Overlapping Faults, Intrabasin Highs, and the Growth of Normal Faults

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ABSTRACT
Normal fault systems bounding extensional basins are typically adjoined by a series of subbasins separated by intrabasin highs. The strata within these basins form syndepositional anticlines and synclines whose axes are transverse to the strike of the main bounding fault. One possible explanation for these intrabasin highs is that they result from persistent along-strike deficits in fault displacement. Such deficits are incompatible with scaling relationships observed between fault displacement and length based on large populations of faults. We present data from active normal faults within the Basin and Range province and from inactive normal faults of the Newark basin of eastern North America demonstrating a clear correlation between the along-strike position of overlapping splay faults and the location of intrabasin highs as well as syndepositional transverse folds. Summed displacements for all faults within an intrabasin high are comparable to the displacements on faults bounding flanking subbasins. Older synextensional deposits exhibit localized tilt maxima within subbasins flanking an intrabasin high whereas younger units exhibit uniform tilt patterns across the entire region. Footwall elevation profiles, used as a proxy for fault displacement, define uniform arcuate patterns independent of along-strike position of intrabasin highs. These characteristics of hanging walls and footwalls suggest that intrabasin highs do not represent locations of long-term fault-displacement deficits, but rather are the location of anastomosing fault segments, which upon linking together, rapidly compensate for initial displacement deficits by increased displacement distributed over several splay faults.

Introduction
Hanging walls of active normal faults are typically observed to form a series of subbasins separated by intrabasin highs. These intrabasin highs commonly expose hanging wall basement in direct contact with the footwall whereas hanging wall basement on either side of the intrabasin high is buried by several kilometers of basin fill. In other instances, intrabasin highs are buried during subsequent sedimentation, resulting in syndepositional transverse synclines and anticlines that are present adjacent to normal faults within several ancient extensional basins (e.g., Wheeler 1939; Withjack and Gallagher 1983; Schlische 1992, 1993; Schlische and Reynolds 1992).

Intrabasin highs commonly serve to define the boundary between two along-strike fault segments of larger active normal faults or fault systems (e.g. Wheeler 1987, 1989). Studies of rupturing on large normal faults suggested that intrabasin highs are regions of persistent barriers to fault rupture, thus resulting in long-term displacement deficits (Schwartz and Coppersmith 1984; Wheeler 1987, 1989; Crone and Haller 1989; Zhang et al. 1991). Furthermore, Wheeler (1987, 1989) proposed that fault segmentation can be established by gravity profiles run parallel to the fault trace. He argued that gravity highs represent shallow basement rocks of partially buried intrabasin highs. These observations constitute a “barrier” model in which the growth or propagation of the fault is restricted at several along-strike locations, usually coincident with intrabasin highs. This provides the basis for the “characteristic earthquake” model used to estimate potential seismic moment and earthquake recurrence intervals (Schwartz and Coppersmith 1984).

Even though segment pinning is thought to be a common occurrence, observations of historic ruptures show a clear pattern of individual rupture events propagating through intrabasin highs (Knuepfer 1989; dePolo et al. 1991). Based
on similar observations made after the 1983 Borah Peak earthquake in Idaho, Crone and Haller (1989) applied the term “leaky barrier” to intrabasin highs that inhibit but do not stop rupture propagation; Wheeler and Krystinik (1987, 1988) used the term “nonpersistent rupture barrier” to define an intermittent barrier. Furthermore, the barrier model predicts irregular along-strike displacement profiles. This, however, is inconsistent with observations on individual or isolated normal faults as well as kinematically linked fault systems which exhibit a uniform displacement distribution, commonly resembling a bell-shaped curve (Chapman et al. 1978; Muroaka and Kamata 1983; Barnett et al. 1987; Walsh and Watterson 1987, 1990, 1991; Peacock and Sanderson 1991; Peacock 1991; Dawers and Anders 1992; Dawers et al., 1993).

In this paper we use footwall uplift patterns, the distribution of overlapping faults along strike, comparisons of total displacement between intrabasin highs and flanking subbasins, tilting patterns of hanging wall units during fault growth, and changes in the patterns of sedimentation to show that some intrabasin highs are not locations of persistent displacement deficit but rather are along-strike regions where displacement during successive ruptures is distributed over numerous overlapping faults. Moreover, we will suggest that intrabasin highs mark the locations two originally isolated fault segments linked together.

The Role of Overlapping Faults in Fault Growth

Well-exposed intrabasin highs typically reveal a history of complex faulting, and many faults within these areas overlap in the direction of extension. Overlapping faults are of two broad types: those that overlap but at some point intersect along strike and those whose surface traces do not intersect (figure 1). Many of the faults that do not appear to link at the surface may do so in the subsurface. Overlapping faults that do not link together at the surface form bridge or relay ramp structures between the segments (figure 1a; Ramsay and Huber 1987; Larsen 1988; Peacock and Sanderson 1991). When propagating fault tips intersect their respective opposite segments, a fault-bounded block is formed that is referred to as a “horse” in map view and “rider block” (Gibbs 1984) or “step-up” (Stewart and Hancock 1991) in cross section (figure 1b). The net effect of overlapping faults is to distribute displacement over several faults, thereby reducing the size of the depression available for sediment deposition and forming an intrabasin high (figure 1). As long as displacement is distributed on several parallel faults, an intrabasin high will persist. Once one of the overlapping faults dominates in accruing displacement, the region of overlap will subside along with the adjoining basins and become a buried intrabasin high. If overlapping faults do not cross-cut one another, they can be active concurrently (Jackson and McKenzie 1983).

The influence of these overlapping faults on developing normal fault systems can be deciphered by examining the uplift patterns of the footwall and the patterns of sedimentation in the associated hanging wall basins. Locating normal faults whose uplift history is preserved and whose basin fill sediments are sufficiently exposed for study is difficult. Active normal faults, like many in the Basin and Range province, have well preserved footwalls, but the hanging wall basin fill is often masked by the youngest deposits. Older basins, like those formed by the rifting of North America and Africa, have well-exposed strata, but the associated footwalls have been removed by erosion. Consequently, we must combine observations from disparate fault systems in order to define a general model of normal fault growth, and wherever possible, use faults where data are available from both the hanging wall and the footwall.

Overlapping Faults and Hanging Wall Basins.

Following Wheeler (1987, 1989) we define strike-parallel gravity highs as the location of intrabasin highs separating hanging wall subbasins. Along-strike Bouguer gravity anomaly profiles for four Basin and Range Province faults and the locations of overlapping fault segments are presented in figure 2b-e. There is a strong correspondence between the location of gravity highs in the hanging wall and the location of overlapping faults, with some notable exceptions. Gravity highs adjacent to the Mollie Gulch/Leadore segments of the Beaverhead fault and the Pass Creek/Mackay segments of the Lost River fault do not correspond to overlapping faults at the surface. These may be locations of inactive intrabasinal splay faults; seismic reflection data could test this hypothesis. At two locations along the 340-km-long Wasatch fault of central Utah, overlapping faults do not correspond to gravity
highs in the hanging wall. Both of the locations are areas where the overlap is relatively recent (lighter shading in figure 2b) compared to overlap at intrabasin highs (Davis 1983a, 1983b, 1985). Nevertheless, the correlation between overlapping faults and gravity highs within the hanging wall of these four faults is good.

Similar to the active normal faults of the Basin and Range Province, intrabasin highs within the early Mesozoic Newark rift basin of New York, New Jersey and Pennsylvania (see summary in Schlische 1992) also correlate well with the location of overlapping faults. The border fault system consists of a number of generally right-stepping fault segments. In the southwest part of the Newark basin, a set of small relay ramps is associated with overlapping fault segments (figure 3c). The presence of basement rocks in the zones of overlap suggests minimal fault displacement in these areas. A series of warps or folds are also associated with the segmented border fault system. The axes of the folds are subperpendicular to the border fault system, and the amplitude of folding decreases away from the fault system, indicating that the folds are fault-related. Furthermore, transverse anticlines occur at the zones of fault overlap, whereas the transverse synclines occur near the present-day centers of the fault segments (figure 3c). This suggests that the anticlines formed in areas of locally lower fault displacement and the synclines formed in areas of locally higher displacement. The transverse anticlines are thus a variety of intrabasin high.

In the Jacksonwald syncline, stratigraphic marker units are thickest in the hinge region and thinnest on both limbs (figure 3c; Schlische 1992). Seismic reflection and drill hole data from the Sassamannsville area (Reynolds et al. 1990; Schlische and Reynolds 1992; D. J. Reynolds, pers. commun. 1992) indicate that (1) the Lockatong Formation is thickest in synclinal troughs and thinnest in the anticlinal crests; (2) the contact between basement rocks and synrift strata is conformably folded with overlying synrift strata; and (3) the border faults are not folded in the subsurface. The transverse folds in the southwestern Newark basin therefore formed syndepositionally.

Although the transverse anticlines likely formed as a consequence of local displacement deficits associated with overlapping faults and relay ramps, the anticlines and the flanking synclines are not basin-wide phenomenon. The folds clearly decrease in amplitude away from the border fault system, and the large-scale basin geometry is not affected by fault segmentation (figure 3b). For example, the trace of the basement-synrift contact is broadly concave toward the border fault system, suggesting a syncline-shaped basin in longitudinal section (figure 3b, section C-C’). This geometry is also corroborated by stratigraphic data, which indicate that formations thicken from the longitudinal ends of the basin toward its center (a proxy of this is the width of the Lockatong Formation in figure 3b; Schlische 1992, 1993). We therefore propose that the displacement deficits responsible for the intrabasin highs are a relatively near-surface phenomenon associated with a segmented fault system but that the large-scale basin geometry was controlled by the displacement distribution on a kinematically linked fault system.

In the central part of the Newark basin, a series of NNE-striking faults splay from the NE-striking border fault segments (figure 3b). These intrabasinal splay faults are believed to have formed normal to the Mesozoic extension direction (in that they are more nearly parallel to Mesozoic dikes) rather than somewhat oblique as for the border fault system (Schlische 1992, 1993). A number of relay ramps are present between the overlapping fault segments, where basin depth is minimal. Basin depth in the zone of multiple intrabasinal splay faults is also less than that immediately to the northeast, yet the cumulative fault heave in the zone of extensive faulting (based on separation of distinctive marker horizons) is only slightly greater (figure 3b).

An intrabasin high within the Triassic-Jurassic Connecticut Valley basin of Connecticut and Massachusetts separates the synclinal Deerfield subbasin from the larger, asymmetric, synclinal Hartford subbasin (figure 3a). The intrabasin high is characterized by a number of basement blocks in contact with the border fault system and exclusively fluvial synrift strata. Longitudinal streams in both subbasins flowed away from the intrabasin high (Hubert et al. 1992). Lacustrine strata and lava flows within the two subbasins thin significantly toward the intrabasin high (figure 3a), indicating that it stood topographically high during the Jurassic accumulation of these units. Furthermore, within the intrabasin high, multiple normal faults overlap in the extension direction (see Wise 1992). The intrabasin high likely marks an area of some displacement deficit as exemplified by the narrow
width of the basin adjacent to the intrabasin high (figure 11a). This region likely marks the site where the two subbasins grew together (Wise 1992; Schlische 1993); faulting and basin formation ceased before the displacement deficit could be overcome.

**Overlapping Faults and Footwall Uplift Patterns.** Footwall uplift is the isostatically driven viscoelastic response to normal faulting (e.g., Vening Meinesz 1950; Zandt and Owens 1980; Spencer 1984; Wernicke and Axen 1988; Buck 1988). Although exactly what role isostasy plays in generating uplift is debated (cf. Barnett et al. 1987, Gibson et al. 1989), it is agreed that footwall uplift is proportional to displacement. Thus, along-strike displacement variations on range-bounding normal faults should be reflected in corresponding footwall elevation patterns. Consequently, Schwartz and Coppersmith (1984) suggested that observed footwall elevation lows in the Wasatch Mountains correlate with locations of fault deficit or segment boundaries on the Wasatch fault. Most of the segment boundaries they defined are located at intrabasin highs (see figure 2e). A multi-humped footwall elevation pattern, as shown in figure 2a(i) (solid lines), would result from long-term pinning of the rupture endpoints.

Footwall elevation profiles of four large Basin and Range province normal faults (figures 2b-e) more closely resemble the dashed line in figure 2a(ii). These elevation patterns have an arcuate distribution that flattens toward the center, with the highest elevation slightly off-center and shoulders somewhat steepened. Assuming footwall elevation is proportional to displacement, the pattern of elevation should reflect the displacement history of each fault segment as discussed by Schwartz and Coppersmith (1984). Qualitatively, there appears to be little correlation between the gravity highs (intrabasin highs) and footwall elevation lows (figures 2b-e).

To test the hypothesis that footwall elevation lows do not correlate with segment boundaries, a more quantitative assessment of the relationship among segment boundaries defined by neotectonic workers, intrabasin highs, gravity highs, and footwall elevation is necessary. To determine footwall uplift, we averaged digital elevation data over a 10-km-wide swath parallel to the range bounding fault and including the estimated line of maximum footwall elevation [see Simpson and Anders (1992) for a further description of this technique]. This method compensates for artifacts induced by choosing one elevation transect that may be dominated by a single anomalously high mountain or a single stream-cut valley.

Once the average elevation profile is determined, it is matched to the along-strike positions of intrabasin highs defined by topography and gravity anomalies and the location of segment boundaries. The range-bounding fault systems of the four mountains shown in figure 2b-e include 26 defined segment boundaries (vertical lines in figure 2b-e), not all of which correspond to intrabasin highs; however, almost all intrabasin highs have been assigned as segment boundaries. Because the lengths of segments are variable, all 26 were normalized to 12 km (equal to about half of the average segment length). For each profile, the midpoint of a defined fault segment was assigned a zero evolution, and data from one side of the midpoint was inverted so that it could be readily compared with the other side, and normalized data could be regressed in order to assess slope. If segment boundaries are zones of displacement deficit as predicted by the barrier model, then the midpoint of each segment should have the highest elevation, and all spot samplings of elevation should have values less than zero. However, over 65% of the elevation data are above the zero elevation line, and the slope of the regression is positive (figure 2f). Clearly, segment rupture boundaries do not correlate with footwall elevation lows.

Superimposed on figure 2f are three slopes (dashed lines) that result from a synthetic fault growth model, which is based on the conservative assumption that the ratio of footwall uplift and hanging wall downdrop (1:5.5) is the same as the coseismic measurements made by Stein and Barrientos (1985) after the 1983 Ms = 7.3 Borah Peak earthquake. The model assumes an initial footwall elevation profile equivalent to the Beaverhead fault (figure 2c), that the synthetic fault is divided into six segments, and that the elevation of each of the five segment boundaries is fixed. The elevation at the center of each segment increases by an increment of 0.33 mm per year, with displacement decreasing to zero at the segment boundaries. Recurrence interval is set at 1000 years for the middle two segments, 2000 years for the two end segments, and 1500 for the intermediate two segments; paleoseismic studies of Basin and Range normal faults indicated that central fault segments have smaller recurrence
intervals than segments at the ends of large fault systems (e.g., Crone and Haller 1989; Machette et al. 1991). Between 0 and 1 m.y. the model results are indistinguishable from the data recorded on the 26 segment boundaries (figure 2f). For intervals over 1 m.y., a pronounced discrepancy is evident between the observed data and the model. It is apparent from the slopes generated by the synthetic fault growth model that footwall elevation differences resulting from pinning at segment boundaries should become measurable in less than a million years, which is significantly younger than the age of any of the faults studied. Furthermore, the 1:5.5 ratio of uplift to downdrop measured for the Borah Peak rupture is inconsistent with longer-term measurements of 1:1-2 (Stein et al. 1988; King et al. 1988). This lower ratio would accentuate any differential relief produced by displacement deficits at segment boundaries.

Because erosion modifies elevation differences of tectonic origin, assessing relative uplift along strike is time dependent and, once uplift stops, the usefulness of footwall elevation pattern decreases with time. In addition, one could argue that a uniform footwall uplift profile across the entire length of the fault system might be produced by erosion preferentially acting on the footwall highs at the centers of fault segments. To maintain such a uniform profile for 1 m.y. requires an erosion rate of about 15 cm/yr at the centers of segments, which is two orders of magnitude higher than the highest measured erosion rates (e.g. Suppe 1985, his figure 1-3). Therefore, if segment pinning occurs for lengths of time in excess of 1 m.y., it should be detectable in the footwall elevation patterns. Moreover, if the ratio of footwall uplift to hanging wall downdrop is 1:1 and 2-meter ruptures recur every 1000 years, pinning would be detectable in elevation patterns on the order of 100 k.y.

Along-strike elevation changes expected from faulting deficits can be obscured by the flexural response of the footwall to differential uplift. Footwall flexural deflection measured normal to fault strike decreases to essentially zero at 10-15 km from the fault for several >100-km-long normal fault systems in the northeastern Basin and Range province (Anders et al. 1993) and at about 30 km for the Wasatch fault (Zandt and Owens 1980). The average spacing for segment boundaries is 22 km on the Lost River and Beaverhead faults and 34 km for the Wasatch fault. Thus, displacement deficits at segment boundaries should not be obscured by the flexural strength of the faults studied.

Intrabasin Highs as Regions of Fault Linkage. Studies of fault growth have demonstrated that longer faults often result from the linkage of several smaller fault segments (Segall and Pollard 1980, 1983; Ellis and Dunlap 1988; Peacock and Sanderson 1991). Furthermore, individual faults at all scales typically exhibit complex zones of anastomosing fault branches at their tips that are often referred to as horsetails. As discussed by Segall and Pollard (1980, 1983) and Granier (1985), two faults in the process of linking together will first join with one or more of the numerous available joint or shear surfaces, which in turn will eventually become the main surface upon which most displacement is accrued. Subsequent to linkage, the area surrounding the point of linkage can be identified by the presence of multiple overlapping faults. Indeed, overlapping faults characterize the regions of intrabasin highs along the normal faults (figures 2b-c). For example, an intrabasin high along the Wasatch fault exhibits a complex pattern of secondary, overlapping faults in both the footwall and hanging wall (Bruhn 1992). The mere presence of overlapping faults does not indicate when the intrabasin high developed with respect to the flanking subbasins. In one scenario, originally favored by Anders (1990), the overlapping faults formed at the same time as the main fault strands (i.e., faulting was simultaneous in both basins and intrabasin highs). Simultaneous growth of overlapping faults has been called incidental by Walsh and Watterson (1990). Conversely, fault overlap may have developed through time as the tips of originally isolated fault segments propagated toward one another. In this case, the intrabasin highs should be younger features than the adjacent subbasins.

Detailed Observation of Fault Growth within Intrabasin Highs

We examined two intrabasin highs that are located along the southern sector of the 120 km long Beaverhead fault of east-central Idaho (figure 4) where excellent exposure permits establishment of a growth history of these highs. The northernmost of these, Middle Ridge, is composed of strata of the 10 Ma Medicine Lodge Beds overlain in the northeast by a mid-Quaternary gravel deposit (Scott 1982). Middle Ridge is
bounded on the south by several scarps that offset latest Quaternary deposits (Haller 1990). The fault defined by these scarps splay off the main range-bounding fault near Nicholia, Idaho (figure 4). North of Nicholia, scarps offsetting latest Quaternary surfaces are absent; however, older mid-Quaternary surfaces have been clearly displaced. The time of initial movement on the intrabasin splay forming the southern boundary of Middle Ridge can only be constrained to sometime after deposition of the Miocene Medicine Lodge Beds.

The southern intrabasin high is exposed northwest of the former town of Blue Dome (figure 4). Miocene through Quaternary deposits have been offset by at least four overlapping, west-dipping faults of approximately the same strike (Rodgers and Anders 1990). The Beaverhead fault in this area is younger than 6.6 Ma (Rodgers and Anders 1990). The best estimate of latest movement on the faults within the intrabasin high is >30 ka (Haller 1990). The section of the main Beaverhead fault between Nicholia and Blue Dome has ruptured several times in the late Quaternary. The scarps on latest Quaternary deposits are directly traceable to scarps in the bedrock to the northeast of Blue Dome. The newly formed scarps in the bedrock (Pennsylvanian Blazer Limestone) form a rider block that is locally referred to as the Blue Dome block (label BDB in figure 4). The offset of the Blazer Limestone on this part of the fault is only a few meters, suggesting that linkage of the Beaverhead fault east of Blue Dome occurred within the latest Quaternary.

An along-strike elevation profile of the southern Beaverhead Mountains is shown in figure 5a along with the hanging wall tilt pattern of the 6.6 Ma tuff of Blacktail. The 6.6 Ma tuff is uniformly tilted in both the two intrabasin highs and the intervening subbasins. This pattern is similar to the hanging wall tilt pattern observed on the northern segments of the 140 km Grand Valley fault (figure 5b), located 100 km to the southwest of the Beaverhead fault. For both faults the tectonic tilt is the same regardless of whether or not sampling is done on an intrabasin high or in the intervening subbasins; therefore, these intrabasin highs cannot have resulted from a displacement deficit relative to the subbasins.

Within the Blue Dome intrabasin high the 10 Ma Medicine Lodge Beds, the 8 Ma basalt flows, and the 6.6 Ma tuff of Blacktail are uniformly tilted (Rodgers and Anders 1990). To the south of Blue Dome a sequence of interbedded volcanic rocks, the 10.3 Ma tuff of Arbon Valley (Kellogg et al. 1989), an intermediate aged basalt flow, and the 6.6 Ma tuff of Blacktail exhibit progressively gentler tectonic tilts, indicating that the southern extremity of the Beaverhead fault was active before the splay faults in the Blue Dome intrabasin high. We therefore conclude that the intrabasinal splay faults in the Blue Dome region developed as the tips of the main faults propagated toward one another.

Although the Blue Dome intrabasin high developed later than the faults bounding the adjacent subbasins, the fault displacements are comparable. The summed hanging wall displacement on all intrabasinal splay faults (see figure 4, X-X’ cross section) is 1.5 km. Adding the approximately 1 km of displacement on the Beaverhead fault results in a cumulative displacement of 2.5 km. The displacement on the Beaverhead fault bounding the subbasin directly to the north of the intrabasin high is estimated to be 3 km based on proprietary seismic reflection lines. Using the tectonic tilting pattern of several ash-flow tuff units in the footwalls of the Beaverhead fault south of the intrabasin high, Anders et al. (1993) calculated a displacement of 2.9 km. Therefore, the summed displacement on all the faults in the region of the intrabasin high is only about 0.5 km less than the displacement of the basin-bounding faults in adjoining basins. The small discrepancy between displacement across this region may be accounted for by insufficient displacement to date in the vicinity of the intrabasin high following the linkage of the propagating fault segments. Alternatively, Peacock and Sanderson (1991) suggested that small displacement deficits will persist in areas of linked faults, most likely due to distributed (ductile) deformation (including folding) that preceded final linkage.

Figure 6 presents an idealized model for the growth of the Middle Ridge and Blue Dome intrabasin highs in the hanging wall of the southern Beaverhead fault. Three isolated faults initiated between 10.3 and 6.5 Ma in response to hotspot-induced extension (figure 6a) (Rodgers and Anders 1990; Anders et al. 1989). Each fault formed an associated hanging wall basin, and the fault tips grew toward one another as displacement on the fault segments increased. A shift in the extension direction from the thermomechanical effect of the hotspot as it approached the southern terminus (left side) of
the Beaverhead fault (Anders and Sleep 1992) caused splay faults to form normal to the extension direction (figure 6b). Activity on the southern segment then increased as the hotspot passed closest to the Beaverhead fault (figure 6c). Faults that splayed off the southern segment formed the Blue Dome intrabasin high. The two northern segments were also linked together. Some time in the Quaternary after the hotspot passed beyond the Beaverhead fault, a zone of fault inactivity (or collapse shadow of Anders et al. 1989) encompassed the southernmost fault segment, causing sharply reduced extension rates, while increased extension rates affected the central fault segment (figure 6d). A new intrabasinal splay fault then formed, creating the Middle Ridge intrabasin high. Growth of the central segment completed linkage of the southern two segments and formed the Blue Dome block.

Evolution of Overlapping Fault Systems and Associated Basins: A General Model

We have seen that intrabasin highs (including relay ramps and transverse folds) are commonly associated with overlapping fault segments. Tilting relations at intrabasin highs in the Basin and Range province suggest that these regions continually evolved. This is not surprising considering that normal faults and associated half-graben basins grow in length as displacement accumulates (Watterson 1986; Walsh and Watterson 1988; Gibson et al. 1989; Cowie and Scholz 1992a, 1992b; Schlische 1993; Dawers et al. 1993). Next we present several end-member models for the evolution of overlapping fault segments and associated basins. We have excluded overlapping faults that have the opposite sense of dip.

In the first case (figure 7, case 1), the amount of overlap between two fault segments is zero. Thus, as the faults grow while displacement accumulates, the fault tips propagate toward one another in the same plane. Initially, each fault segment is associated with a discrete sedimentary basin. Shortly after linking together, the composite basin consists of two synclinal subbasins separated by an intrabasin high. The footwall elevation profile is multi-humped at the moment of linkage. The displacement deficit in the region of linkage is then overcome. The mechanical basis for this is described by Cowie and Scholz (1992b) who suggested that the yield strength at the fault tips governs the along-strike growth. The greatest concentration of stress is focused in the linkage area where there is a bulk deficit of displacement. If the linkage zone did not experience greater displacement rates, the newly linked fault zone would exhibit values of n<1, where the relationship between fault length (L) and displacement (D) is given by:

\[ D \propto L^n \]

and n is some exponent. Attempts to characterize D versus L scaling relationships indicate that n = 1 (Watterson 1986; Walsh and Watterson 1988; Scholz and Cowie 1990; Marrett and Allmendinger 1991; Dawers et al. 1993).

Because there are no overlapping splay faults for case 1, the intrabasin high will gradually disappear. This type of basin evolution can be recognized in the geologic record because the oldest synextensional stratal units will form restricted sequences marking the positions of the originally isolated subbasins; these deposits will be thickest near the centers of the original subbasins. The youngest stratal units deposited after linkage and the elimination of the displacement deficit are deposited basin-wide and thicken toward the center of the composite basin. The Connecticut Valley basin (figure 3a) may represent a good example of a Case 1-type basin in which faulting ceased before the displacement deficit was completely eliminated.

The second case involves two propagating fault segments that eventually begin to overlap in the direction of extension. The faults are so closely spaced that the interacting stress fields direct the propagating faults toward each other (Segall and Pollard 1980, 1983). There are two possible variations: (1) the back fault propagates at the expense of the front fault, and thus the former footwall of the front fault is transferred to the hanging wall of the back fault (figure 7, case 2a); (2) the front overlapping fault segment grows into the back fault and consequently the former hanging wall of the back fault is transferred to the footwall of the front fault, forming an intrabasin high in the zone of overlap (figure 7, case 2b). As long as displacement is distributed along both coevally active faults, overall basin depth will be shallower than in non-overlap areas, and the intrabasin high will persist.

In the third case (figure 7, case 3), widely spaced fault segments that overlap in the extension direction do not intersect. Nonetheless the summed fault displacement in the extension direction still forms the typical bell-shaped
profile. Synextensional stratal units are restricted to each subbasin and thicken toward the center of each subbasin. The two overlapping faults may or may not anastomose at depth. Whether or not this occurs, an intrabasin high will persist as long as the multiple overlapping segments actively accrue displacement.

As discussed above, there is a strong correlation between overlapping fault segments (and segment boundaries) and gravity highs. On the other hand, there is little or no correlation between footwall elevation profiles and the pattern of hanging wall subsidence. The lack of correlation can best be accounted for if the arcuate footwall elevation profile is a manifestation of displacement at depth rather than near-surface displacement on splay faults. In other words, footwall elevation profiles reflect the total displacement on major range-bounding fault systems whereas hanging wall basin depth represents the near-surface displacement on the nearest fault segment. As documented earlier, the summed displacements on overlapping faults at some intrabasin highs is comparable to that of the flanking subbasins.

Seismological Implications

Zones of normal fault overlap and associated intrabasin highs are commonly used to define fault segment rupture boundaries. If earthquake ruptures are indeed halted at geometric segment boundaries, then neotectonicists can estimate segment length and potential rupture length and thus assess potential earthquake magnitude and seismic risk. The observations presented in this report suggest that this approach may not always be valid.

First, many intrabasin highs are not associated with persistent displacement deficits and thus do not mark regions of repeated rupture termination (Knuepfer 1989; dePolo et al. 1991). Rather intrabasin highs commonly mark regions where displacement is accommodated on several faults that overlap in the extension direction. Thus, instead of stopping most ruptures, persistent intrabasin highs are regions where rupture will be transferred to a given fault splay fault one time and another splay the next. Such transfer reduces detectability because the total offset is distributed on several splays and therefore cumulative displacement on each fault will be less. Second, the concept of fault growth requires that some intrabasin highs—namely relay ramps—are not likely to stay fixed over the long-term.

Third, smoothly varying total fault displacement profiles (responsible for the uniform single-humped footwall elevation profiles and syncline-shaped basins in longitudinal section) suggest that many faults bounding and within the basin are kinematically linked at depth, even though they are clearly geometrically segmented at the surface. As most large-displacement earthquakes nucleate at depth (Scholz 1990), it is unlikely that upward- and outward-propagating ruptures will be significantly inhibited by surface discontinuities, although they will affect the near-surface displacement distribution and associated structures. Thus, we believe that fault segmentation and associated structures are not a good predictor of potential earthquake rupture lengths for assessing seismic risk. Instead, regional footwall uplift profiles, overall geometry of sedimentary basins, and summed displacement profiles normal to the extension direction will give a more accurate picture of regions of displacement deficit that are likely to mark persistent rupture termination locations.

Conclusions

Studies on faults at all scales show that larger faults are often the result of linking together of several smaller faults which interconnect to form a larger fault system. Our model of normal fault growth that suggests that hanging wall elevations highs, here called intrabasin highs, are locations of antecedent linkage. We further suggest that these intrabasin highs are regions of significant post-linkage strain accommodation and not regions of displacement deficit as previously suggested. Were this not the case, linked fault systems would exhibit a scaling relationship between displacement and length that is inconsistent with those of all previously studied fault populations. The strain in intrabasin highs is accommodated by distributed displacement on several overlapping faults that break-up the footwall and/or hanging wall into smaller blocks restricting the volume available for sediment deposition. In support of our model of normal fault growth we have shown: (1) intrabasin highs commonly correlate with regions of multiple overlapping faults; (2) summed displacement in intrabasin highs is approximately equal to displacement on the faults which bound adjoining subbasins; (3) footwall elevation profiles, a proxy for fault
displacement, are unaffected by the location of intrabasin highs; (4) the tilt of units deposited after formation of intrabasin highs have the same total tectonic tilt as the coeval units deposited within adjoining subbasins; and (5) the location of syndepositional anticlins in older exhumed basins correlates with the locations of overlapping faults or intrabasin highs. Our model also suggests that intrabasin highs appear no more likely to serve as segment boundaries than any another geometric feature and thus are unreliable for use in defining normal fault rupture lengths that are commonly used to assess seismic hazards.

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Figure 1. Geometry of overlapping faults. 

a. Faults overlap in the extension direction (north-south) but do not intersect in map view. Displacement decreases toward the ends of each segment, resulting in a relay ramp in the zone of overlap and a transverse anticline near the western end of the southern segment. In each cross section, the summed fault displacement is the same, resulting in shallow basins in the zones of overlap.

b. Faults overlap in the extension direction and intersect in map and perhaps cross-section view. This geometry is particularly likely if the extension direction changes from north-south to northeast-southwest. In map view, a fault horse is located in the zone of overlap; rider blocks are observed in cross-section view. In each cross section, the summed fault displacement is the same.
Figure 2. a. Hypothetical footwall elevation profile (solid) and depth to basement on hanging wall block (shaded) along segmented normal fault system. In (i) it is assumed that the fault segment boundaries are regions of slip deficit, whereas in (ii) the boundaries are regions of fault splaying and overlap. b-e. Elevation and gravity profiles for four large normal faults in the Basin and Range province. In all profiles the upper line is elevation based on highest elevation within a 10 km swath and the lower line is a Bouguer gravity anomaly profile. Shaded areas are regions of multiple fault overlap, lightly shaded areas are newly formed overlapping faults. Fault segments are delineated by vertical lines. b. Lost River fault, Idaho; gravity data from Bankey and Kleinkopf (1988), fault segmentation from Crone and Haller (1989). c. The Beaverhead fault of east-central Idaho; gravity data from Bankey and Kleinkopf (1988) and segment boundaries are from Crone and Haller (1989). d. Pleasant Valley fault, Nevada; gravity data from Erwin (1974), fault segmentation from Wallace (1984). e. Wasatch fault, Utah; gravity data from Zoback (1983); fault segmentation from Machette et al. (1991). f. Compilation of normalized elevations between segment ends and midpoints for 26 segments shown in b-e. The regression is fixed at the origin and has a slope of $9.2 \times 10^{-2}$. Dashed lines represent the slope of elevation data from a synthetic growth model after 1 m.y., 2 m.y. and 4 m.y. Erosion is not considered.
Figure 4. Generalized geological map of two intrabasin highs found along the southern portion of the Beaverhead fault in northeast Idaho. Modified from Crone and Haller (1989) and Rodgers and Anders (1990). Also shown is cross-section X-X’ through the Blue Dome intrabasin high.
Figure 5. Along-strike tectonic tilt of the 6.6 Ma tuff of Blacktail in the hanging walls of the (a) Beaverhead and (b) Grand Valley/Star Valley faults. Solid lines represent the elevation of the hanging wall with the locations of intrabasin highs indicated by dashed lines. The averaged tilt is represented by the horizontal line. Bars represent the quadratic error associated with measuring the absolute or tectonic tilt. Note that there is little correlation between the intrabasin highs and the post-6.6 Ma tilting of the tuff of Blacktail within the hanging wall. Map locations and tilt data are compiled in Anders et al. (1989), Rodgers and Anders (1990), and Anders et al. (1993).
Figure 6. Schematic diagram of the growth history of the southern Beaverhead fault during the last 10.3 Ma. Thick lines are faults, thin lines are structural contours of the basin fill. Pluses represent relative extension rates and the double-headed arrow is the extension direction.
Table 1. Features associated with four cases of overlapping, propagating normal faults. The top row shows the map patterns of faults and associated sedimentary basins. Fine lines represent contours of the basement-sediment interface of the hanging wall. The second row represents a view from the hanging wall normal to the fault surface; the dark shading represents the overlapping surface or intrabasinal splay fault. The light stipple represents the extent of the fault at an early stage in its evolution, whereas the dashed line represents its geometry following an increase in displacement. Subsequent rows show the expected behavior of the respective footwalls and hanging walls. 

**a. Case 1**: the faults grow together in plane. 

**b. Case 2a**: the rear overlapping fault intersects the front or basinward fault. 

**c. Case 2b**: the front or basinward overlapping fault grows into the rear fault. 

**d. Case 3**: the overlapping faults do not intersect either at the ground surface or the subsurface.