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Speculations on the Geometry of the Initiation and Termination Processes of Earthquake Rupture and its Relation to Morphology and Geological Structure

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Abstract—Earthquake initiation and termination processes are commonly described in terms of barriers and asperities. Barriers fall into two classes: Geometric barriers are associated with places where the orientation of a failure surface changes, and relaxation barriers, where stress is low because aseismic creep processes outpace tectonic loading. Geometric barriers fall into conservative and nonconservative subgroups, according to whether finite fault motion can proceed without the creation of new structures or whether it demands the creation of new faulting or void space. The multiple faulting, or 'fragmentation', associated with some nonconservative barriers can disrupt fault planes and form asperities. By means of selected examples it is shown that a description in terms of these barriers can help one to visualise the processes of earthquake rupture and its relation to the geological environment.

Key words: Fault geometry, earthquake rupture.

Introduction

This paper discusses ideas about barriers that have proved to be key concepts in the development of our understanding of earthquake processes in the last ten years (DAS and AKI, 1977; AKI, 1979). The relation of barriers to asperities (KANAMORI, 1978) is also considered. Most of the earlier work has considered the problem from the perspective of seismogram modelling, or of developing models relating laboratory studies of rock friction to the dynamics of rupture on faults (e.g., RICE, 1980). Here the purpose is to emphasize what may be understood from geometrical considerations alone.

The description of barriers adopted has been discussed by KING and YIELDING (1983), who used the El Asnam earthquake as an example, and by KING (1983) and KING and NABELEK (1985), who considered the application of self-similar geometries (fractal geometries) to the behavior of fault systems.

The concepts have implications for the relation between repeating earthquakes and the formation of geological structures and growth of morphological features

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(KING and VITA-FINZI, 1981; CISTERNAS et al., 1982; KING and BREWER, 1983; KING and STEIN, 1983; STEIN and KING, 1984; STEIN, 1985). The relation of these ideas to the barrier classification is illustrated in this paper.

The views presented here, on the role of fault bends, are implicit in the work of BAKUN and MCEVILLY (1979), SYKES and SEEBER (198), and BAKUN and MCEVILLY (1984). The discussion by LINDH and BOORE (1981) of the geometry of faulting associated with the 1966 Parkfield earthquake, extended by BAKUN and LINDH (1985) in the context of earthquake prediction, also emphasizes the role of bends in controlling rupture in individual and repeating earthquakes.

Fault jogs or offsets may be regarded as consisting of one or more bends of opposite sign. It is difficult to believe that motion in the brittle zone transfers between segments by continuum processes, except in the short term. Sets of bends associated with jogs or offsets are considered in this paper to behave collectively in a manner similar to individual bends (see also: KING, 1983; KIING and NABELEK, 1985).

Characterization of barriers

Barriers may be classed according to whether they are a consequence of fault geometry or of rock properties.

A planar fault is associated with two vectors, the slip vector and the vector normal to the fault plane. Where a fault meets a second fault or some other structure that may also be described by two vectors (e.g., dykes, sills, microcracks, and the Earth's stress-free surface), the nature of that junction may be classified according to the relations between the four vectors. This forms the basis for classifying *geometric* barriers.

Relaxation barriers result from the properties of the rock or fault material. Only the most basic consequence of these rock properties needs to be considered: Can the material relieve stress by creep as fast as, or less fast than, the tectonic loading can restore it? If stress cannot accumulate in the long term, then rupture can propagate only as a result of rapid slip in an adjacent stressed region. Such low stress regions form relaxation barriers; their properties have been discussed by HUSSEINI et al. (1975).

In the following discussion we consider barriers in pairs for the convenience of referring to the rupture zone between them as the *main* fault. Barriers, of course, need not occur in pairs of the same type, and a barrier will, in general, have both geometric and relaxation characteristics. Furthermore, the assumption that the main fault is simple and planar is correct only in broad terms.

A large fault presumably consists of many barriers of different types operating over a range of scales. Thus, although clear examples of the types of barrier discussed can be found, real earthquakes usually are more complex geometrically than the simple models. Some clear examples are discussed here.

Geometric barriers

Geometric barriers occur where the normal to a failure surface changes. If we allow the slip vector to change also, nine possible combinations can occur. Thus eight variations from one plane and slip direction are possible. Not all of these are easy to represent pictorially or appear to be physically reasonable, and the following discussion is confined to the geometries illustrated in Figure 1. These fall into two

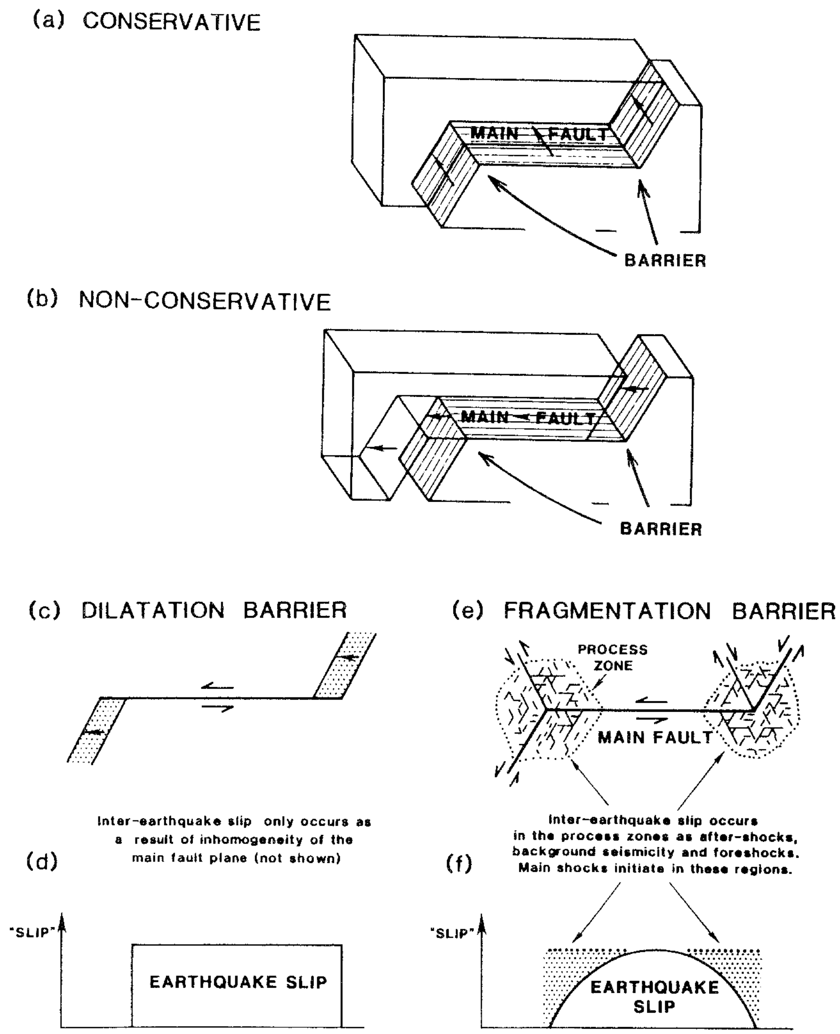

Figure 1
Geometric barrier pairs: (a) conservative barriers; (b) nonconservative barriers; (c) dilatation barriers that occur when confining pressure is relatively low or voids are small; (d) slip amplitude at main fault ends is not constrained by dilatation barriers; (e) fragmentation barriers occur under conditions of higher confining pressure; (f) slip amplitude for fragmentation barriers must taper and become zero at each end of main fault.

categories, the *conservative* and the *nonconservative* barriers. In conservative barriers (Fig. 1a) the slip vector lies in both failure planes, and slip can occur without either volume change or the creation of new faulting. It has been conjectured that such barriers can briefly arrest but not terminate dynamic rupture (KING and YIELDING, 1983). Thus they can produce features on a seismogram but they are not significant in the initiation and termination processes of earthquakes.

Nonconservative barriers (Fig. 1b) require either a volume change or the creation of new faulting, and on this basis they may be divided into *dilatation* and *fragmentation* barriers (Fig. 1c and 1d respectively). The former occur where faults end on open or fluid-filled voids. Examples on a large scale are where transform faults join ocean ridges or where faults meet dykes, sills, magma bodies, or the Earth's stress-free surface. On a small scale, fracture can terminate in pore space or microcracks.

Figure 1d shows the form that the slip function on the main fault can take between dilatational barriers. The barriers do not require the slip to taper at the fault ends, and finite motion can occur without the creation of new structures or the development of large strains in the medium around the fault. Earthquakes terminate because the features at the fault ends are free-moving.

For this geometry earthquakes can initiate anywhere on the fault, because no part of the fault is left with a slip deficit as a result of a tapered slip function. The significance of this will become apparent when other barriers are considered.

Fragmentation barriers occur under conditions of confining pressure such that large voids cannot open (Fig. 1e). The slip function on the main fault is constrained to be zero at both ends (Fig. 1f), and although some motion can be accommodated elastically, finite motion must be accommodated by further faulting. The multiple faulting, in the long term, must be sufficient to accommodate the slip that cannot be accommodated on the main fault. Thus, in the absence of creep processes—that is, the bend is without a relaxation character—the deficit of slip (indicated by shading in Fig. 1f) must be accommodated by aftershocks, background seismicity, and foreshocks. When a slip deficit or moment-release deficit occurs at the ends of the main fault, a consequence is that these regions will maintain a higher mean stress than the main fault segment. Therefore earthquakes may be expected to begin at or near bend regions.

The fragmentation process cannot proceed indefinitely with ever-decreasing scale without the creation of an infinite fault area (KING, 1983). At some scale an opening must occur, and thus the fragmentation process must be associated with many small dilatational barriers. SIBSON (1986) describes this process specifically in terms of fault offsets. Some component of dilatation, however, is an inevitable property of any fault bend and not a specific one of offsets or fault jogs.

Relaxation barriers

Relaxation barriers occur where creep processes can relax tectonic loading on a part of a fault system as fast as it is applied. The creep may be localized at the fault (Fig. 2a) or occur in a volume (Fig. 2b). (The principal difference between the two is not in the nature of the earthquake rupture but in the geodetic and geological deformation associated with fault motion.) For short time periods, certainly for the times involved in seismic rupture, the faults and surrounding rock must behave in an elastic-brittle fashion. For a relaxation barrier to occur, the stress relaxation time must be short compared with the earthquake repeat time and long compared with the rupture time. Since these differ by more than eight orders of magnitude, rocks of widely varying rheological properties will behave in essentially the same way. When an earthquake occurs, seismic slip extends outside the loaded zone (A–B in Fig. 2) into the creeping region to produce a tapered slip function of the form shown in Figure 2c. Rupture is constrained because of the low stress in the creeping region, but nonetheless it can extend well beyond the zone A–B (see HUSSEINI et al., 1975).

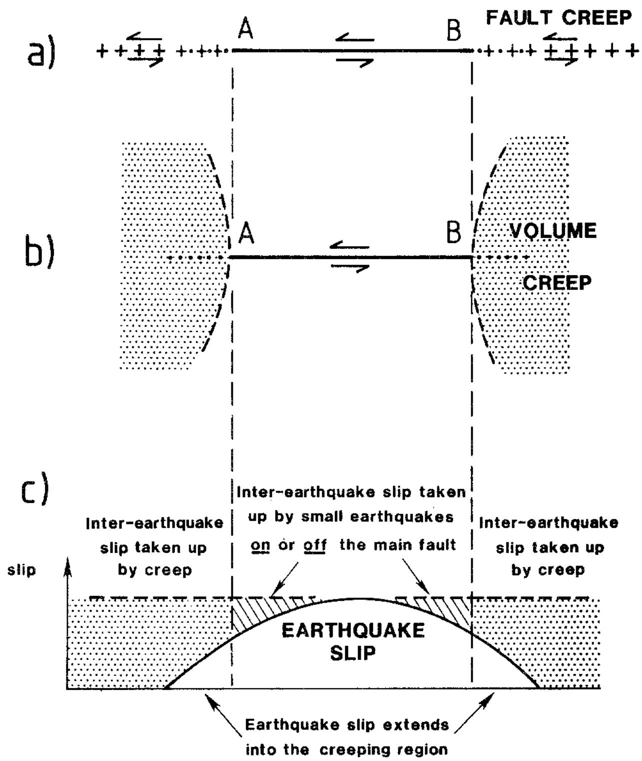

Figure 2
Relaxation barrier pairs: (a) barriers with fault creep; (b) barriers associated with volume creep; (c) slip function associated with either (a) or (b).

Because of the form of the slip function, a deficit of slip occurs along part of the fault in zone A–B. As creep proceeds outside this region after an earthquake, the zone is reloaded and the slip deficit will be taken up by aftershocks and foreshocks. In contrast to fragmentation barriers, though, there is no geometric requirement that events lie off the main fault. In common with fragmentation barriers, the amplitude of the slip diminishes at the ends of the fault. No seismicity need occur in the creeping region. For a creeping fault (as opposed to a creeping volume), however, this will be true only if the whole fault is planar and creeps over its entire surface. For example, the low-magnitude seismicity of the creeping part of the San Andreas may be regarded as resulting from geometrical irregularities in the fault and (sticking) patches where the creep relaxation rate is less than the loading rate.

Changes of fault behavior with depth

Using the foregoing descriptions of barriers, we can discuss the behavior of faulting as a function of depth. The conditions for a strike-slip fault are shown

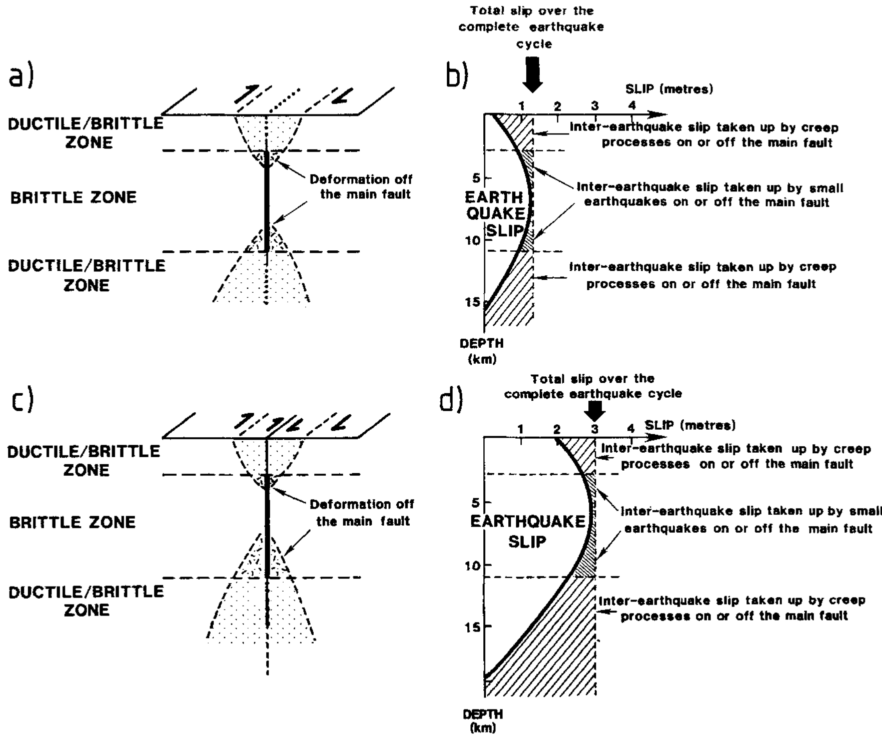

Figure 3

Fault behavior as a function of depth for strike-slip faults: (a,b) region where slip at depth does not greatly exceed 1 m; (c,d) where slip is close to 3 m at depth. In both cases an earthquake leaves a deficit of slip at the top and bottom of the brittle zone.

schematically in Figure 3. The upper figures (3a and 3b) are for a fault for which the repeating earthquakes have magnitudes of about 6.0 and scarcely form surface breaks, and the lower figures (3c and 3d) are for faults which have events between half and one magnitude greater and produce significant surface breaks. In a vertical direction two types of barrier occur. Relaxation barriers occur at depth and near to the surface. Data supporting the existence of these barriers are presented by KING and VITA-FINZI (1981), KING and BREWER (1983), VITA-FINZI and KING (1985), and EYIDOGAN and JACKSON (1985). It is proposed that near-surface stress relaxation occurs by chemical processes in joints and fissures, while at depth creep processes activated by temperature increasing with the geothermal gradient perform the same function. In the absence of magma or other fluid bodies (which terminate rupture on a dilatational barrier) rupture can continue downwards to an extent determined only by the magnitude of slip in the brittle layer. The observation that the rupture of large earthquakes apparently penetrates to greater depth than that of smaller ones was made by SCHOLZ (1982), although his original explanation differs slightly from that given here.

Slip can extend indefinitely downwards, but rupture above the brittle stressed zone can extend only as far as the dilatation barrier presented by the Earth's stress-free surface. For small earthquakes the upper relaxation barrier is sufficient to prevent slip from reaching the surface, and hence such earthquakes generally have

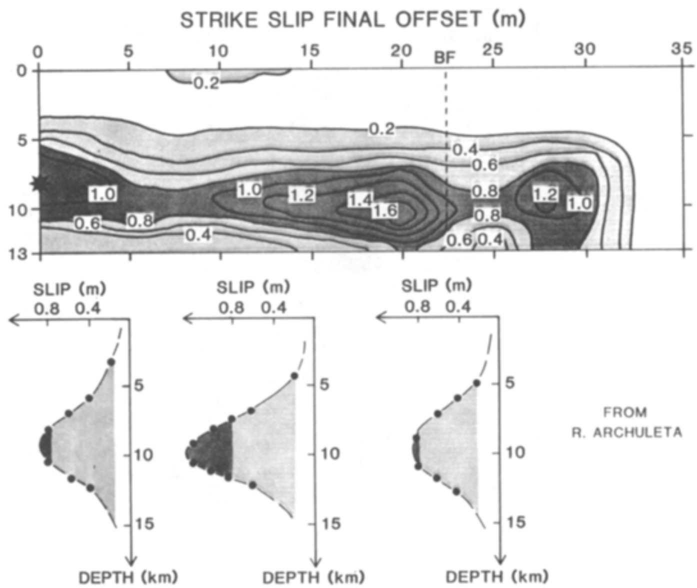

Figure 4
Slip distribution of the 1979 Imperial Valley earthquake. Contour plot: horizontal and vertical distances in kilometers, slip in meters; star indicates hypocenter, BF indicates the Brawley Fault. Lower graphs: 5, 25, and 25 km positions of upper figure. From ARCHULETA (1985).

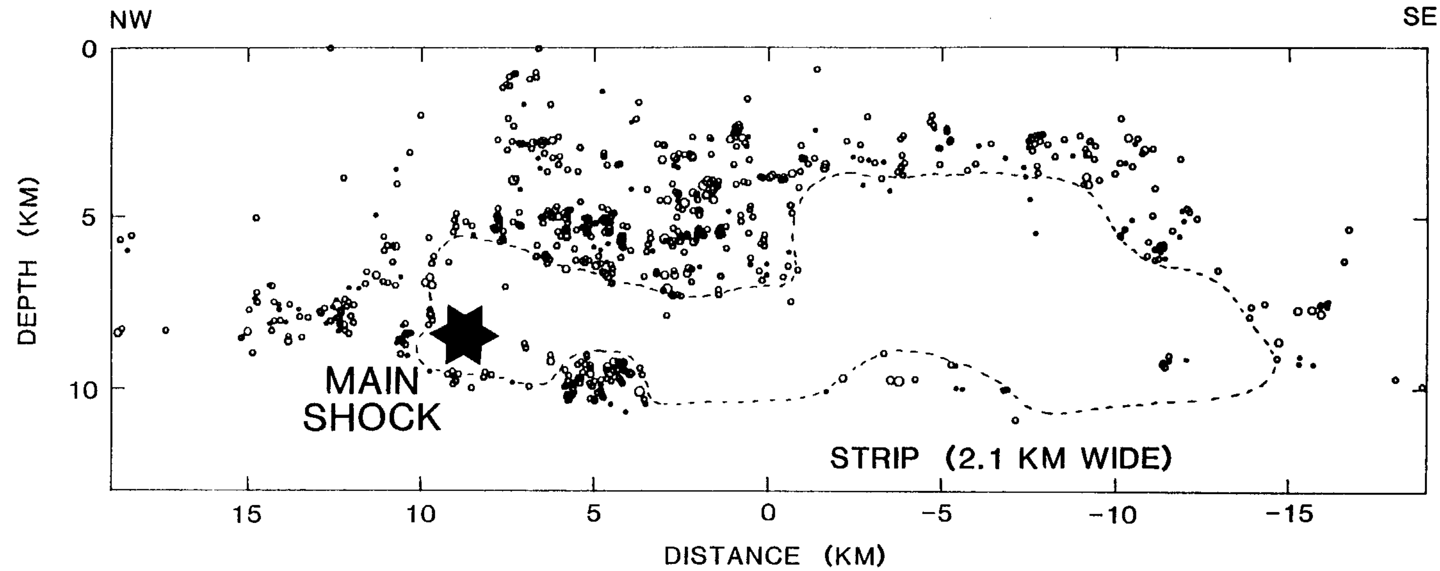

Figure 5
Aftershocks of 1984 Morgan Hill, California, earthquake for a strip 2.1 km wide aligned along the dip and strike of the main Calaveras fault. The section extends from Hall's valley to Coyote Lake. The region largely devoid of aftershocks, is apparently associated with rupture in the main event. From COCKERHAM and EATON (1985).

no surface breaks. The amount of slip, however, that can be absorbed by the surface barrier is apparently limited to about one meter, and usually only events with more than one meter of slip at depth have surface breaks (see also VITA-FINZI and KING, 1985).

The effect of the relaxation barriers is to localize the maximum slip within the center of the brittle zone (see Fig. 4), leaving a deficit to be taken up above and below that region by interseismic processes, in the manner discussed earlier. In the brittle zone the deformation occurs by aftershocks (see Fig. 5), background seismicity, and foreshocks; outside that region the deformation is accommodated by creep either on the fault or in the surrounding rock.

Whether the motion is ductile or brittle, the deformation away from the main fault accumulates over many earthquake cycles and produces dragfolds. These are the horizontal equivalent of the anticlinal structures demonstrated to result from repeating dip-slip earthquakes (KING and VITA-FINZI, 1981; KING and BREWER, 1983; STEIN and KING, 1984; VITA-FINZI and KING, 1985). These folds develop as a consequence of inhomogeneous fault motion as a function of depth, which is in turn dictated by the nature of the barriers that control the vertical distribution of earthquake slip. Drag folds (discussed in the next section) also form as a result of inhomogeneous motion along strike, in this case controlled by the distribution of barriers along strike.

Because geological markers are usually horizontal, drag folding associated with strike-slip faulting is less easy to examine than folding associated with dip-slip faulting. On seismic profiles, for example, the features that are observed are vertical folding and 'flower structures' (e.g., BALLY, 1983). These structures result from Poisson's ratio uplift and subsidence. While the vertical deformation is readily observed, both in profiles and in the morphology, it is a secondary manifestation of the ductile processes of strike-slip faulting. If, in general, rock bedding were vertical and not horizontal, the flower structure folding would pass largely unnoticed, in the same way that vertical-axis folding presumably exists but is not observed in association with dip-slip faults.

The behavior of a dip-slip fault is more complex (Fig. 6). Near the surface the same conditions prevail as those described for strike-slip earthquakes. Since motion is vertical, however, the surface folding produced by repeated earthquakes is readily identified by its morphological and structural effect. As pointed out earlier, there is no difference in the mechanics of formation between folds formed above dip-slip faults (sometimes referred to as drape folds) and drag folds caused by similar barriers associated with strike-slip faulting; only their observability differs.

The 1980 El Asnam earthquake, besides providing an unequivocal example of earthquake-related folding, provided an important warning about the interpretation of surface faulting. The greatest total length of faulting caused by that compressional event was extensional and was a secondary consequence of anticline uplift (Fig. 7). Normal faulting could be traced even in regions where no thrust surface breaks

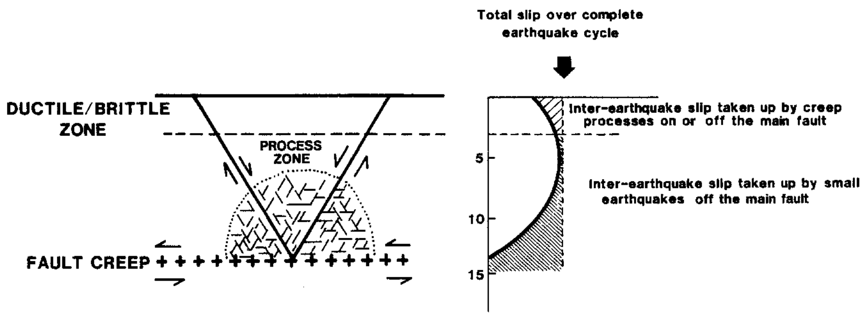

Figure 6
Processes as a function of depth for dip-slip faults. Near-surface creep results in relaxation barriers in a manner similar to the process in strike-slip faults (Fig. 3). At depth the effect of creep can produce both relaxation-barrier and fragmentation-barrier processes, the latter because of the presence of detachment surfaces.

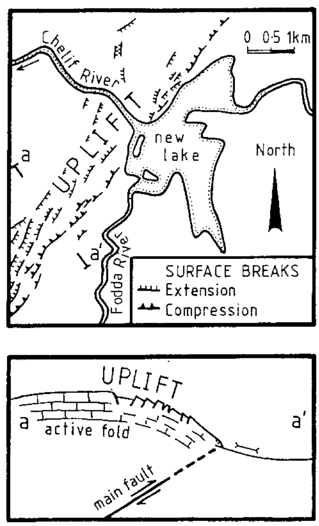

Figure 7
Faulting and anticline uplift associated with the 1980 El Asnam earthquake. Upper figure: surface ruptures and lake that appeared after earthquake. Lower figure corresponding to the section (a–a′ on upper figure): form of anticline and associated faulting. Extensional faulting at surface is secondary and cannot extend to depths greater than 1 or 2 km. The reverse fault associated with main event is shown. Contraction, indicated by telescoped irrigation channels, occurred in main fault footwall.

occurred. Similar, but less spectacular, faulting was associated with the Corinth 1981 earthquake (VITA-FINZI and KING, 1985). The important feature of secondary faulting is that it is due to surface strains, which result when the main faulting at depth does not reach the surface. Hence, such faulting must die out with depth and

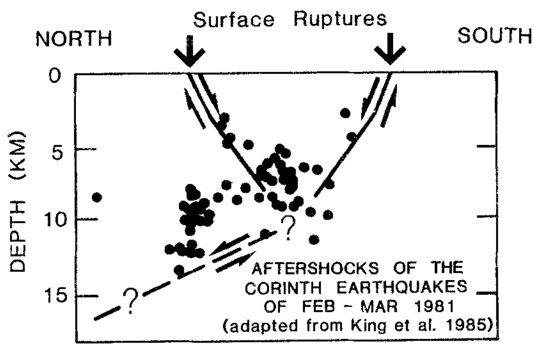

Figure 8
Aftershocks of 1981 Corinth earthquakes.

is *not* a direct surface manifestation of the seismic faulting at depth. This distinction between *primary* and *secondary* surface faulting is important in understanding the nature of surface features in active regions.

At depth, in dip-slip regimes, less information is available about the processes that occur. Nevertheless it is useful to speculate from the structures believed to be seen in reflection records and interpretations of seismograms and aftershock distributions.

If normal faults flatten into 'detachment' surfaces, the bend must produce fragmentation barriers. The depth at which the flattening occurs, however, is apparently also related to the depth at which creep processes occur (SIBSON, 1982; SIBSON, 1983; SMITH and BRUHN, 1984). Thus, the same depth range may be expected to be associated with relaxation barriers. Deformation thought to be associated with fault bends or fragmentation barriers can be seen in the aftershock sequences of some earthquakes; an example is shown in Figure 8.

Changes of fault behavior along strike

Figure 9 summarizes the information about the propagation of rupture along fault strike for some recent major earthquakes. While interpretations differ, the hypothesis suggested by a number of authors (SYKES and SEEBER, 198?; BAKUN and LINDH, 1985), that bends and hence, in the terminology of this paper, fragmentation barriers, play an important role in earthquake initiation and termination, is supported by a growing number of well-documented examples. Figure 10 illustrates, for two dimensions (biaxial deformation), the processes at a fragmentation barrier that create the conditions causing earthquakes to initiate and terminate in these places. The process will actually be three-dimensional.

Only one fragmentation barrier is considered in Figure 10, and no allowance is

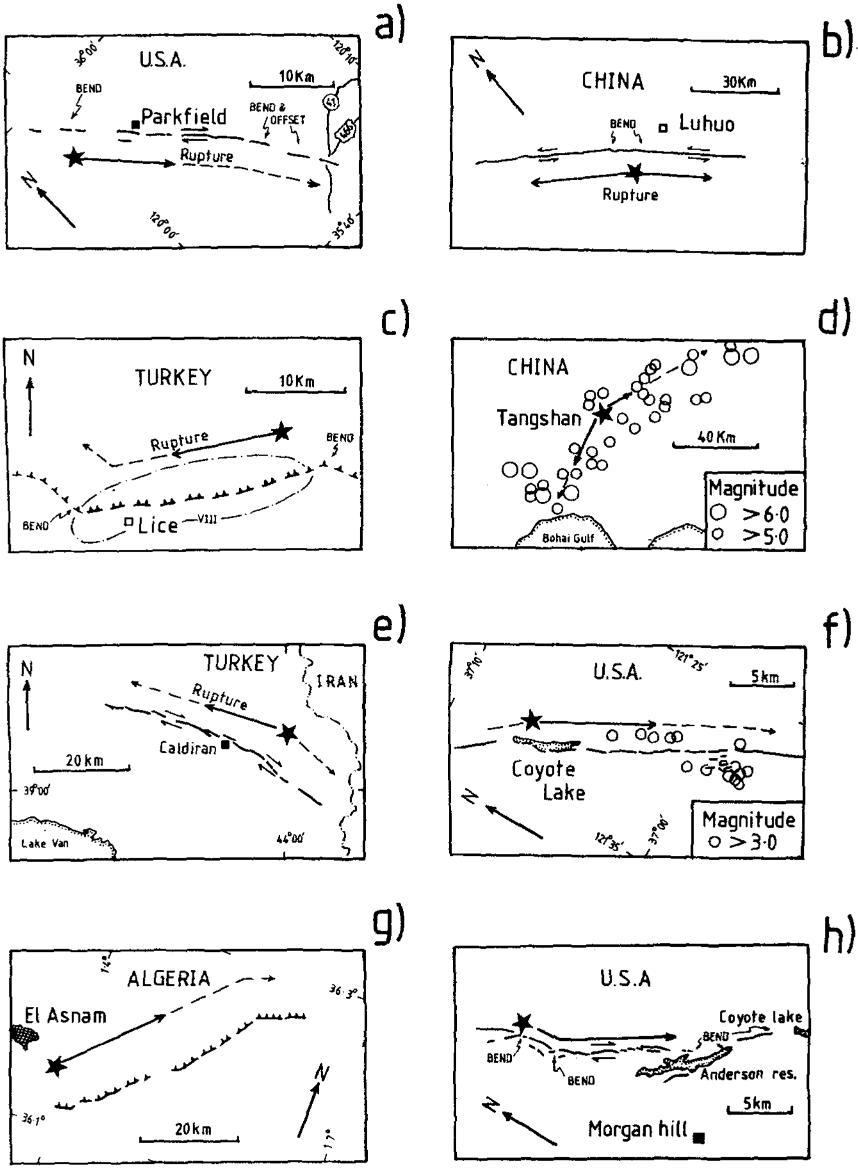

Figure 9a–h
Simplified fault geometry, in regions of rupture initiation and termination, for some well-studied recent earthquakes. Adapted from KING and NABELEK (1985); see same for discussion of data on which these figures are based.

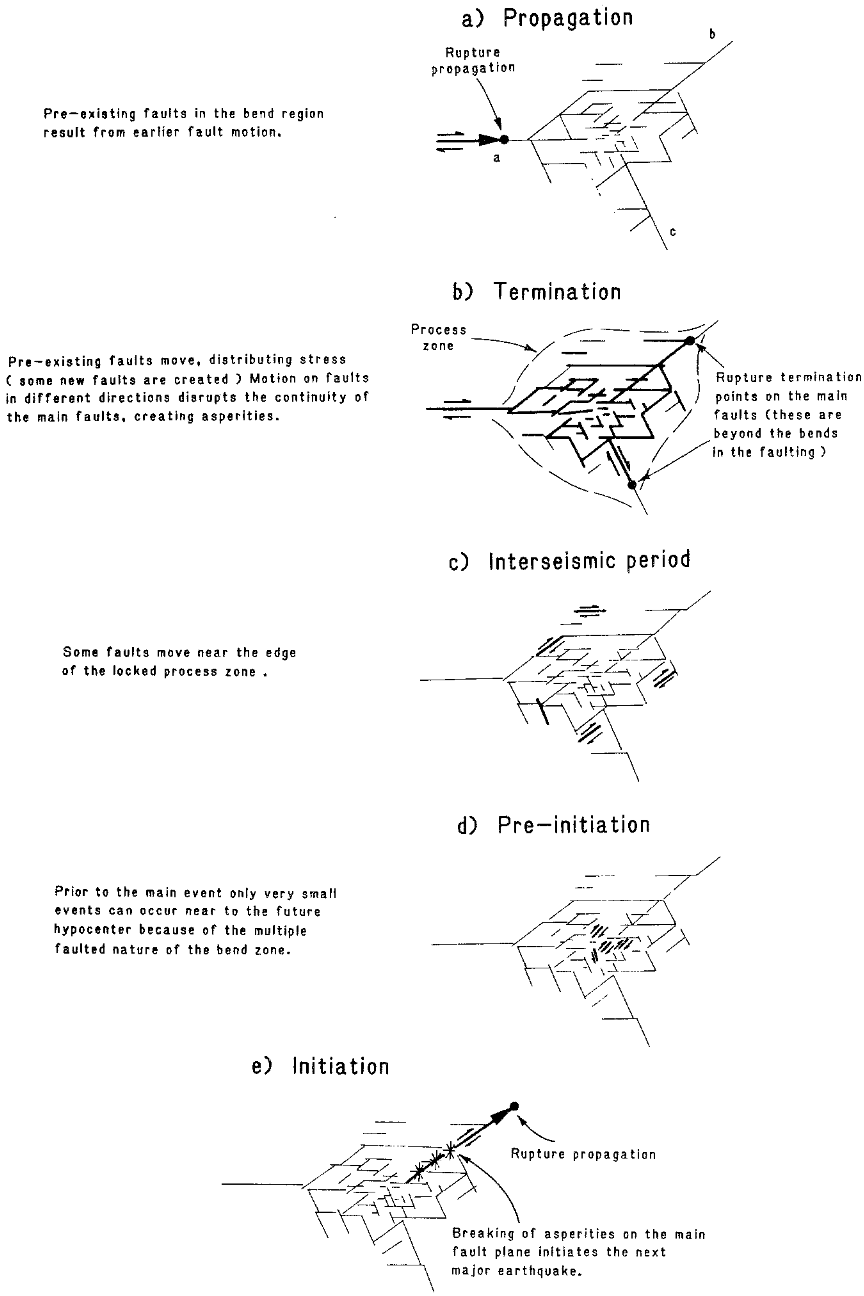

Figure 10a–e

Geometric processes occurring at a fault bend. The figures indicate the way in which a fault bend acts as a barrier to terminate earthquake rupture and at the same time sets up the conditions for future earthquake initiation. Modified from KING and NABELEK (1985). For convenience, figure shows initiation of a second earthquake in termination region of previous event; it should be interpreted simply as showing how termination processes 'set up' initiation conditions for some future event, not necessarily the next event.

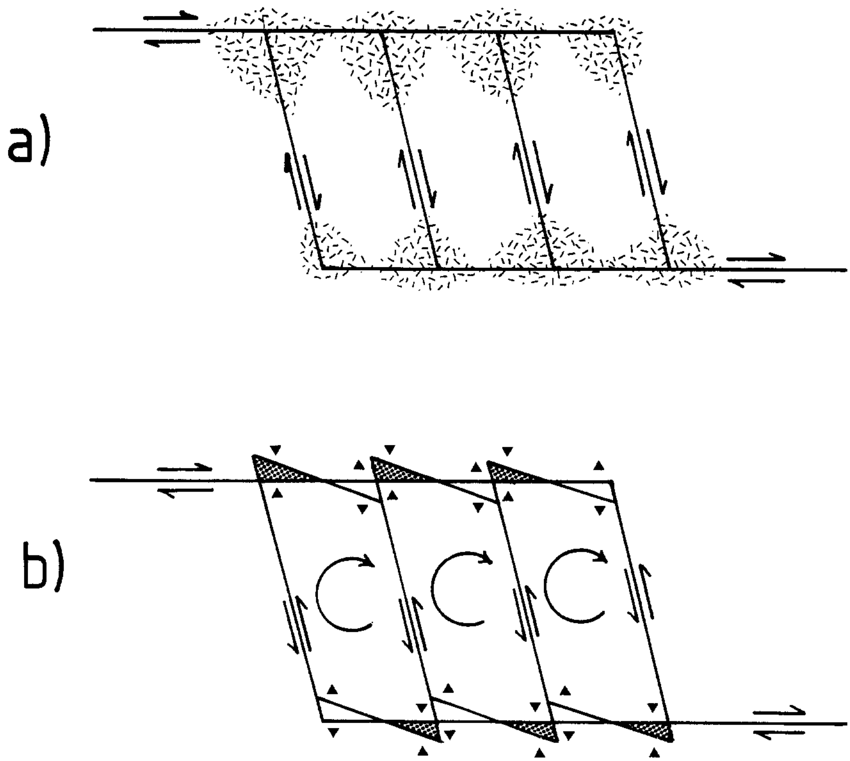

Figure 11
Fault offset mechanisms requiring multiple faulting: (a) nonrotational mechanism discussed by KING (1983); (b) rotational mechanism proposed by SEEBER and NICHOLSON (1985).

made for the way tectonic loading is applied or for the interaction between adjacent fault segments and associated barriers.

The model has two important features:

1. Rupture must extend past the bend to bring the 'barrier' into existence. The rupture is stopped by the zone of fractured rock that distributes stress.
2. Motion on faults of different orientations causes interlocking. The termination process of the barrier creates the *asperites* that must break to initiate a future earthquake.

Earlier it was pointed out that fault offsets or jogs may be regarded as composed of sets of bends and, hence, sets of fragmentation barriers at depth. This is true whether the motion transfers between fault segments by the processes shown in Figure 11a or by those shown in Figure 11b. Both mechanisms cause the main fault displacement to diminish as the offset zone is approached, and both require multiple small-displacement faulting to exist at depth. The figures shown only motion in the horizontal plane. Actually offset zones are associated with dip-slip motion which cause uplift or subsidence. Fragmentation processes will therefore be three-dimensional rather than two-dimensional.

Whatever the detailed geometry, since short faults have small displacements rupture on the faults at depth is limited to small incremental movements. Thus, these faults will not normally traverse the surface relaxation barrier and form surface breaks during earthquakes. If the surface barrier results from volume creep, the cumulative deformation will form folds and no primary faulting will be observed.

Since large surface strains can cause secondary faulting, this does not mean that surface faulting will not occur in barrier regions. However, the relation between surface features and seismogenic structures at depth must be treated with caution.

There is little or no evidence that earthquakes are limited along strike by relaxation barriers, although in principle such barriers should occur. This, in part, may be because creeping regions cannot readily be identified. The only well-documented segment of a fault that creeps in a depth zone that is normally seismogenic is a 200 km segment of the San Andreas between Parkfield and San Juan Batista. Attempts to discover fault creep on a similar scale elsewhere have proved unsuccessful, although geodetic results suggest that a system of faults in New Zealand moves by creep (WALCOTT, 1984).

The only place that argues for a relaxation barrier that limits the horizontal extent of earthquake faulting is the middle mountain region where Parkfield earthquakes start, but since the initiation region is also associated with a fault bend and has not been extensively studied with instruments until recently, the roles of fragmentation and relaxation processes cannot yet be assessed.

Tectonic loading and the location of rupture initiation

A feature of the barriers described, with the exception of dilatation barriers, is that fault slip in earthquakes tapers in amplitude as the barrier is approached. Thus, following an earthquake, a slip deficit is left in these regions. Some of this deformation lag can be accommodated by aftershocks, background seismicity, foreshocks, and creep, but the overall effect of motion on a fault system consisting of a series of segments separated by barriers is that the barrier brittle zones will be subject to a greater load than will other parts of the fault. This is illustrated in Figure 12. If the events shown have identical displacement-to-length ratios, then the middle-sized segment must move twice for each movement of the largest, and the smallest must move four times. Of course, segments can trigger each other to form multiple events. The shaded regions of Figure 12 indicate the deformation that must occur off the main faults between events.

Certain features are apparent from the representation. Some slip deficit will remain at all but the centres of fault segments. Hence, if we also allow for tapering with depth, off-fault seismicity may be expected almost everywhere, but with the

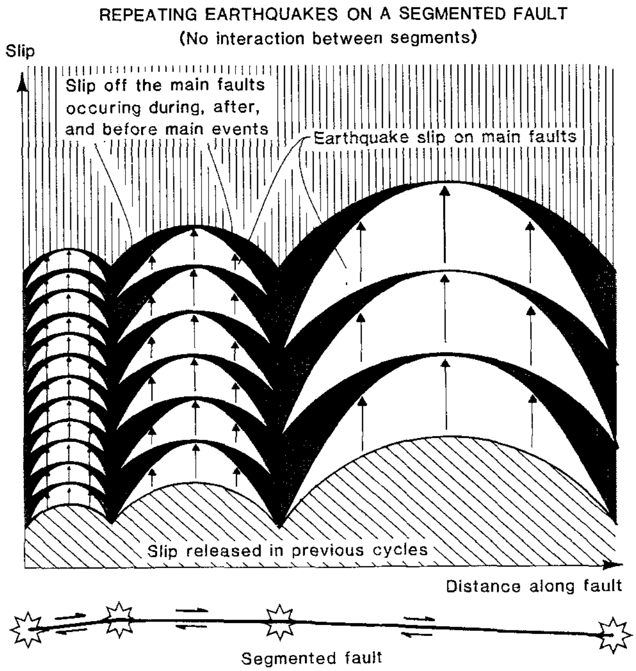

Figure 12

Plot of slip on a segmented fault over several earthquake cycles for fragmentation barriers. The three segments differ in length by factors of 2. For every slip event on the right-hand segment the central segment must rupture twice and the left-hand segment four times. Slip in the barrier regions between segments lags behind slip at the present centres.

most moment release near the barrier regions. A consequence of the slip deficit near the barriers is that it is in these regions that earthquake initiation is to be expected.

Since rupture initiation close to a barrier is a consequence of the tapering of the slip function of the previous earthquake, relaxation, as well as fragmentation barriers, may be expected to be associated with earthquake initiation. Thus, because of the form of slip as a function of depth (Figs. 3 and 6), rupture may be expected to start either at depth or near the surface. Considering that larger events cut the surface (and the associated dilatation barrier is not expected to be associated with rupture initiation), most substantial earthquakes should begin near the base of the brittle zone. This is commonly observed (e.g., SIBSON, 1983). It is worth noting that this geometrical reason for earthquake initiation at the boundaries of the brittle zone may be regarded as an additional, rather than an alternative explanation, to that provided by DAS and SCHOLZ (1984). They consider the problem from the perspective of rock strength.

Conclusions

The three types of barrier that result from the geometric features of faulting or from variations of rock properties have been simply classified. An understanding of the nature of these barriers provides a framework within which to understand many features of fault behavior. Repeating earthquakes occur between pairs of barriers (or along segments separated by several barriers, in multiple-event earthquakes). Except in the case of dilatational barriers, slip amplitude in earthquakes tapers towards the barriers. Consequently, a deficit of slip remains at the ends of a fault after a main event, and this slip is accommodated by creep process or by on-fault or off-fault aftershocks, according to the nature of the barrier. The slip deficit in the barrier regions also means that earthquakes will begin in these regions rather than alsewhere. This explains not only why earthquakes start near fault bends but also why they commonly start near the base of the brittle zone.

Off-fault deformation, whether occurring by brittle processes at fragmentation barriers or by ductile processes at relaxation barriers, in the long term, must create geological features in the form of drag folds and drape folds. Since these record the summed deformation of a series of seismic and interseismic periods, a study of them may be expected to reveal information about barrier processes.

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References

Aki, K. (1979), *Characterization of barriers on an earthquake fault.* J. Geophys. Res. *84*, B11, 6130–6147.

Archuleta, R. J. (1984), *A faulting model for the 1979 Imperial Valley earthquake.* J. Geophys. Res. *89*, 4559–4585.

Bakun, W. H. and Lindh, A. (1985), *The Parkfield, California, prediction experiment.* Science *229*, 619–624.

Bakun, W. H. and McEvilly, T. V. (1979), *Earthquakes near Parkfield, California: Comparing the 1934 and 1966 sequences.* Science *205*, 1375–1377.

Bakun, W. H. and McEvilly, T. V. (1984), *Recurrence models and Parkfield, California, earthquakes.* J. Geophys. Res. *89*, B5, 3051–3058.

Bally, A. W., *Seismic expression of structural styles.* A.A.P.G. Studies in Geology Ser. No. 15, vol 3, 1983.

Cisternas, A., Dorel, J. and Gaulon, R. (1982), *Models of the complex source of the El Asnam earthquake.* Bull. Seism. Soc. Am. *72*, 2245–2266.

Cockerham, R. S. and Eaton, J. P., 'The April 24, 1984, Morgan Hill earthquake and its aftershocks: April 24 through September 30, 1984,' (eds., J. Bennett and R. Sherburne), *The 1984 Morgan Hill, California, earthquake.* Calif. Dept. Conserv., Div Mines Geol., Spec. Publ. *68*, 1985.

DAS, S. and AKI, K. (1977), *Fault plane with barriers: A versatile earthquake model.* J. Geophys. Res. *82*, 5658–5669.

DAS, S. and SCHOLZ, C. (1983), *Why large earthquakes do not nucleate at shallow depths.* Nature *305*, 621–623.

EYIDOGEN, H. and JACKSON, J. (1985), *A seismological study of normal faulting in the Demirci, Alasehir and Gediz earthquakes of 1967–70 in Western Turkey: Implications for the nature and geometry of deformation in the continental crust.* Geophys. J. Roy. Astr. Soc. *81*, 569.

HUSSEINI, M. I., JOVANOVICH, D. B., RANDALL, M. J. and FRUEND, L. B. (1975), *The fracture energy of earthquakes.* Geophys. J. Roy. Astr. Soc. *43*, 367–385.

KANAMORI, H., 'Use of seismic radiation to infer source parameters,' Proc. Conf. III, *Fault mechanics and its relation to earthquake prediction.* U.S. Geol. Surv. Open-file Rept. 78–380, 1978.

KING, G. C. P. (1983), *The accommodation of large strains in the upper lithosphere of the Earth and other solids by self-similar fault systems: The geometrical origin of b-value*, Pure Appl. Geophys. *121*, 761–1093.

KING, G. C. P. and BREWER, J. (1983), *Fault related folding near the Wind River thrust, Wyoming, U.S.A.* Nature *306*, 147–150.

KING, G. C. P. and NABELEK, J. (1985), *The role of bends in faults in the initiation and termination of earthquake rupture.* Science *228*, 984–987.

KING, G. C. P., OUYANG, Z. X., PAPADIMITRIOU, P., HOUSEMAN, G., JACKSON, J. A. VIRIEUX, J., SOUFLERIS, C. and DESCHAMPS, A. (1985), *The evolution of the Gulf of Corinth (Greece): An aftershock study of the 1981 earthquakes.* Geophys. J. Roy. Astr. Soc. *80*, 677–693.

KING, G. C. P. and STEIN, R., 'Surface folding, river terrace deformation rate and earthquake repeat time in a reverse faulting environment: The Coalinga, California, earthquake of May 1983,' in *The 1983 Coalinga California earthquake.* Calif. Div. Mines Geol. Spec. Publ. 66, Sacramento, Calif., 1983.

KING, G. C. P. and VITA-FINZI, C. (1981), *Active folding in the Algerian earthquake of 10 October 1980.* Nature *292*, 22–26.

KING, G. C. P. and YIELDING, G. (1983), *The evolution of a thrust fault system: Processes of rupture initiation, propagation and termination in the 1980 El Asnam (Algeria) earthquake.* Geophys. J. Roy. Astr. Soc. *77*, 915–933.

LINDH, A. G. and BOORE, D. M. (1981), *Control of rupture by fault geometry during the 1966 Parkfield earthquake*, Bull. Seism. Soc. Am. *77* (1), 95–116.

RICE, J. R., 'The mechanics of earthquake rupture,' in (eds. A. M. Dziewonski and E. Boschi) *Physics of the Earth's Interior.* Proc. Int. School Physics, 'Enrico Fermi,' Course 78, 1979, Italian Phys. Soc., North Holland Publ., p. 555–649, 1980.

SCHOLZ, C. H. (1982), *Scaling laws for large earthquakes: Consequences for physical models.* Bull. Seism. Soc. Am. *72* (1), 1–14.

SEEBER, L. and NICHOLSON, C., *Slip along discontinuous wrench faults and the role of block rotation in seismic and interseismic deformation.* Lamont-Doherty Geol. Observ., Palisades, N.Y. (preprint), 1985.

SIBSON, R. H. (1982), *Fault zone models, heat flow, and the depth of distribution of earthquakes in the continental crust of the United States.* Bull. Seismol. Soc. Am. *72* (1), 151–163.

SIBSON, R. H. (1983), *Continental fault structure and the shallow earthquake source.* J. Geol. Soc. London *140*, 741–767.

SIBSON, R. H. (1985), *Stopping of earthquake ruptures at dilatational fault jogs.* Nature 316.

SMITH, R. B. and BRUNHN, R. L. (1984), *Intraplate extensional tectonics of the eastern basin-range: Inferences on structural style from seismic reflection data, regional tectonics and thermal-ductile deformation.* J. Geophys. Res. *89* (B7), 5733–5762.

STEIN, R. and KING, G. C. P. (1984), *Seismic potential revealed by surface folding: The 1983 Coalinga, California, earthquake.* Science *224*, 869–872.

STEIN, R. S. 'Evidence for surface folding and subsurface fault slip from geodetic elevation changes associated with the 1983 Coalinga, California, earthquake,' *Proc. Workshop XXVII.* U.S. Geol. Surv. Open-file Rept. p. 85–44, 1985.

SYKES, L. and SEEBER, L. (198?), *Great earthquakes and great asperites, San Andreas Fault, Southern California.* Pure Appl. Geophys. (not available at press time).

VITA-FINZI, C., and KING, G. C. P. (1985), *The seismicity, geomorphology and structural evolution of the Corinth area of Greece. Phil. Trans. Roy. Soc. London, A314*, 379–407.

WALCOTT, R. I. (1984), *The kinematics of the plate boundary zone through New Zealand: A comparison of short- and long-term deformations.* Geophys. J. Roy. Astr. Soc. *79*, 613–633.

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