earthquake travel times to a permanent array of seismograms. The resolution of these inversions is controlled by several factors, particularly station and hypocentral distribution. In general, features less than 2 km in width will not be resolved by this method.

The seismological evidence summarized above indicates that the creeping sections of the San Andreas fault system may be characterized by a pronounced seismic LVZ. WANG (1984) considers geologic, geophysical, and geochemical evidence for the composition of the LVZ material. He shows that the seismic velocity models which show a deep LVZ along the central San Andreas fault are consistent with laboratory measurements of fault-gouge materials at elevated pressures. Electrical resistivity values for the fault zone are consistent with a gouge composition. Citing the laboratory results of CHRISTENSEN (1966, 1978), WANG (1984) concludes that the seismic velocity of serpentine is higher than the velocities reported for the fault zone. This contrasts with earlier suggestions that the San Andreas fault in central California is rich in serpentine at depth (ALLEN, 1968). WANG (1984) concludes from this and additional geologic and geochemical evidence that throughout its seismogenic regime (0 to 15 km) the creeping portion of the fault in central California is composed of saturated fault gouge. WU et al. (1975) and WANG et al. (1978) reached the same conclusion from a smaller amount of evidence. WANG (1984) is cautious about extending his deep fault-gouge model to outside the creeping section of the San Andreas fault, but he admits the possibility that it is applicable in those other regions. We examine this possibility below.

Several studies indicate that the San Andreas fault does not have a substantial LVZ along the locked portions of the fault. Although the inversion results of THURBER (1983) do show a significant deep LVZ running just northeast of the San Andreas fault in the area of Hollister (Fig. 2), further northwest along the San Andreas fault the LVZ terminates approximately where creep and seismic activity cease, near Watsonville (Fig. 2). The lack of an LVZ on the fault north of Watsonville is confirmed by two seismic refraction profiles across the fault: neither a profile just north of Watsonville (MOONEY and COLBURN, 1985) nor one 40 km further north (BOKEN and MOONEY, 1982) detected an LVZ at the fault.

Across the locked portion of the San Andreas fault south of Parkfield, California, a seismic refraction profile has been variously interpreted as being with (TREHU and WHEELER, 1986) and without an upper crustal LVZ (MEREU, 1986). Thus the available evidence may indicate that substantial LVZ's do not occur along locked sections of the fault, and accordingly we hypothesize that *substantial* LVZ's occur only on creeping sections of the fault. It appears that LVZ's are conducive to fault creep—for example, through high pore pressures due to the high water content.

Two observations contradict our hypothesis that substantial LVZ's occur only along the creeping section of the fault. First, due to the volumetric increase (and density decrease) that accompanies its formation, fault gouge often upwells along the fault. The resultant weltlike ridge is readily visible along many portions of the fault, such as the locked section in the Carrizo Plain (WALLACE, 1949; ALLEN, 1968, 1981). Small amounts (30 m) of surficial fault gouge are also reported along the locked portion of the fault north of Watsonville (HALL, 1984). These observations, however, do not demonstrate that a thick zone of fault gouge exists at depth, only that some gouge exists in the near-surface.

The second observation indicating thick fault gouge at depth is the narrow, elongate Bouguer gravity low which is nearly coincident with much of the fault trace. Several investigators have modelled this gravity low as due to a low-density zone (presumably fault gouge) 10 km deep and between one and several kilometers wide (PAVONI, 1973; WANG et al., 1978; STIERMAN, 1984). Of these studies, only that of STIERMAN (1984) concerns a locked portion of the fault.

STIERMAN (1984) considered the Bouguer gravity data along with borehole and other geophysical evidence and concluded that fracturing and chemical alteration exist in a wide zone along the fault that extends to seismogenic depths. He demonstrates that the altered zone is not limited to a single narrow trace a few tens or even hundreds of meters wide, that rocks a few kilometers from the primary dislocation surface are also modified to some extent by the earthquake process.

To evaluate the properties within the fault zone, STIERMAN (1984) modeled gravity data across the San Jacinto fault of southern California at locations where the observed lows are clearly not attributable to surficial basins. He modeled these lows with a series of crustal prisms 10 km deep by 1 to 2 km wide with density reductions of 0.1 to 0.025 gm/cm^3 on either side of the fault, and a zone 200 m wide with a density reduction of 0.4 gm/cm^3 directly at the fault. This model, with several kilometers of low-density prisms, appears to be consistent with a pronounced LVZ at this locked portion of the fault, and STIERMAN (1984) concluded that a fault-zone LVZ is probably not confined to the creeping portion of the San Andreas fault. To test this hypothesis we convert the density model to a velocity model, using an approximation to a velocity–density curve (LUDWIG et al., 1970):

$$\Delta V_p = 3.33\ \Delta\rho$$

This relationship yields a velocity reduction of only 0.33 km/sec corresponding to a density reduction of 0.1 gm/cm^3, and a velocity reduction of 1.33 km/sec corresponding to a density reduction of 0.4 gm/cm^3. Thus, if a *pronounced* LVZ is indicated by a velocity reduction of 0.5 to 1.0 km/sec or greater (such as found along creeping faults), then STIERMAN's model (1984) implies an LVZ that is only 200 m wide across the San Jacinto fault, as compared to one that is 3 to 4 km wide along the creeping

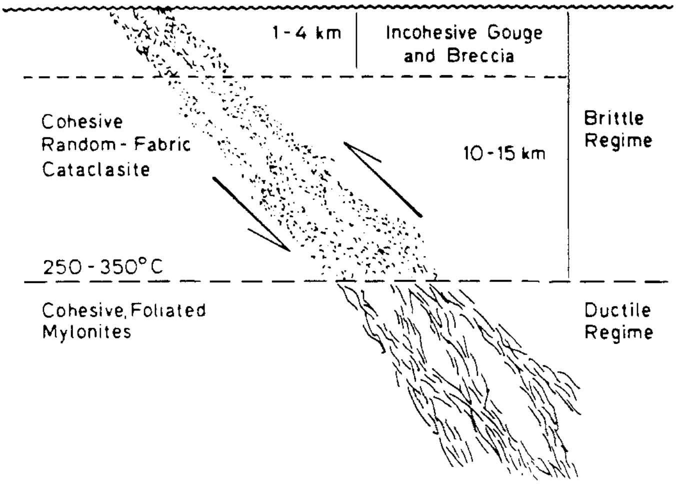

Figure 4
Geologic model of a deep continental fault (SIBSON, 1977). In the ductile regime, cohesive, foliated mylonites are expected.

segment of the San Andreas fault in central California.

We note that our hypothesis of a substantial LVZ along creeping sections of the San Andreas fault is not inconsistent with STIERMAN's hypothesis (1984) that rocks adjacent to locked portions of the fault are fractured and locally altered; it is merely that the velocity reduction there will be significantly less than within the pronounced LVZ found on the creeping central San Andreas fault.

In conclusion, a substantial body of geologic and geophysical evidence is consistent with an LVZ as wide as 3 to 4 km along the creeping section of the San Andreas fault, and a narrow LVZ (200 m or much less) along the locked sections. Thus, fault slip behavior correlates with the internal properties of the fault zone, and we speculate that high pore pressure within thick LVZ's is conducive to fault creep. Important questions raised by this hypothesis are whether locked faults evolve into creeping faults as gouge is developed and, indeed, whether fault creep creates fault gouge or fault gouge leads to creep.

Deep structure of faults

On the basis of geological and physical considerations SIBSON (1977) proposed a conceptual model of the deep structure of faults. In his model (Fig. 4) fault gouge and breccia are formed in the upper 1 to 4 km of the crust, where brittle processes dominate and water is available to alter the crushed rocks into clay-rich gouge. At depths greater than 4 km, low-porosity cataclasites are formed as a result of elasticofrictional deformation. These rocks generally have random fabrics and reduced grain size. At depths greater than about 15 km and temperatures of about 300 °C ductile processes dominate and mylonites are formed. The mylonites in this quasi-plastic zone are well foliated and show tectonically induced grain-size reduction. The depth of the boundary between the elastic and the quasi-plastic styles of deformation will be determined by several factors, including heat flow, presence of fluids, and composition of the rocks. Numerous seismic reflection and refraction profiles completed in the course of crustal studies and exploration activities have detected faults, and in this study we summarize the seismological evidence for the internal structure of fault zones and use these to evaluate SIBSON's model.

SIBSON (1977) hypothesizes that both vertical and low-angle faults should in some instances penetrate to the lower crust. The primary evidence he cites for deep faults is the existence of mylonites in ancient fault zones that have been uplifted and exposed. In many cases the mineral assemblages and metamorphic grades of the rocks within and enclosing the fault zone give unquestionable evidence that they come from deep crustal levels. Given this geologic evidence, we consider how deeply into the crust faults can be traced on seismic reflection profiles. A profile recorded northwest of Scotland shows a reflection that projects through the crust nearly to the Moho (Fig. 5). PEDDY (1984) shows that after correction for a near-surface sediment wedge the apparently reactivated thrust penetrates through the entire lower crust and into the uppermost mantle. Another reflection profile across the Wind River mountains of Wyoming shows a low-angle fault that can be traced from the surface to depths approaching 32 km (SMITHSON et al., 1979; LYNN, 1979). Given these examples, the question remains whether these seismic observations can be explained by the presence of mylonites (SIBSON, 1977) within faults.

The origin of reflections from deep faults was considered by JONES and NUR (1984) in a combined analysis of field observations of ductile fault zones, laboratory measurements of the physical properties of mylonites, and seismic modelling of fault-zone velocity structures. Their field observations show that the mylonites are characterized by both compositional banding and strong foliation parallel to the boundaries of the shear zone. Velocities were measured in mylonite samples recovered from the fault. No significant difference in average velocity and density, between the mylonites and adjacent rocks, was found, but a seismic anisotropy of 7 per cent or greater was found in several mylonites. In particular, quartz and phyllosilicates are readily deformed ductilely to effect a strong preferred orientation of crystallographic axes. Phyllosilicates have a much greater single-crystal anisotropy than does quartz, so the phyllosilicate composition and fabric will control the seismic anisotropy. JONES and NUR (1984) showed, using synthetic seismograms, that the anisotropy and finely laminated structure in the lower crust could produce strong reflections from mylonite zones.

FOUNTAIN et al. (1984) also assessed the evidence that mylonite zones are the cause of deep crustal reflections. They note that the shallow dips (less than 30°) and lateral continuity of these zones are conducive to producing laterally continuous,

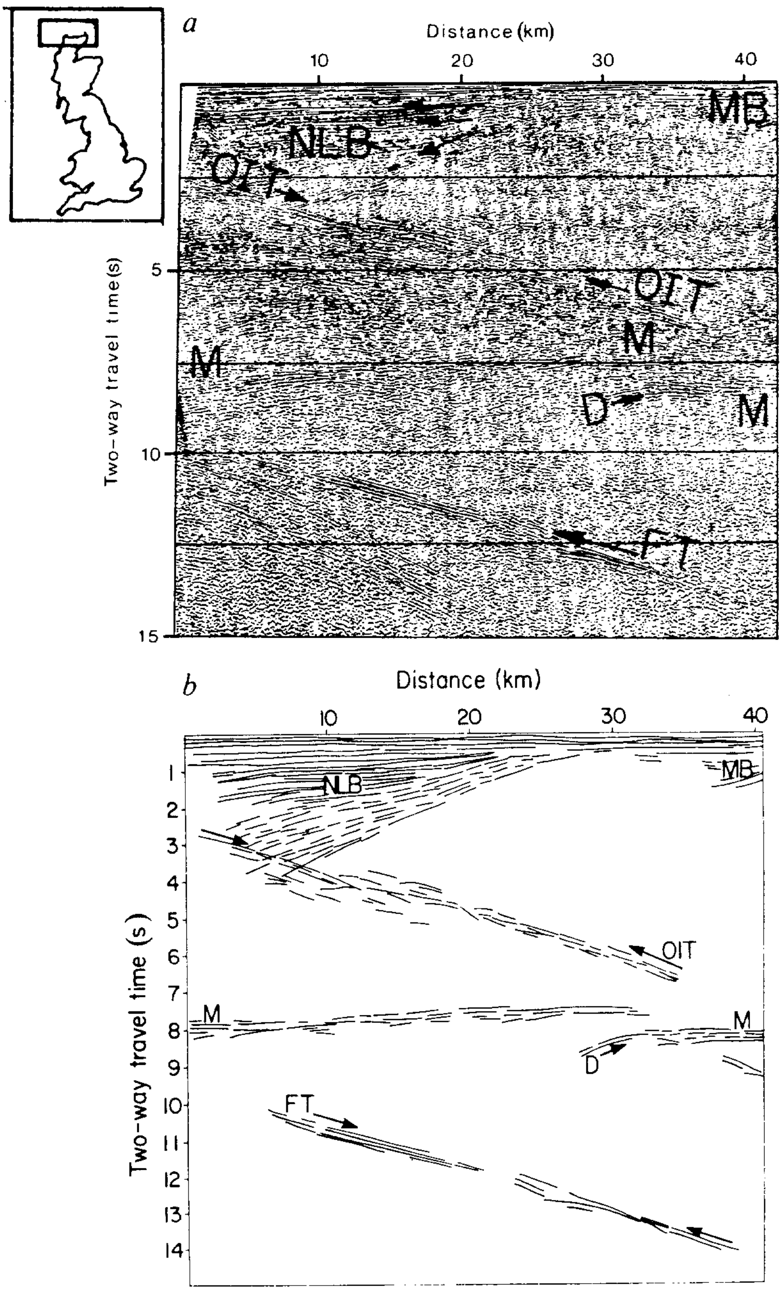

Figure 5
(a) Deep seismic reflection data from a marine survey north of Scotland (Brewer et al., 1983): NLB, North Lewis Basin; MB, Minch Basin; OIT, Outer Island Thrust; M, Moho; D, diffraction; FT, Flannel thrust. (b) Line drawing of main events in (a) above. The OIT appears to reach almost to the Moho; PEDDY (1984) shows that it does in fact reach and cross the Moho, after data corrections are made for the surficial basins. The Flannel thrust indicates that faults exist as well in the mantle.

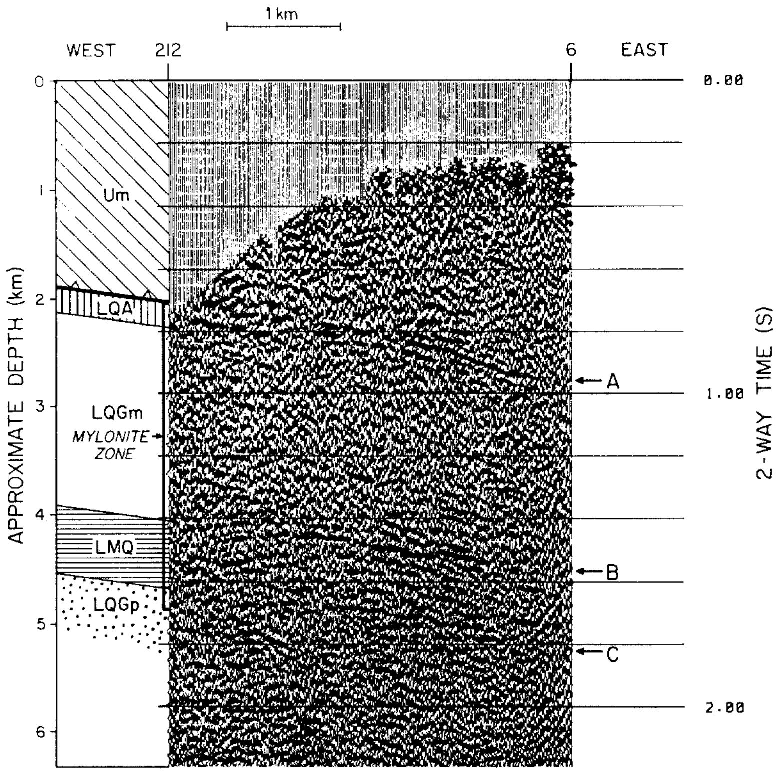

Figure 6
Seismic reflection section showing a correlation between the downdip projection of an exposed mylonite zone and seismic data (HURICH et al., 1985), 1:1 scale: A, reflection from fault and upper part of mylonite zone; B, reflections from interlayered metamorphic rocks; C, reflection from top of crystalline basement; Um, upper-plate marble, phyllite; LQA, lower-plate quartzite, ampphibolite; LQGm, lower-plate quartzofeldspathic gneiss; LMQ, lower-plate marble, quartzite, schist; LQGp, lower-plate porphyroclastic gneiss.

correlatable reflections. They also note that many mylonite zones exhibit either a laminated or anastomosing geometry. Both geometries produce relatively high-amplitude seismic reflections where constructive interference occurs. Furthermore, field and petrographic studies demonstrate that mylonitic zones commonly exhibit strong retrograde metamorphic effects, compared to undeformed zones. The effect of this retrogression will be to lower the velocity of mylonitic zones relative to neighboring rocks, or to enhance the seismic anisotropy by the synkinematic growth of new phyllosilicates (FOUNTAIN et al., 1984).

A direct demonstration of the seismic response of a mylonite zone was obtained by HURICH et al. (1985), who recorded a seismic reflection profile over an exhumed zone in eastern Washington State. The presence of sillimanite within the zone indicated a minimal depth of about 15 km for the deformation in the fault zone. A downdip projection of the exposed mylonite zone into the seismic profile indicated that the compositionally layered mylonite is reflective and traceable to a depth of at least 5 km, as a complex zone of multicyclic reflections (Fig. 6).

In summary, deep seismic reflection data sometimes show pronounced reflection events that penetrate the entire crust. These events consist of multicyclic reflections with relatively high amplitudes and good lateral continuity. Modelling of these events and field tests confirm that fault zones consisting of mylonite zones 0.5 to 2 km thick in the crust can explain these observations. The properties of mylonite zones that appear to make them reflective include velocity anisotropy, laminated or anastomosing geometry, and retrograde metamorphism.

Summary

The deformation of upper crustal rocks due to faulting may be described in three levels, corresponding to successive reductions in the seismic velocities of the rocks. Faulting at the lowest level leads to microcracking and limited macrofracturing of the rocks and does not have a significant effect on their seismic velocities, particularly if the rocks are saturated, as is generally the case in situ. The middle level, densely fractured rocks, such as are found adjacent to major strike-slip fault zones, show greatly reduced V_p and V_s and increased V_p/V_s. If the fractures are oriented by a deviatoric stress field, velocity anisotropy results and can be determined with precise velocity measurements and shear-wave polarization analysis. At the third level, large amounts of continuous deformation produce fault gouge which is generally 30% to 40% lower in seismic velocity than is the protolith, at a given confining pressure. Fault gouge therefore is easily detected within the fault zone by seismic measurements.

Seismic studies in central California have identified densely fractured rocks adjacent to the San Andreas fault and a pronounced low-velocity zone (LVZ) extending to seismogenic depths along the creeping segment of the fault. This LVZ has been attributed to deep fault gouge, and WANG (1984) presents geologic and geochemical evidence that fault gouge is stable to depths of 10 to 12 km in the crust, which lends credence to this hypothesis. Evidence is presented here that the creeping segments of the San Andreas fault in central California contain thick LVZ's, whereas the locked segments of the fault have either no LVZ or very thin ones (200 m or much less). Thus the slip behavior along segments of the fault appears to be related to the presence or absence of an LVZ along the fault, and it is possible that high pore pressure within thick LVZ's is conducive to fault creep.

Faults may be traced locally on seismic reflection sections into the lower crust. According to SIBSON's model (1977) for continental fault zones, deep faults consist of mylonitic rocks. Recent seismic reflection data and computer modelling of these data are consistent with this model.

The discussion of the observations of fault-zone properties given here indicates that the study of the internal properties of fault zones is at a rather rudimentary stage. Particular attention needs to be paid to shear-wave velocities and the seismic quality factor (Q). Progress may be expected with continued interpretation of seismic reflection and refraction data, borehole measurements, and array recordings made with three-component sensors deployed directly across and along fault zones.

Acknowledgments

Critical reviews of an early draft by M. C. Andrews, S. Klemperer, and A. W. Walter significantly improved this paper. Discussions with S. Peacock, J. McCarthy, and E. L. Ambos were helpful. Review comments by N. I. Christensen, A. Lindh, T. V. McEvilly, and R.-S. Zeng (Beijing) are greatly appreciated.

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(Received 5th November 1985, revised 20th March 1986, accepted 23rd March 1986)