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Geometric pattern, rupture termination and fault segmentation of the Dixie Valley–Pleasant Valley active normal fault system, Nevada, U.S.A.

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Abstract—Active fault systems of both interplate and intraplate settings clearly show patterns of segmentation. Fault segmentation, therefore, is very important for understanding fault behavior and assessing seismic hazard. The complexity of geometric patterns along fault systems has a strong influence on the propagation of earthquake ruptures, because it can form structural heterogeneities that tend to inhibit rupture propagation. This implies that there should be relations between fault segmentation and the complexity of fault geometry. The active normal faults of the Dixie Valley–Pleasant Valley system provide an excellent example to study such relations and to establish the geometric method of normal fault segmentation in the Basin and Range province. Studies of fault segmentation in the Dixie Valley–Pleasant Valley fault system indicate that there are changes of both geometric pattern and geomorphic character in the area of the segment boundary. The characteristics of segment boundaries on normal faults of the Dixie Valley–Pleasant Valley system can be explained in terms of the nature of rupture propagation and termination. These changes may be helpful criteria for distinguishing normal fault segmentation at least in the Basin and Range province.

INTRODUCTION

ONE OF the interests of structural geologists is how faults behave. Seismotectonic studies during the past two decades reveal that most large faults do not rupture along their entire length during individual rupture events, but instead consist of segments that may rupture independently of each other, each with its own rupture history (Allen 1967, Swan *et al.* 1980, Schwartz & Coppersmith 1984, Wheeler 1987, 1989). Seismic belts of both interplate (e.g. Gulf of Alaska, Sykes 1971) and intraplate (e.g. central Nevada–eastern California, Wallace & Whitney 1984) settings clearly manifest segmented patterns, and future earthquakes are expected to occur in 'seismic gaps'—historically unruptured fault segments.

Fault segmentation, therefore, is very important for understanding fault behavior and assessing seismic hazard. The identification of segments and their boundaries, however, is not easy, and the method is still in an early stage of development (Schwartz 1988). At present, the most reliable method of determining fault segmentation is based on paleoseismic and fault behavior data that require dating of prehistorical earthquake events. This requirement is very difficult to satisfy for many faults, especially in the Basin and Range province, where recurrence intervals between earthquakes are typically several thousand years (Wallace 1984a) and datable material is difficult to obtain. Thus, there is a need to develop other reliable methods of fault segmentation based solely on easily obtainable geologic, geomorphic and geometric data.

The key point in fault segmentation is to identify persistent segment boundaries, where most or all of a propagating rupture is arrested event after event. Minor tectonic slip or triggered secondary slip may rupture through the segment boundary, but the main rupture associated with an earthquake does not appear to cross the segment boundary. Detailed studies suggest that the observed seismic behavior of a fault can be correlated with mapped complexities of fault geometry, such as change in strike, bifurcations and stepovers, and that mapped complexity on the surface may persist through the thickness of the seismogenic zone (Eaton *et al.* 1970, Bakun *et al.* 1980, Reasenberg & Ellsworth 1982, Okubo & Aki 1987). Thus the earthquake faulting process, from nucleation to termination, may be heavily controlled or influenced by the complexity of fault geometry (Segall & Pollard 1980, Okubo & Aki 1987). Studies of historical surface ruptures and active faults (late Quaternary) in the Basin and Range province indicate that the rupture termination and segment boundaries are often associated with complex fault geometry (Slemmons 1957, Wallace 1984b, Bruhn *et al.* 1987, Fonseca 1988, Crone & Haller 1989, Machette *et al.* 1989). Seismological studies of the nucleation and termination of earthquakes also show that the majority of earthquake ruptures are terminated in areas with complex fault geometry (Aki 1989), such as in the 1983 Borah Peak earthquake (Crone *et al.* 1987). There appears to be a connection between segment boundaries and the complex geometry of faults, and therefore it may be possible to develop a method of fault segmentation based only on fault rupture geometry and fault zone geomorphology.

Studies of historical ruptures world-wide, including the Basin and Range province, show that there are no typical geometric patterns that characterize the termination of a rupture zone (dePolo *et al.* 1989, Knuepfer 1989). The reliability of the available data set, however,

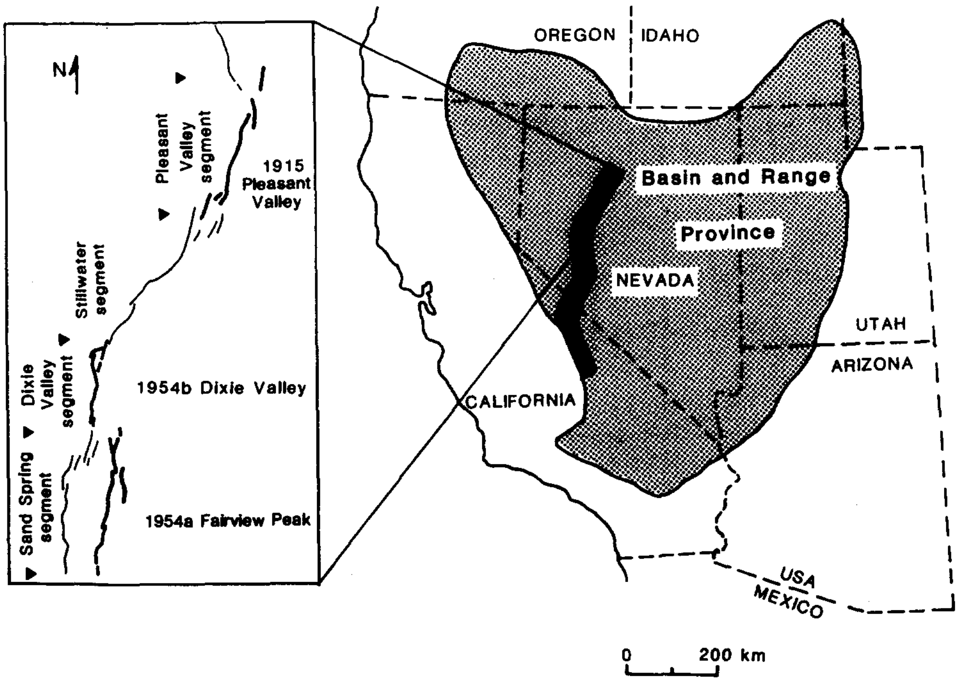

Fig. 1. Location of the Dixie Valley–Pleasant Valley active normal fault system in the Basin and Range province. The dotted area is the Great Basin. The black strip shows the central Nevada and eastern California Seismic Belt. The insert shows the Dixie Valley–Pleasant Valley active normal fault system. The thick lines are historical ruptures and the thin lines are late Quaternary fault scarps. The proposed fault segments of the system are also shown.

may be influenced by many factors such as preservation of historical ruptures and the detail of the original study. Another aspect not included in the comparative studies of dePolo *et al.* (1989) and Knuepfer (1989) is the relation between historically ruptured and unruptured fault scarps in the area of rupture termination. The interaction between two major fault segments may be important in controlling earthquake rupture termination (Segall & Pollard 1980, Sibson 1987). Thus, the ideal way to approach the establishment of geometric and geomorphic methods of fault segmentation would be to: (1) select a fault system with both historically ruptured and unruptured segments; (2) study the historical rupture and its termination; (3) study the relation between the historically ruptured and unruptured fault segments; (4) establish criteria to identify the segmentation boundaries; and (5) apply the observed relations to other fault systems to verify the validity of the method.

The active normal faults of the Dixie Valley–Pleasant Valley system, west-central Nevada, may be one of the best places in the entire Basin and Range province to study the segmentation of normal fault systems, as several historical earthquakes have occurred along this fault system (1915 Pleasant Valley, 1954 Fairview Peak and 1954 Dixie Valley earthquakes) in this century, and each produced large surface rupture zones that are still well preserved (Slemmons 1957, Wallace 1984b). These historical rupture zones are separated by Holocene fault scarps that have not ruptured during historical time (Bell 1984, Bell *et al.* 1984, Wallace & Whitney 1984). This fault system has shown its natural segmentation pattern by the historical earthquakes and ruptures and, thus, provides an ideal opportunity to study the geometric and geomorphic pattern of each segment and each segment boundary, and to establish geometric and geomorphic criteria of fault segmentation for normal fault systems in the Basin and Range province.

GEOLOGICAL SETTING OF THE DIXIE VALLEY–PLEASANT VALLEY FAULT SYSTEM

The Dixie Valley–Pleasant Valley fault system is in the northwestern part of the Basin and Range province (Fig. 1), and forms the northern part of the central Nevada and California Seismic Belt (Wallace & Whitney 1984). The fault system consists of fault scarps along the western flank of the Sand Spring Range, the western flank of the Stillwater Range, the eastern flank of the Tobin Range and the eastern flank of the Sonoma Range (Fig. 2). The Stillwater Range, west of the fault system, consists of Mesozoic pelites, gabbros and granodiorites overlain by Tertiary volcaniclastic rocks and basalts (Page 1965, Okaya & Thompson 1985). Seismic-reflection profiles and gravity data across Dixie Valley show that the valley is: (1) an asymmetrical graben filled with about 1.8 km of Quaternary lacustrine and alluvial fan deposits; (2) bounded by a major E-dipping high-angle (~60°) normal fault on the western side; and (3) has three minor, W-dipping, normal faults on the eastern side (Okaya & Thompson 1985). The rupture associated with the 1915 Pleasant Valley earthquake and the Sonoma range-front fault bound the Pleasant Valley on the west (Fig. 2), which is filled with Quaternary sedi-

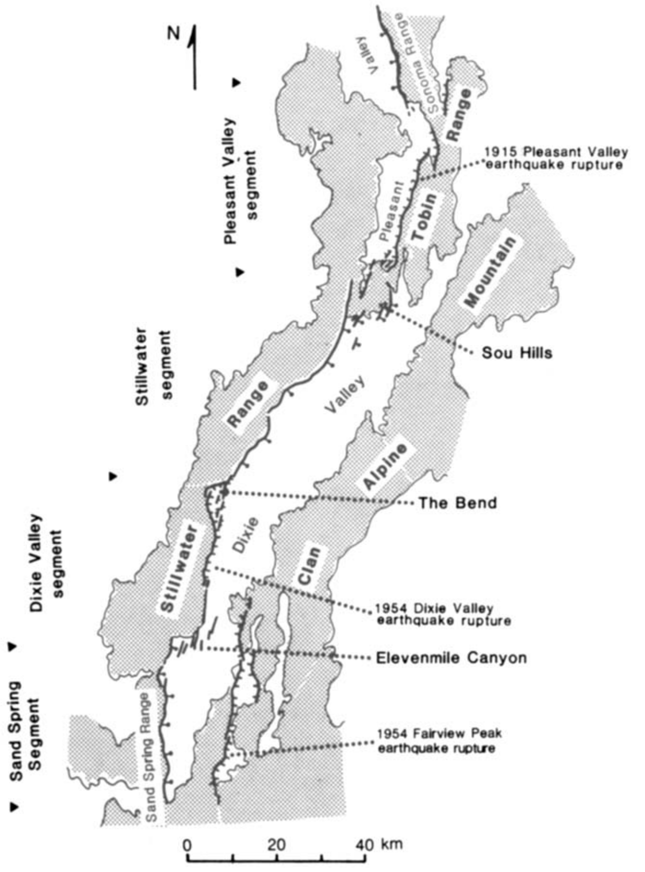

Fig. 2. Major faults and proposed fault segments in the Dixie Valley–Pleasant Valley active normal fault system. The thick lines are the major faults. The balls are on the downthrown side of prehistorical normal faults. The thick lines with ticks are the historical earthquake rupture zones, and the ticks are on the downthrown side of the historical ruptures.

ments that are deepest in the northern part according to Bouguer gravity surveys.

Both Holocene and historical faults are abundant along the western side of Dixie Valley (Bell 1984). Most of these faults are present along the range-front and separate Quaternary basin sediment from bedrock. Piedmont faults are well displayed in places such as The Bend and Elevenmile Canyon. Discontinuous Quaternary faults are also present along the central and eastern side of the valley (Fig. 2) (Whitney 1980, Bell 1984), however, the scarps along these faults are less fresh and more subdued than the fault scarps to the west. They are tectonically less important than the range-front faults along the western side of the valley because they do not control its morphology and subsidence. The valley floor is tilted to the west and the lowest portions of the valley are along its western boundary. Correspondingly, the alluvial fans on the eastern side are larger than those on the western side. These geomorphic features suggest that faults on the western side of the valley are more active (i.e. higher slip rates) and that deformation on this side has controlled late Cenozoic evolution of the valley.

The late Cenozoic deformation of this fault system appears to be dominated by extension and normal faulting. Bell & Katzer (1987, 1990) calculated the Holocene slip rate in the Bend area to be about 0.5 mm a^{-1}. The long-term slip rate is estimated to be 0.2 mm a^{-1} over the past 200,000 years (Bell & Katzer 1990). This rate is in good agreement with that estimated by Wallace & Whitney (1984) and by Okaya & Thompson (1985). In this paper we only discuss the geometric patterns of the major faults of the fault system.

THE 1954 EARTHQUAKE RUPTURE

On 16 December 1954 an earthquake of M_s7.1 occurred east of Fairview Peak, and was followed 4 min later by a shock of M_s6.8 (Slemmons 1957, Doser 1986). The first earthquake, the Fairview Peak earthquake, was associated with a 44 km (end-to-end) long complex surface rupture zone. The displacement is oblique-slip with predominantly right-lateral strike-slip component (Slemmons 1957, Zhang unpublished data). The ruptures are present along eastern flank of the Fairview Peak east of Dixie Valley (Fig. 2). The epicenter of the second earthquake is not well constrained, but is believed to be east of the main surface rupture zone, near the eastern side of Dixie Valley (Doser 1986). Surface ruptures associated with the second event, the Dixie Valley earthquake, flank the eastern side of the Stillwater Range on the western side of Dixie Valley (Figs. 2 and 3). Details of the geometric pattern of this rupture zone can be found in Slemmons (1957), Bell & Katzer

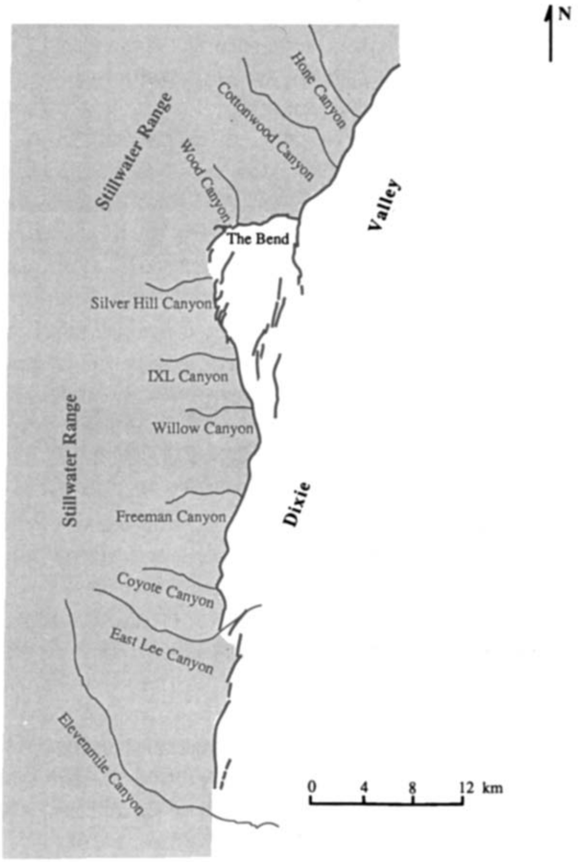

Fig. 3. Surface ruptures associated with the 1954 Dixie Valley earthquake. Dotted area is the Stillwater Range.

(1987). In the following we briefly summarize the general geometric pattern of the rupture zone associated with the Dixie Valley earthquake.

The southern end of the Dixie Valley rupture zone observed during field study is about 2 km north of Elevenmile Canyon (Fig. 3). The 1954 fault scarps associated with the 1954 earthquake in this area are less than 0.5 m in height. A continuous low fault scarp associated with the 1954 earthquake can be traced about 18 km northward, and it also follows a prehistorical fault scarp along the range front. The slope of scarp associated with the 1954 earthquake is usually degraded to between 30° and 35° and that of the prehistorical scarp is between 15° and 20°. Together they form a composite scarp that is over-steepened at its base. The height of the prehistorical fault scarp is usually between 3 and 6 m.

From Coyote Canyon northward to Willow Canyon (Fig. 3), the 1954 earthquake rupture is characterized by a major range-front fault scarp zone that consists of E-facing scarps and grabens along its base. The fault scarps are 2–3 m in height. The width of the graben varies from several meters to about 100 m, and is generally about 1–1.5 m in depth. The trend of the rupture zone in this section is about N15°–20°E. The rupture zone follows the boundary between bedrock and alluvium and shows a zig-zag pattern (Fig. 3). The morphology of the scarps along this section clearly shows no evidence of a Holocene event prior to the 1954 earthquake, an observation made earlier by Wallace & Whitney (1984) and Bell & Katzer (1987). Ruptures and associated deformation in this part of the fault zone appear to be concentrated only along the range front, as no surface ruptures have been found outside this range-front rupture zone.

Northward to The Bend area, major displacement associated with the earthquake continues along the range-front fault, however, the geometric pattern of the rupture zone becomes much more complex than the ruptures to the southern (Fig. 3). The strike of the range-front fault changes from about N15°E in the south to about N15°W. The range-front rupture commonly splays out into wide zones, and most of the branch faults are present on the eastern side of the main ruptures (Fig. 3). At The Bend, the trend of the range-front rupture becomes N80°E. In 1954, earthquake ruptures also occurred along the pre-existing piedmont fault system (Fig. 3) (Slemmons 1957, Bell & Katzer 1987, 1990, Zhang unpublished data). The Holocene fault scarp along the piedmont fault is about 2–3 m high, whereas the 1954 scarp is less than 0.5 m and is at the base of the Holocene scarp.

North of The Bend, the 1954 earthquake ruptured about 14 km along the Stillwater range-front fault (Fig. 3). The 1954 ruptures exactly follow the 4 to 6 m high Holocene fault scarp along the range-front. The style of deformation, however, is dramatically different. The displacement decreases significantly from 2 to 3 m south of The Bend to less than 0.5 m in most places to the north. Thus, the earthquake rupture appears to be significantly attenuated across The Bend and most of the earthquake energy seems to be released to the south. The main earthquake rupture appears to have been almost arrested at The Bend. Another possibility is that this section of the rupture may have triggered slip due to severe shaking during the 1954 earthquake. This northern section of the rupture is similar to that along the Lost River range-front fault during the 1983 Borah Peak, Idaho, earthquake, where about 8 km ruptures with small displacement north of the Willow Creek Hills are considered to have been triggered by the 1983 earthquake to the south of Willow Creek Hills (Crone *et al.* 1987, Crone & Haller 1989).

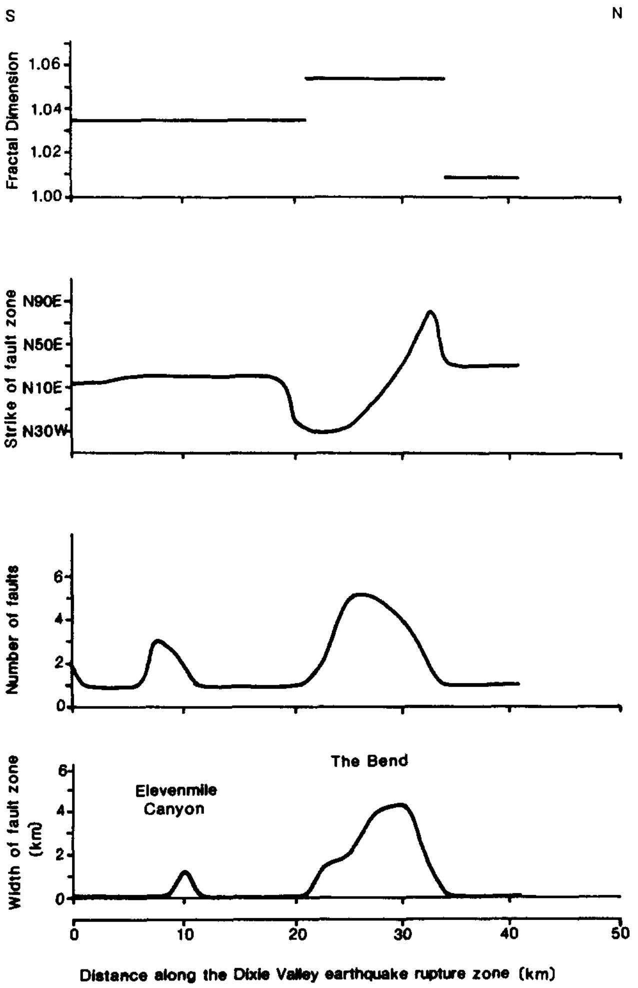

Fig. 4. Changes in character of the 1954 Dixie Valley earthquake rupture zone.

Figure 4 summarizes geometric changes along the 1954 Dixie Valley surface rupture zone. Significant geometric changes occur in The Bend area, where the Dixie Valley segment overlaps with the Stillwater segment (see discussion below) and where the 1954 Dixie Valley rupture terminated. We have measured fractal dimension, a measurement of geometric complexity (see discussion below), of the 1954 Dixie Valley earthquake rupture. The fractal dimension is large in The Bend area where the rupture is almost arrested, suggesting that complexity of fault geometry does influence earthquake rupture propagation. If ruptures within a fault system share the same kind of rupture process and faulting mechanics, the relation between rupture termination and geometric pattern of the 1954 Dixie Valley rupture should serve as an example for other fault segments within the system. In other words, similar geometric pattern or similar levels of geometric complexity should be observed in other segment boundaries of the fault system.

GEOMETRIC PATTERN OF SEGMENTS

The inherent segmentation of the Dixie Valley–Pleasant Valley active normal fault system is demonstrated by its historical earthquake rupture pattern. From south to north these segments are the Sand Spring Range, the Dixie Valley, the Stillwater Range, the Pleasant Valley and the Sonoma (Fig. 2). The above segmentation of the fault system is supported by paleoseismic studies of each individual segment (Fig. 5) (Wallace 1984b, Bell & Katzer 1990, Bell 1990 personal communication). Although the age of each paleoearthquake contains large uncertainty, it is clear that all proposed segments rupture independently of each other. The geometry of these segments is relatively simple, but they are separated by areas with complex geometry. Geometric and geomorphic anomalies similar to those at the northern end of the 1954 Dixie Valley ruptures can be found in almost all segment boundaries of the fault system.

The Sand Spring segment

The Sand Spring segment is the southernmost segment of the fault system. It is just south of the 1954 Dixie Valley surface rupture zone along the eastern flank of the Sand Spring Range, which is the southern extension of the Stillwater Range (Fig. 2). This 37 km long fault zone strikes approximately N–S. The fault scarp exactly follows the range-front, suggesting that it controls the uplift of the range to the west and the subsidence of the valley to the east. No evidence of lateral displacements have been found along this fault.

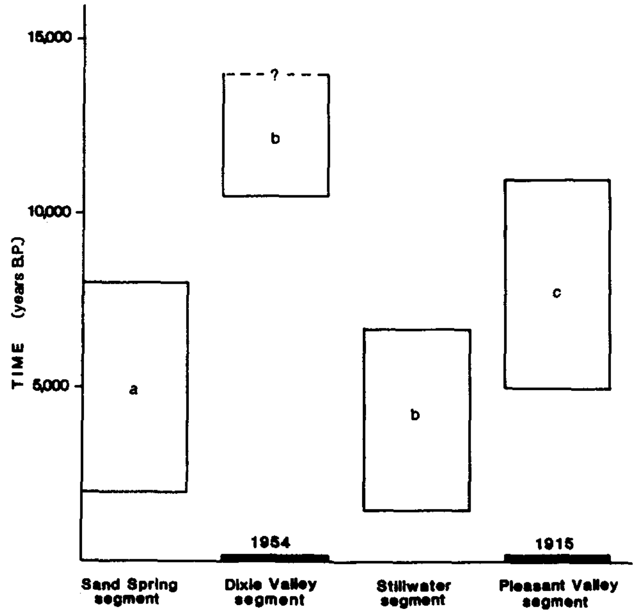

Fig. 5. Space–time diagram of paleoseismic events along the Dixie Valley–Pleasant Valley active fault system during the past 15,000 years. Rectangles show the time range when the paleoearthquake occurred. The question marks show that either the lower bound or upper bound of time of the paleoearthquake is not well constrained or unknown. Source of data: (a) Bell (1984, 1989 personal communication), (b) Bell & Katzer (1990) and (c) Wallace (1984b).

The fault zone geometry of the main part of this segment is relatively simple. Most of the fault scarps are concentrated on a narrow zone along the range-front. Very few branching fault scarps are present within 500 m of the range-front scarp, which usually has steep slopes of 20–30°. The height of the range-front fault scarp varies from 1 to 4 m. On the basis of scarp morphology and soil development, Bell (1984) estimated that the fault scarp was probably formed in mid to late Holocene time. Absence of bevels on the fault scarp along this segment suggests that the scarp was formed by a single faulting event.

North of this segment is a complex area that separates the Sand Spring Range segment from the Dixie Valley segment (Fig. 2), called the Elevenmile Canyon area. The Sand Spring range-front fault seems to splay out into this area. Numerous fault scarps are present within an 8-km-wide zone and the trends of both individual fault scarps and the zone itself changes from north to north-northeast. The length of individual fault scarps in this area is relatively short and they are subparallel to each other, dipping both east and west. In addition, the range-front area and the valley floor near Elevenmile Canyon is uplifted with respect to the valley floor to the north and south of it (Fig. 6a).

The Dixie Valley segment

The Dixie Valley segment is north of the Sand Spring segment (Fig. 2). As described above, the main part of the fault scarp associated with the 1954 earthquake follows the range-front fault of the Stillwater Range. Ruptures and significant displacements are concentrated within a narrow zone of the fault scarp and graben. Scarp morphology and paleoseismic studies (Wallace & Whitney 1984, Bell & Katzer 1990) suggest that there was no Holocene event prior to the 1954 faulting along the range-front fault zone.

The geometric pattern of the northern end of the Dixie Valley segment, The Bend area, is much more complex than the southern and central parts (Figs. 3 and 4). Surface ruptures and displacement associated with the 1954 earthquake are distributed along both the range-front and piedmont faults, as well as the area in between. We think the main rupture of the 1954 earthquake might have terminated or significantly attenuated in this area, only minor or secondary ruptures continue northward.

The piedmont fault in The Bend area is probably the southward continuation of the Stillwater range-front fault, another major segment of the system, and it becomes a piedmont fault in The Bend area only due to the range-front embayment (Figs. 2 and 6). During

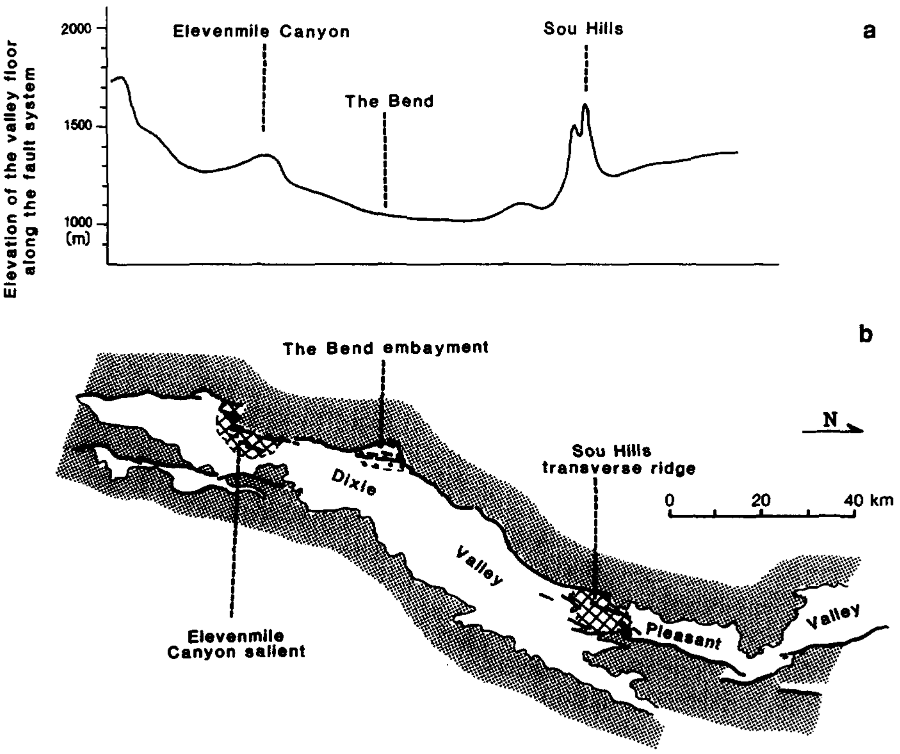

Fig. 6. Elevation of valley floor (a) and range-front geomorphology (b) along the Dixie Valley active normal fault system. The elevation profile (a) is along the middle of the valley. Note the coincidence between the geomorphologic anomalies of the valley floor and the range-front anomalies (b) in the proposed segment boundaries.

prehistoric rupture along the Stillwater range-front fault, the piedmont fault in The Bend was part of the major rupture of the prehistorical earthquake. To the south the piedmont fault dies out or merges with the range-front near Willow Canyon (Figs. 2 and 3). Thus, The Bend area is at an overlap between two major segments that each ruptured independently. During the 1954 earthquake, rupture on one segment of the range-front fault, might have triggered minor slip along the piedmont fault of the other segment. Even if the northern 14 km minor rupture was not triggered by the earthquake, the main rupture associated with the 1954 earthquake was significantly attenuated in The Bend area. Thus it probably has less tectonic meaning for the 1954 earthquake than the rupture south of The Bend. That these two faults (range-front and piedmont), representing two major segments, may have ruptured independently of each other is supported by detailed paleoseismological study in The Bend area by Bell & Katzer (1990). The paleo-scarp along the piedmont fault was probably formed between 1500 and 6800 years ago (Hecker 1985, Bell & Katzer 1990), whereas the event prior to the 1954 rupture along the range-front fault probably occurred in late Pleistocene time (Bell & Katzer 1990).

The Stillwater segment

A zone of prehistorical fault scarps approximately 40 km long is present along the eastern flank of the Stillwater Range (Fig. 2). Historical earthquake ruptures occur both north and south of this fault zone but not within the Stillwater segment. Wallace & Whitney (1984) term this "the Stillwater Seismic Gap", and postulated that it is the probable site of future faulting, perhaps in the next few decades. The Stillwater range-front fault scarps generally strike about N35°E and face southeast. Most of the fault scarps are concentrated in a narrow zone along the range-front and closely follow the sinuous range-front and hug the valley. Short (<1 km) branch faults and antithetic faults in the hangingwall are present in only three places.

No bevels have been found along the 4–6 m high range-front fault scarps, suggesting that they are the product of single-event offset. This is also supported by the fact that mid-Holocene soil offset is about the same as late Holocene soil offset (Hecker 1985, Fonseca 1988). The age of this Holocene event is constrained to be between 1500 and 6800 years ago (Hecker, 1985, Bell & Katzer 1987, Fonseca 1988).

Northward, the range-front fault splays out into the Sou Hills, an uplifted mid-valley bedrock transverse ridge, that has represented a profound barrier to the propagation surface rupturing throughout the Quaternary (Fonseca 1988). The structural style changes markedly in the Sou Hills. Fault scarps in the southern Sou Hills trend about 10–15° more to the east, whereas the length of individual scarps is <4 km, in contrast to >10 km-long scarps along the Stillwater range-front.

The Pleasant Valley segment

Unlike the segments in the south, the fault scarp within the Pleasant Valley segment is present mainly along the eastern side of the valley rather than the western side. The Pleasant Valley fault trends about N15°E along the western margin of the Tobin Range, and dips to the northwest. The 1915 Pleasant Valley earthquake occurred along this fault, and formed a 59 km long (end-to-end) surface rupture having a maximum vertical displacement of 5.8 m (Wallace 1984b).

The surface rupture zone of 1915 Pleasant Valley earthquake consists of four discontinuous fault scarps (Wallace 1984b, dePolo *et al.* 1989). A 1.5 km long ruptures in the Stillwater Range south of the Sou Hills are considered as landslide features by Wallace (1984b). Although some small-scale features, such as branching, discontinuities, abrupt changes in strike and gaps, exist in places, the overall geometric pattern of the rupture zone north of the Sou Hills area is relatively simple (Fig. 2) (plate 1 of Wallace 1984b). Most of the ruptures are concentrated within a narrow zone along the western range-front of the Tobin Range.

The southern end of the 1915 rupture zone is in the Sou Hills transverse ridge, where the 1915 ruptures, as well as prehistorical ruptures, are distributed over an approximately 8 km wide zone, rather than being confined to a well-defined range-front (Fig. 2) (plate 1 of Wallace 1984b). The displacement measured on scarps in the Sou Hills is less than that of the main part of the rupture zone (Wallace 1984b, Fonseca 1988). The strike of the fault scarps also differ; Fonseca (1988) reported about a 20° change in strike compared to the strike of the main rupture zone to the north. Prehistorical fault scarps in the Sou Hills show the same pattern and style of the 1915 rupture.

The Sonoma segment

North of the Pleasant Valley segment is another ~40 km long, late Quaternary fault scarp which flanks the eastern side of the Sonoma Range (fig. 4 of Wallace 1989). Its southern end and the northern end of the Pleasant Valley segment form a geometrically complex area (fig. 4 of Wallace 1989). Thus, the fault scarps along the Sonoma Range probably constitute another segment of the fault system, and the segment boundary is located in the area between the southern end of the Sonoma Range and the northern end of the Tobin Range (Fig. 2). This segment and its associated segment boundary will not be discussed in detail in this paper.

The Fairview Peak segment

The Fairview Peak segment, delineated by 1954 earthquake rupture, is to the east of Dixie Valley and parallels the Sand Spring segment at about the same latitude (Fig. 2). It is unclear how the Fairview Peak segment is related to the Dixie Valley–Pleasant Valley fault system. One possibility is that the 1954 Fairview Peak earthquake rupture may have no direct relation to the tectonic development of the Dixie Valley–Pleasant Valley active normal fault system, but was a trigger for the 1954 Dixie Valley earthquake. The 1954 Fairview Peak earthquake rupture is separated from the Dixie Valley active fault system by Fairview Peak, an uplifted mountain range that bounds the southern end of Dixie Valley on the west (Fig. 2). Discontinuous and subdued fault scarps along the western flank of Fairview Peak may be continuations and correlatives of the minor fault scarps along the central and northern part of the eastern side of Dixie Valley. Moreover, the Fairview Peak segment is located in the transitional area where the NNW-trending central Nevada and California Seismic Belt changes to NNE-trending (Fig. 1). The ruptures within the NNW-trending belt are characterized by predominantly right-slip, whereas the NNE-trending belt is dominated by normal dip-slip. The Fairview Peak earthquake rupture is clearly characterized by a predominant component of right-lateral strike-slip (Slemmons 1957, Zhang unpublished data), and it supports the suggestion that the Fairview Peak segment belongs to a different fault system.

Another possibility is that the Fairview Peak rupture zone is part of the southern end of the Dixie Valley–Pleasant Valley fault system but that the rupture pattern and process are complex at the end of this fault system. The rupture may break through one strand such as the Sand Spring segment at one time, and another strand such as the Fairview Peak rupture at another time. The Fairview Peak segment causes problems for fault segmentation hypotheses in the Dixie Valley fault system and further studies are needed.

CHARACTERISTICS OF SEGMENT BOUNDARIES

Based on the description give above, the Dixie Valley–Pleasant Valley active normal fault system can be divided into four geometric fault segments, which coincide with those defined by historical earthquakes. The geometry of each individual segment is relatively simple in contrast to the areas that separate them (Fig. 2). The boundaries of segments show much more complex geometric patterns than the interiors of segments. Three segment boundaries can be identified, from south to north they are: the Elevenmile Canyon, The Bend, and the Sou Hills areas. In the following discussion, we summarize the geometric and geomorphic characteristics of these segment boundaries.

(1) *Changes in strike*

Where segments enter boundary areas, the orientation of fault scarps often changes. Fonseca (1988) noticed the change of fault scarp orientation in the Sou Hills. The approximate N20°E trend of 1915 Pleasant Valley earthquake ruptures change to about N40°E in the Sou Hills (Fig. 2) (Wallace 1984). Where the Stillwater range-front fault splays out to the Sou Hills the trend of fault scarps changes from N30°E to N50°–80°E (Fig. 2). Fault scarps also change strike in The Bend area

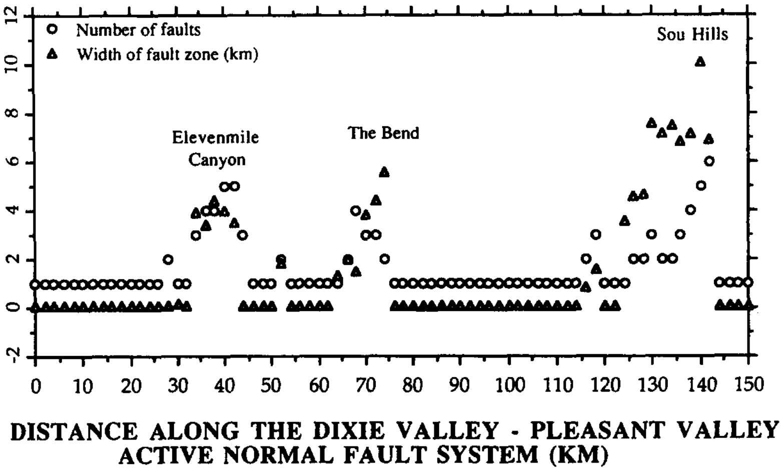

Fig. 7. The change of fault zone width and number of faults along the Dixie Valley–Pleasant Valley active normal fault system.

where the Stillwater segment overlaps the Dixie Valley segment. The Dixie Valley range-fault changes orientation from about N15°W to N60°–70°E (Fig. 3). The Holocene piedmont scarps also change from about N30°E along the Stillwater range-front to about N10°E in The Bend area. The Holocene fault scarps in the Elevenmile Canyon area trend N20°–30°E (Fig. 2), but the range-front faults of both the Dixie Valley segment to the north and the Sand Spring segment to the south generally trend north.

(2) *Increase in the number of faults and the width of the fault zone*

In general, most earthquake ruptures and fault scarps are confined to narrow zones of <200 m wide along a range-front in the Dixie Valley–Pleasant Valley fault system. The number of faults and width of the fault zone increase only at the segment boundaries (Fig. 7). In the Elevenmile Canyon area, the fault zone is as much as 4.5 km wide and the fault number increases to as many as five. In The Bend area there are four faults and the width of the fault zone increases to more than 5 km. In the Sou Hills area, six faults form a fault zone more than 11 km wide. Bell (1990 personal communication) noted the increasing of fault zone width and the number of faults correspond to areas of old alluvium (or bedrock) as do segment boundaries. This relation suggests two possibilities. One of them is that the central part of segments have as many faults as, and are as wide as, that of segment boundaries, but that downthrow on the youngest prehistorical rupture event along the range-front and the subsequent sedimentation buried all the faults in the piedmont and the valley floor except at segment boundaries. The other possibility is that the width of the fault zone and the number of faults increase only in the segment boundary areas as a result of buried or surface asperities. The 1915 and 1954 earthquake ruptures show that the increases appear only in the rupture termination areas. This evidence favors the second possibility.

(3) *Increase of complexity*

It is clear from Fig. 2 that the structural pattern is more complex in the segment boundary areas than in the main part of segments. In order to compare the complexity, we measured the fractal dimension, a quantitative measurement of complexity and irregularity. Fractal geometry is the description of forms more complex than the standard topographic dimension (Mandelbrot 1982). For a two-dimensional surface fault trace the fractal dimension D is between 1 and 2 ($D = 1$ suggests that the fault is a straight line, and $D = 2$ suggests that the fault is a very complex curve that continuously fills up the whole surface). The basic assumption in fractal analysis is self-similarity of the fault traces. Evidence suggests that earthquakes and faulting are fractal processes (King 1983, Scholz & Aviles 1986, Aviles *et al.* 1987, Okubo & Aki 1987). In this study we use the cycle method (Mandelbrot 1982) to measure the fractal dimension (Okubo & Aki 1987), because it covers the fragmentation property of the faulting. Figure 8 shows the measured fractal dimension along the fault system. It is obvious that the main segments have small fractal dimensions that imply less complexity, whereas the segment boundaries have large fractal dimensions and more complex geometry. Fractal geometry studies of several other normal fault systems in the Basin and Range province show similar relations (Zhang unpublished data). Because the segment boundary is the place of rupture termination, this study is independently consistent with the conclusions from seismological studies of earthquake rupture nucleation and termination (Aki 1989). It therefore indicates that the geometry of faults does indeed control or influence the propagation of earthquake rupture as inferred by Segall & Pollard (1980), and that the geometric segmentation method can probably be as reliable a method as those based on paleoseismicity and fault behavior.

(4) *Stepover of main segments*

Some stepovers or en échelon offsets of otherwise continuous fault scarps have been identified as indicators of segmentation boundaries (e.g. Wheeler 1987, 1989, Crone & Haller 1989, Machette *et al.* 1989), because they reflect along-strike discontinuities or geometric irregularities in the adjacent fault that may inhibit slip propagation (Sibson 1987). Stepovers can be divided into two categories: overlap and underlap stepovers (see Fig. 10 for definition), each of which create different geomorphic and geometric patterns in the stepover area. All three segment boundaries in the Dixie Valley fault system are associated with major (several km wide) stepovers of fault segments (Fig. 2). The Dixie Valley segment underlaps the Sand Spring segment to form an intensely deformed and uplifted area, Elevenmile Canyon, between them. The Stillwater Range overlaps the Dixie Valley segment to the right in The Bend area, where a complex rupture pattern and a major range-front embayment have formed. The Pleasant Valley segment underlaps the Stillwater segment to the right across the Sou Hills, an area of structural complexity and a mid-valley bedrock transverse ridge.

(5) *Change of range-front and valley-floor morphology*

This change includes major salients, embayments and bends of the range-front. Major range-front salients associated with segmentation boundaries have been found along the Lost River fault, Idaho (Crone *et al.* 1987, Crone & Haller 1989), and the Wasatch fault, Utah (Machette *et al.* 1987). The segmentation boundaries in the Dixie Valley–Pleasant Valley active normal fault system are also associated with morphological changes of the range-front. There are prominent range-front bends in both the Elevenmile Canyon area and The Bend area. The Sou Hills connects the southern end of the Tobin Range and the northern end of the Stillwater Range, and are a bedrock transverse zone. The valley floor is uplifted in the two underlap areas, Elevenmile Canyon and Sou Hills (Fig. 6). The embayment of the range-front and subsidence of the valley floor formed in The Bend area are associated with overlap of the two major segments (Fig. 6).

The geometric and geomorphic characteristics described above suggest that fault segment boundaries may be identifiable even without paleoseismicity and behavioral data, illustrated by the Dixie Valley–Pleasant Valley active normal fault system. Thus, these characteristics can probably be used as criteria for segmentation of other normal fault systems. Of course they are not complete criteria. Other characteristics of segment boundaries are identified along other normal fault systems in the Basin and Range (Crone & Haller 1989, Machette *et al.* 1989). Combinations of at least several criteria must be used when applying these characteristics to segment a fault system. A scaling factor must also be considered (Crone & Haller 1989, Zhang unpublished data). For example, bends or stepovers of <1 km width are probably not sufficient to terminate a rupture of 30–40 km long with more than 2 m displacement.

RUPTURE TERMINATION AND SEGMENTATION BOUNDARIES

Segment boundaries are areas of structural or material heterogeneity where ruptures terminate (Aki 1984, Sibson 1987, 1989, Wheeler 1987). These areas are often characterized by a complex structural and geomor-

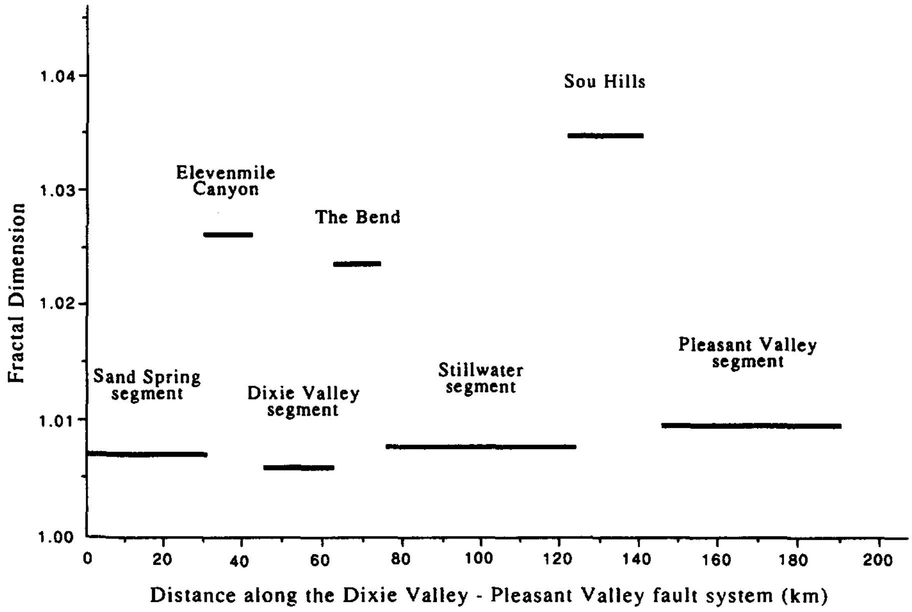

Fig. 8. Fractal dimension along the Dixie Valley–Pleasant Valley active normal fault system.

phic pattern (Wheeler 1987, Aki 1989). A change in complexity along a fault system may have a strong influence on the rupture propagation of an associated earthquake and on the persistence of segment boundaries. If this is true, the geometric characteristics and the patterns of complexity must be explained in terms of rupture propagation and termination.

Studies of fracture mechanics suggest that a propagating rupture often bifurcates at its end to form out-planar (different strike from the main part of rupture) extensional ruptures (e.g. Lawn & Wilshaw 1975, Bahat 1982). Bifurcation at the ends of rupture zones has been recognized for many earthquake rupture zones and active faults (Bahat 1982, 1984, Deng & Zhang 1984, Bruhn *et al.* in press). Bifurcation may be the nature of brittle fracture propagation, at least for some earthquake ruptures. Bruhn *et al.* (in press) suggests that bifurcation may be a useful criteria for identifying fault segment boundaries. It is unknown where the 1954 Dixie Valley rupture initiated, but it was preceded by the Fairview Peak earthquake 4 min earlier to the south. Thus, major propagation of the Dixie Valley earthquake rupture was probably northward although bilateral rupture is suggested (Doser 1986). The complex rupture pattern in The Bend area associated with the Dixie Valley earthquake may result from bifurcation at the rupture termination. This bifurcation may cause a change in rupture orientation, increase of fault zone width, and increase of rupture number near the northern end of the 1954 earthquake rupture zone. These changes also occurred at other segment boundaries within the Dixie Valley fault system, and in other prehistorical fault systems of the Basin and Range province (Crone *et al.* 1987, Crone & Haller 1989, Bruhn *et al.* 1987, in press). Thus, it may be bifurcation at rupture terminations that creates the complexity of geometry at segment boundaries.

Another important aspect in understanding earthquake mechanics is slip distribution. This includes the slip distribution of single earthquake events along segments of fault zones and the overall slip along the entire fault system. Previous studies of the slip distribution along earthquake rupture zones show complex and irregular patterns (Thatcher & Bollina 1989), but all earthquake slip distributions seem to decrease toward the ends of a rupture zone. If uniform or linear slip is assumed along the entire fault system, then a slip deficit is left near the rupture termination area of each segment after each offset event, because of decreased displacement toward the rupture end (King 1986). Some of the slip deficit can be accommodated by aftershocks, background seismicity, foreshocks and creep, but the overall strain accumulation near the termination area is still higher than other parts of the fault system during subsequent loading. Therefore the termination area, the segment boundary or the barrier, may be a nucleation area for the next rupture (King 1986). There is no evidence in the Basin and Range province, however, that most earthquake ruptures have the tendency to initiate at barriers or segment boundaries that separate fault segments (Doser 1989) although the 1983 Borah Peak earthquake may initiate at the boundary of two segments of the Lost River fault system. Thus, an alternative view of slip distribution is that slip is not uniform along the entire fault system, and the so called 'slip deficit' is 'slip low', a characteristic of slip distribution along a rupture zone (Fig. 9). In this view, the rupture is terminated at a 'dilatation barrier' that does not require the creation of new structures or the development of large strain at rupture terminations (King 1986).

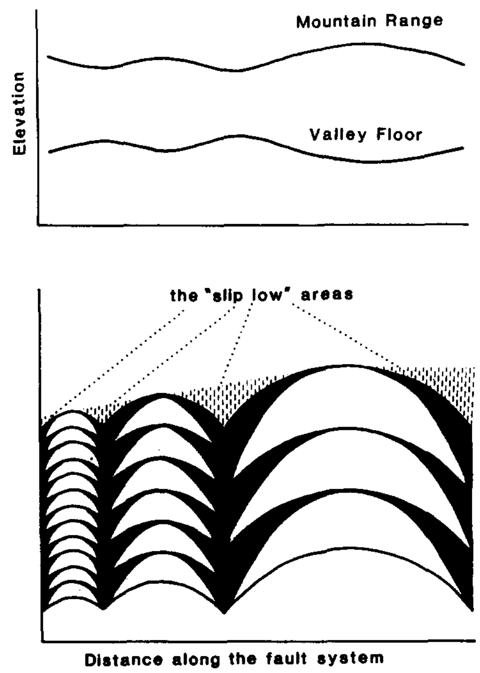

Fig. 9. Conceptual diagram to show the relations between 'slip low' and the elevation of mountain range and valley floor (modified from King 1986).

If the 'slip low' at the end of a rupture is the nature of earthquake rupture termination, and if recurrent earthquakes on the same segment retain even grossly characteristic slip distribution from one event to another (Schwartz & Coppersmith 1984, 1986), an accumulation of these slip deficits would create geomorphic anomalies, such as low topographic relief of the mountain range and high topographic relief of the valley floor (Fig. 9). The topographic change in the segment boundaries reported on this fault system (Fig. 6) and other fault systems (Crone *et al.* 1987, Schwartz & Coppersmith 1984, Bruhn *et al.* 1987, Crone & Haller 1989, Machette *et al.* 1989) are probably the result of the accumulation of the 'slip low' at the end of a segment. This may indicate the persistence of segment boundaries (Wheeler 1987, 1989).

In the Dixie Valley–Pleasant Valley active normal fault system, all three segment boundaries are associated with stepover of fault segments (Fig. 2). The valley floor is uplifted in the underlap boundaries (Elevenmile Canyon and Sou Hills), and subsided in a range-front embayment in the overlap segment boundary (The Bend) (Fig. 6). The faults at underlap segment boundaries appear to bear less displacement than that of main

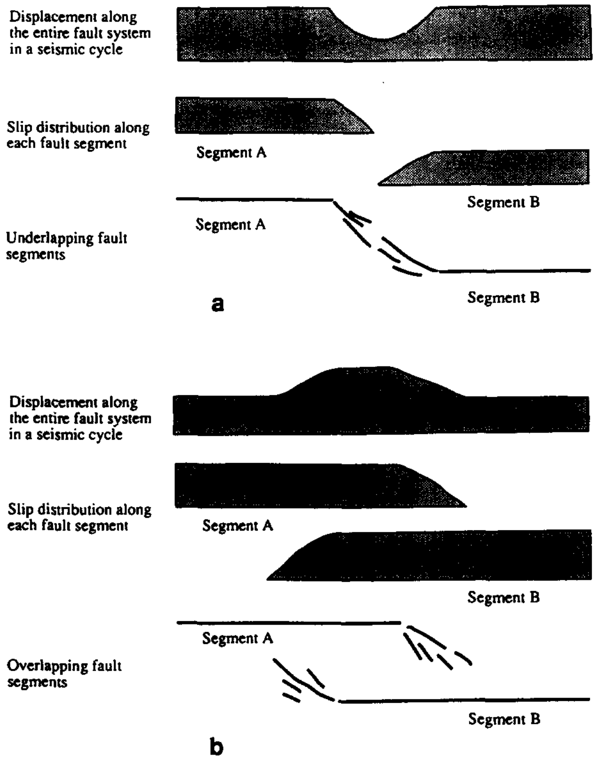

Fig. 10. Conceptual diagrams to show the relation among fault stepovers, slip distribution along each fault segment and total displacement along the entire fault system after a seismic cycle. (a) shows the case of underlap, where the total displacement in the segment boundary is less than that along the main segments. (b) shows the case of overlap, where the total displacement in the segment boundary is larger than that along the main segments.

segments. As shown in Fig. 10(a), when rupture along one segment enters the segment boundary, it ruptures only part of the segment boundary with small displacements, as was the case in the Sou Hills during the 1915 Pleasant Valley earthquake (Wallace 1984b, Fonseca 1988). The same thing occurs during subsequent events when ruptures along the adjacent segment enter the segment boundary. Thus, the total displacement of the entire fault system is less in underlap areas than along main segments after a seismic cycle (Fig. 10a). The overlap segment boundary, on the other hand, may have larger total displacements than along the main segments, because displacements on two different segments overlap in a seismic cycle (Fig. 10b).

CONCLUSIONS

The complexity of fault geometry has a strong influence on the propagation of earthquake ruptures (Segall & Pollard 1980, Okubo & Aki 1987, Aki 1989) because it forms structural heterogeneity that tends to inhibit rupture propagation. This implies that there should be a relation between fault segmentation and the complexity of fault geometry, and that the segmentation of fault zones should be indicated by their geometric patterns. The Dixie Valley–Pleasant Valley active normal fault system provides an excellent example to establish the geometric method of normal fault segmentation in the Basin and Range province owing to its recent movement. Studies of this active normal fault system indicate that there are changes of both geometric pattern and geomorphic character in the segment boundary areas. Within the segment boundary areas the following changes are noted: the strike of ruptures changes, the width of the fault zone and number of faults within the fault zone increase, and the geometry of fault zones is much more complex, as indicated by their fractal dimension. Segment boundaries are often associated with stepovers or discontinuities within the fault zone. Changes in geomorphology of the range-front and the valley floor also coincide with segment boundaries, reflecting their long-term persistence. Combinations of these characteristics may be used in conjunction with other methods as criteria for normal fault segmentation in the Basin and Range province.

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REFERENCES

Aki, K. 1984. Asperities, barriers, characteristic earthquakes and strong motion prediction. *J. geophys. Res.* **89**, 5867–5872.

Aki, K. 1989. Geometric features of a fault zone related to the nucleation and termination of an earthquake rupture. *U.S. geol. Surv. Open-file Rep.* **89-315**, 1–9.

Allen, C. R. 1967. The tectonic environments of seismically active and inactive areas along the San Andreas fault system. *Stanford Univ. Publs Geol. Sci.* **11**, 70–82.

Aviles, C. A., Scholz, C. H. & Boatwright, J. 1987. Fractal analysis applied to characteristic segments of the San Andreas fault. *J. geophys. Res.* **92**, 331–344.

Bahat, D. 1982. Extensional aspects of earthquake induced ruptures by an analysis of fracture bifurcation. *Tectonophysics* **83**, 163–183.

Bahat, D. 1984. New aspects of rhomb structures. *J. Struct. Geol.* **5**, 591–601.

Bakun, W. H., Stewart, R. M., Bufe, C. G. & Marks, S. M. 1980. Implication of seismicity for failure of a section of the San Andreas fault. *Bull. seism. Soc. Am.* **70**, 185–201.

Bell, J. W. 1984. Quaternary fault map of Nevada–Reno sheet. *Nevada Bur. Mine & Geol. Map* **79**.

Bell, J. W. & Katzer, T. 1987. Surface geology, hydrology, and late Quaternary tectonics of the IXL Canyon area, Nevada, as related to the 1954 Dixie Valley earthquake. *Bull. Nevada Bur. Mines & Geol.* **102**, 1–50.

Bell, J. W. & Katzer, T. 1990. Timing of late Quaternary faulting in the 1954 Dixie Valley earthquake area, central Nevada. *Geology* **18**, 622–625.

Bell, J. W., Slemmons, D. B. & Wallace, R. E. 1984. Roadlog: Reno to Dixie Valley–Fairview Peak earthquake areas, field trip 18: Geological Society of America, 1984 Annual Meeting. *Western Geol. Excursions* **4**, 425–472.

Bruhn, R. L., Gibbler, P. R. & Parry, W. T. 1987. Rupture characteristics of normal faults: an example from the Wasatch fault zone, Utah. In: *Continental Extensional Tectonics* (edited by Coward, M. P., Dewey, J. F. & Hancock, P. L.). *Spec. Publs geol. Soc. Lond.* **28**, 337–353.

Bruhn, R. L., Yonkee, W. A. & Parry, W. T. In press. Geometrical barriers in extensional fault zones. *Tectonophysics* **175**.

Crone, A. J. & Haller, K. M. 1989. Segmentation of Basin and Range normal faults: Examples from east-central Idaho and southwestern Montana. *U.S. geol. Surv. Open-file Rep.* **89-315**, 110–130.

Crone, A. J., Machette, M. N., Bonilla, M. G., Lienkaemper, J. J., Pierce, K. L., Scott, W. E. & Bucknam, R. C. 1987. Surface faulting accompanying the Borah Peak earthquake and segmentation of the Lost River fault, central Idaho. *Bull. seism. Soc. Am.* **77**, 739–770.

Deng, Q. & Zhang, P. 1984. Research on the geometry of shear fracture zone. *J. geophys. Res.* **89**, 5699–5710.

dePolo, C. M., Clark, D. G., Slemmons, D. B. & Aymard, W. H. 1989. Historical Basin and Range Province surface faulting and fault segmentation. *U.S. geol. Surv. Open-file Rep.* **89-315**, 131–162.

Doser, D. I. 1986. Earthquake processes in the Rainbow Mountain–Fairview Peak–Dixie Valley, Nevada, region 1954–1959. *J. geophys. Res.* **90**, 12,572–12,586.

Doser, D. I. 1989. The character of faulting processes of earthquakes in the intermountain region. *U.S. geol. Surv. Open-file Rep.* **89–315**, 163–180.

Eaton, J. P., O'Neill, M. E. & Murdock, J. N. 1970. Aftershocks of 1966 Parkfield–Cholame earthquake: A deailed study. *Bull. seism. Soc. Am.* **60**, 1151–1197.

Fonseca, J. 1988. The Sou Hills: a barrier to faulting in the central Nevada seismic belt. *J. geophys. Res.* **93**, 475–489.

Hecker, S. 1985. Timing of Holocene faulting in part of the seismic belt, west-central Nevada. Unpublished M.S. thesis, University of Arizona, Tucson.

King, G. C. P. 1983. The accommodation of large strains in the upper lithosphere of the earth and other solids by self-similar fault system: The geometric origin of *b*-value. *Pure & Appl. Geophys.* **121**, 761–815.

King, G. C. P. 1986. Speculations on the geometry of the initiation and termination process of earthquake rupture and its relation to morphology and geological structure. *Pure & Appl. Geophys.* **124**, 567–585.

Knuepfer, P. L. K. 1989. Implications of the characteristics of end-points of historical surface fault ruptures for the nature of fault segmentation. *U.S. geol. Surv. Open-file Rep.* **89-315**, 193–228.

Lawn, B. R. & Wilshaw, T. R. 1975. *Fracture of Brittle Solids*. Cambridge University Press, Cambridge, U.K.

Machette, M. N., Personius, S. F., Nelson, A. R., Schwartz, D. P. & Lund, W. R. 1989. Segmentation models and Holocene movement history of the Wasatch fault zone, Utah. *U.S. geol. Surv. Open-file Rep.* **89-315**, 229–245.

Mandelbrot, B. B. 1982. *The Fractal Geometry of Nature*. W. H. Freeman, San Francisco.

Okaya, D. A. & Thompson, G. A. 1985. Geometry of Cenozoic extensional faulting: Dixie Valley, Nevada. *Tectonics* **4**, 107–125.

Okubo, P. G. & Aki, K. 1987. Fractal geometry in the San Andreas fault system. *J. geophys. Res.* **92**, 345–355.

Page, B. M. 1965. Preliminary geological map of a part of the Stillwater Range, Nevada. *Nevada Bur. Mines & Geol. Map* 28.

Reasenberg, P. & Ellsworth, W. L. 1982. Aftershocks of the Coyote Lake, California, earthquake of August 6, 1979: A detailed study. *J. geophys. Res.* **87**, 10,637–10,655.

Segall, P. & Pollard, D. D. 1980. Mechanics of discontinuous faults. *J. geophys. Res.* **85**, 4337–4350.

Scholz, C. H. & Aviles, C. H. 1986. The fractal geometry of faults and faulting. In: *Earthquake Source Mechanics—Maurice Ewing Series* **6** (edited by Das, S., Boatwright, J. & Scholz, C. H.). *Am. Geophys. Un. Geophys. Monogr.* **37**, 147–155.

Sibson, R. H. 1987. Effects of fault heterogeneity on rupture propagation. *U.S. geol. Surv. Open-file Rep.* **87-673**, 362–373.

Sibson, R. H. 1989. Earthquake faulting as a structural process. *J. Struct. Geol.* **11**, 1–14.

Schwartz, D. P. 1988. Geologic characterization of seismic sources moving into the 1990s. *Geotech. Spec. Publs Am. Soc. Civil Engrs* **20**, 1–42.

Schwartz, D. P. & Coppersmith, K. J. 1984. Fault behavior and characteristic earthquakes: Examples from the Wasatch and San Andreas Fault zones. *J. geophys. Res.* **89**, 5681–5698.

Schwartz, D. P. & Coppersmith, K. J. 1986. Seismic hazard: new trends in analysing using geologic data. In: *Active Tectonics* (edited by Wallace, R. E.). National Academy Press, Washington, DC, 215–230.

Slemmons, D. B. 1957. Geological effects of the Dixie Valley–Fairview Peak, Nevada, earthquakes of December 16, 1954. *Bull. seism. Soc. Am.* **47**, 353–375.

Swan, F. H., III, Schwartz, D. P. & Cluff, L. S. 1980. Recurrence of moderate to large magnitude earthquakes produced by surface faulting on the Wasatch fault zone, Utah. *Bull. seism. Soc. Am.* **70**, 1431–1462.

Sykes, L. 1971. Aftershock zones of great earthquakes, seismic gaps, and earthquake prediction of Alaska and Aleutians. *J. geophys. Res.* **76**, 8021–8041.

Thatcher, W. & Bonilla, M. G. 1989. Earthquake fault slip estimation from geologic, geodetic, and seismologic observations: implications for earthquake mechanics and fault segmentation. *U.S. geol. Surv. Open-file Rep.* **89-315**, 386–399.

Wallace, R. E. 1977. Profiles and ages of young fault scarps, north-central Nevada. *Bull. geol. Soc. Am.* **88**, 1267–1281.

Wallace, R. E. 1984a. Patterns and timing of late Quaternary faulting in the Great Basin Province and relation to some regional tectonic features. *J. geophys. Res.* **89**, 5763–5769.

Wallace, R. E. 1984b. Fault scarps formed during the earthquakes of October 2, 1915, Pleasant Valley, Nevada and some tectonic implications. *Prof. Pap. U.S. geol. Surv.* **1274-A**, 1–33.

Wallace, R. E. 1989. Fault-plane segmentation in brittle crust and anisotropy in loading system. *U.S. geol. Surv. Open-file Rep.* **89-315**, 400–408.

Wallace, R. E. & Whitney, R. A. 1984. Late Quaternary history of the Stillwater seismic gap. *Bull. seism. Soc. Am.* **74**, 301–314.

Wheeler, R. L. 1987. Boundaries between segments of normal faults—criteria for recognition and interpretation. *U.S. geol. Surv. Open-file Rep.* **87-673**, 385–398.

Wheeler, R. L. 1989. Persistent segment boundaries on Basin and Range normal faults. *U.S. geol. Surv. Open-file Rep.* **89-315**, 432–444.

Whitney, R. A. 1980. Structural–tectonic analysis of northern Dixie Valley, Nevada. Unpublished M.S. thesis, University of Nevada, Reno, Nevada.