TECTONICS OF MID-TERTIARY EXTENSION ALONG A TRANSECT THROUGH WEST CENTRAL ARIZONA

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Abstract. Large-magnitude Miocene extension in west central Arizona occurred primarily along three imbricate, northeast dipping normal faults. The structurally highest of these faults, the gently dipping Buckskin-Rawhide detachment fault, accommodated approximately 66 km of crustal extension, whereas the two structurally lower faults accommodated a total of about 20 km extension. Due to this large-magnitude extension, an area at the Earth’s surface that was 10 to 20 km wide is now over a 100 km wide, and crystalline rocks with mid-Tertiary mylonitic fabrics, uncovered by detachment faulting, are exposed over roughly 2000 km² in the Harcuvar metamorphic core complex. Most of the upper plate of the Buckskin-Rawhide detachment fault was largely undeformed by normal extension; only the thin, tapered end of the upper plate was highly extended. During extension the lower plate must have flexed to conform to the listric underside of the upper plate and to have flattened to its present subhorizontal form as it was uncovered. Grooves on the underside of the upper plate were apparently imposed on the pliable lower plate as it was denuded, forming extension-parallel folds in the lower plate. Low flexural strength characterized the lower plate during denudation, and a highly mobile, low-viscosity deeper crust must have effectively decoupled the upper crust from the mantle lithosphere.

INTRODUCTION

The kinematics and dynamics of large-magnitude extension are a major focus of research in continental geodynamics. Cenozoic extensional strains of 100% or more are now recognized in some areas of the Basin and Range province, but in general it has been difficult to quantify such strains and to determine the kinematic evolution of rocks in affected areas. In this study we integrate data from previous studies in order to examine structural styles and magnitudes of extension along an extension-parallel transect in west central Arizona. Analysis of geologic maps, most of them recently published, reveals that extension in the study area was accommodated in the upper crust by displacement on three imbricate, regionally northeast dipping normal faults, the most important of which is the Buckskin-Rawhide detachment fault and its correlatives in the Harcuvar and Whipple mountains. Structural analysis and restoration of normal faults allow estimates of displacement on each of these faults and lead to a tectonic reconstruction that realigns Mesozoic structural features.

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Detachement faults have been considered to be the extensional analogs of thrust faults [Wernicke, 1981]. In detail, however, the mechanical behavior of the crust during large-magnitude extension may be quite different from that typically associated with crustal shortening. Specifically, the uplift and warping of large-displacement, low-angle normal faults (detachament faults) and their commonly mylonitic lower plates (metamorphic core complexes) during extension indicate that the flexural strength (resistance to bending) of detachment-fault lower plates was far less during extension than was the flexural strength of the lower plates of foreland fold and thrust belts during thrusting. Indeed, simplistic models of detachment faulting assuming essentially zero lower plate flexural strength yield fault geometries that are quite similar to some known geometries [Spencer, 1984].

Several types of evidence indicate that the Buckskin-Rawhide detachment fault and its down-dip extension as a ductile (e.g., plastic [see Rutter, 1986]) shear zone had a broadly listric form and that the hanging wall largely maintained this form during extension. The lower plate therefore must have undergone major flexural deformation as it was displaced up and out from beneath the upper plate. A significant conclusion of this study is that, during extension, the low flexural strength of the lower plate of the Buckskin-Rawhide detachment fault and a highly mobile middle to lower crust allowed rapid warping and short-wavelength isostatic adjustment during extension. We thus support earlier proposals for styles of extensional deformation in which a migrating monocline in the lower plate follows the tapered end of the upper plate during extension [Buck, 1988; Hamilton, 1980; Wernicke and Amos, 1983]. In addition, we support previously proposed models for a highly mobile deeper crust [Gans, 1987; Spencer and Reynolds, 1989b; Wernicke, 1989; Block and Royden, 1990] that accommodated the migrating monocline. Extension-parallel antiforms and synforms in the lower plates of the detachment faults, which are reflected by the physiography of the Buckskin, Rawhide, and Harcuvar mountains, are interpreted in part as folds that formed under conditions of low flexural strength.

GEOLOCIC SETTING

The southwestern part of the Basin and Range province can be divided into Tertiary tilt-block domains within which normal-fault bounded blocks are tilted in one dominant direction (figure 1) [Spencer and Reynolds, 1989b]. Tertiary mylonitic fabrics in metamorphic core complexes have been interpreted as exhausted ductile shear zones that formed down dip from detachment faults (shear zone model of metamorphic core complexes) [Wernicke, 1981; Davis, 1983; Davis et al., 1986]. The Tertiary age of mylonitic fabrics in many metamorphic core complexes in the southwestern United States is indicated by geochronologic data from the Whipple [Wright et al., 1986], Buckskin [Bryant and Wooden, 1989], South [Reynolds et al., 1986b], and Tortolita [Keith et al., 1980] mountains. The sense of shear during mylonitization, as interpreted from mylonitic petrofabric, was the same as the general sense of displacement on normal faults in the associated tilt-block domains [Davis et al., 1986; Reynolds and Lister, 1987; Spencer and Reynolds, 1989b]. This kinematic coordination supports the shear zone model for the origin of Tertiary mylonitic fabrics in Cordilleran metamorphic core
complexes, and all of our data from the study are consistent with this interpretation.

The Whipple tilt-block domain in western Arizona, southeastern California, and southernmost Nevada (Figure 1) [Howard and John, 1987; Spencer and Reynolds, 1989b] contains some of the most areally extensive exposures of Tertiary mylonite fabrics in North America [Davis et al., 1980; Davis and Lister, 1988; Spencer and Reynolds, 1989a]. The study area is a transect across the southern part of this tilt-block domain and includes areas at the ends of the transect that have undergone little or no Tertiary extension at exposed crustal levels.

The Transition Zone physiographic province of Arizona, which is at the northeastern end of the transect, separates the Basin and Range province from the Colorado Plateau. It is structurally continuous with the Colorado Plateau, but its Paleozoic and Mesozoic sedimentary cover has been largely removed by erosion. It has undergone little Tertiary extension at exposed structural levels. The southwestern regional slope of the Transition Zone reflects a decrease in crustal thickness from more than 40 km to the northeast to less than 30 km to the southwest [Warren, 1969; McCarthy et al., 1987; HAUSER et al., 1977; HAUSER and LUNDY, 1989]. Before mid-Tertiary extension, the area that is now the Transition Zone sloped northeastward and sediments were shed onto the Colorado Plateau from what is now the Basin and Range province [Young and McKee, 1978; REEVE et al., 1992]. The mid-Tertiary reversal of drainage and sediment-transport direction is interpreted as a reflection of crustal thinning of the Basin and Range province from thicknesses greater than to thicknesses less than those of the Colorado Plateau. The reversal in the surface slope of the area that is now the Transition Zone was presumably caused by southward displacement and thinning of lower and middle crustal rocks, with progressively greater crustal thinning toward the southwest [Wernicke, 1985; Reynolds and Spencer, 1985; Spencer and Reynolds, 1989b].

Ranges at the southwestern end of the transect have undergone little extension at exposed crustal levels. These ranges include the southern Pomoa [MILLER, 1970] and central Dome Rock (R. Tosdal, unpublished map, 1988) mountains in Arizona and the Big Maria [HAMILTON, 1987], Little Maria [EMERSON, 1982], and Paten [Stone and Kelley, 1989] mountains in California (Figure 2). This part of the Basin and Range province now has a crustal thickness of 22 to 28 km [HEARN, 1984; HAUSER et al., 1987] but almost certainly had a much greater crustal thickness immediately after Mesozoic crustal shortening and granitoid plutonism [e.g., HAMILTON, 1978].

**Structural Geology and Rock Types Along the Transect**

Structural styles of Miocene extension vary greatly across the southern part of the Whipple tilt-block domain (Figure 2). The southwestern part of the transect (Ranegrass Plain area) crosses an area where displacement on initially steeply to moderately northeast dipping normal faults was associated with substantial rotation of faults and fault blocks. The middle part of the transect crosses the northern part of the Harcuvar metamorphic core complex (Buckskin and Rawhide mountains). The northeastern part of the transect crosses the boundary between the Basin and Range and Transition Zone physiographic provinces (Artillery and Poachiie mountains).
Ranegas Plain Area

The southern Plomosa Mountains (southwest corner of Figure 3) consist largely of rocks that have not been significantly affected by Tertiary normal faulting [Miller, 1970; Sherrod and Koch, 1987]. These rocks are bounded to the northeast by a northeast-southeast normal fault (breakaway fault on Figure 3) that separates the largely unextended southern Plomosa Mountains from gently to steeply southwest tilted fault blocks bounded by normal faults in the Ranegas Plain area. Gently southwest dipping Tertiary mafic volcanic rocks that rest depositionally on pre-Tertiary rocks along the southwestern flank of the Little Harquahala Mountains [Spencer et al., 1985] are similar and probably coeval with southwest tilted mafic volcanic rocks in the New Water and central Plomosa Mountains [Sherrod et al., 1990]. Northeast dipping normal faults along the southwest side of Ranegas Plain are thus inferred to extend under the Little Harquahala and Granite Wash mountains.

Crystalline rocks that underlie the tilted Tertiary section in the central Plomosa Mountains continue unbroken for almost 20 km to the north where they form much of the northern Plomosa Mountains [Stoneham, 1985; Scarborough and Meader, 1989]. These crystalline rocks are bounded to the east and north by the Plomosa fault, a moderately to gently dipping normal fault that is overlain by an extended array of south to southwest tilted fault blocks. The Plomosa fault projects eastward beneath the western Bouse Hills, which consist of Proterozoic crystalline rocks that are depositionally overlain by moderately to steeply southwest dipping Tertiary volcanic and sedimentary rocks [Spencer and Reynolds, 1990a]. The tilted Miocene strata in the western Bouse Hills,

Fig. 2. Map showing locations of mountain ranges and regional structural plates. Plate 1 consists of rocks that form the footwall block of the southwesternmost northeast dipping normal fault (breakaway) within the Whipple tilt-block domain. Plate 2 consists of rocks that are structurally above the breakaway but are below the large detachment faults of the region. Rocks above the detachment faults comprise Plate 3 and extend northeastward to the Transition Zone physiographic province and Colorado Plateau. Some correlations are speculative; rocks shown as Plate 2 in California may actually be Plate 1 [e.g., Frost and Okaya, 1986].

Fig. 3. Simplified geologic map of the Ranegas Plain area showing location of transects for which the magnitudes of extension have been estimated (estimates in Table 1 and Appendix). Abbreviations are Qtv, undeformed Quaternary and upper Tertiary sedimentary and volcanic deposits; Tsv, middle Tertiary sedimentary and volcanic rocks; MzPsv, Mesozoic and Paleozoic sedimentary and volcanic rocks; and TRe, Tertiary to Proterozoic crystalline rocks.
including mafic volcanic rocks, are on strike with the tilted Miocene mafic volcanic rocks on the west flank of the Granite Wash and Little Harquahala mountains (Figure 3), and all three areas appear to be in the same basic mid-Tertiary structurally setting.

**Buckskin and Rawhide Mountains**

The undulating Buckskin-Rawhide detachment fault separates a variably mylonitic lower plate from highly faulted upper plate rocks that include Miocene volcanic and sedimentary rocks (Figure 4) [Rehrig and Reynolds, 1980; Spencer and Reynolds, 1989a]. The detachment fault has a corrugated form with corrugation axes trending approximately N55°E. Upper plate rocks are typically preserved in synforms, whereas resistant lower plate crystalline rocks make up the topographically higher antiforms.

Lower plate crystalline rocks in the Buckskin, Rawhide, and central and eastern Harcuvar mountains consist largely of amphibolite-grade gneisses intruded by granitic to dioritic sills and plutons of Proterozoic (?) to Tertiary age [Rehrig and Reynolds, 1980; Spencer and Reynolds, 1989a]. These

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**Fig. 4.** Simplified geologic map of the Rawhide and central and eastern Buckskin mountains, and adjacent areas.
crystalline rocks are overprinted by a Tertiary mylonitic fabric that grades from well developed and penetrative in northeastern areas to weak and restricted to discrete shear zones in southwestern areas. The degree of mylonitic fabric development generally decreases structurally downward.

In the Rawhide and northern Buckskin mountains, the corrugated form of the Buckskin-Rawhide detachment fault is well constrained by numerous exposures of the fault. Farther south, the form of detachment faults is revealed primarily by the geomorphology of resistant lower plate rocks (Figure 3) [Fatt, 1987]. Contour lines within the lower plate appear to be arched in the eastern Harcuvar and eastern Buckskin mountains, which suggests that the corrugations partially reflect folds in lower plate rocks. The approximate parallelism of corrugation axes and mylonitic lincations suggests that folding and extension were related.

The Harcuvar, Little Buckskin, and Buckskin mountains are separated by intervening Quaternary surficial deposits in Butler Valley (Figure 4). Varably mylonitic, lower plate crystalline rocks are essentially identical in these ranges and the detachment fault corrugations defined by physiography are all parallel (Figure 5). The detachment fault at the east end of the Harcuvar Mountains, known as the Bullard fault (Figure 2) [Reynolds and Spencer, 1985], thus appears to be correlative with the Buckskin-Rawhide detachment fault; both faults form the upward bounding surface of what is probably a single arsally extensive lower plate and both faults have the same sense and inferred age of displacement.

Lower plate rocks in the Buckskin and Rawhide mountains strongly resemble those in the eastern Whipple Mountains. Overlying detachment faults in the two areas have the same sense of displacement, and their upper plates appear to be part a single large area of tilt blocks and half grabens that includes the Bill Williams Mountains (Figure 2) [Sherrod, 1988; Spencer, 1989]. The two detachment faults are thus probably correlative [Davis et al., 1980; Spencer and Reynolds, 1990b].

Artillery and Poachie Mountains

The northeastern part of the transect includes the area from the northeasternmost trace of the Buckskin-Rawhide detachment fault to the Transition Zone physiographic province (Figure 6). The detachment fault in the northeastern Rawhide Mountains dips gently northeastward beneath southwest-tilted Tertiary strata that overlie crystalline rocks in the Artillery and Poachie mountains [Laskey and Wobber, 1949; Otten, 1982; Bryant, 1988; Spencer et al., 1989a]. Proterozoic crystalline rocks that underlie the tilted Tertiary strata include a variety of rock types, some of which form northeast trending belts that extend into the Transition Zone physiographic province of central Arizona [Bryant, 1988]. These belts are not significantly disrupted by Tertiary faults (locations A and B in Figure 6).

The base of mid-Tertiary strata in the Artillery Mountains and southern Poachie Mountains is tilted 30°-50° to the southwest. Farther northeast and east, similar-age strata are gently dipping and rest on an undulating subhorizontal to gently dipping disconformity [Bryant, 1988; Brooks, 1985] (volcanic rocks around location B in Figure 6). The magnitude of tilting of underlying pre-Tertiary rocks, which are not separated from the Transition Zone physiographic province by faults with large displacement, thus decreases to the northeast. Distributed brittle deformation within these

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Fig. 5. Minimum-elevation contour map of the lower plate of the Buckskin-Rawhide-Bullard detachment fault showing fault-surface corrugations with axes parallel to fault-displacement direction. Contours are drawn to intersect topographically highest areas such as ridge crests and therefore to display the minimum elevation of the detachment fault over the lower plate.
rocks appears to have at least partially accommodated progressively greater tilting toward the southwest.

**MAGNITUDE AND DIRECTION OF EXTENSION**

Various geologic relationships allow fairly well constrained estimates of extension along four transects in the southern part of the Whipple tiltblock domain (Figures 3 and 7). Each of these transects is divided into segments; detailed analysis of each segment is given in the Appendix, and estimates of extension are in Table 1. The following summarize the results of this analysis.

Transect 1 (Figure 3) crosses a belt of extension between the ranges directly southwest of the Whipple tilt-block domain and the lower plate rocks in the Harcuvar metamorphic core complex, whereas transects 2, 3, and 4 (Figure 7) cross an area of extension associated with movement on the Whipple-Buckskin-Rawhide-Harcuvar detachment system or within its upper plate. Lower plate rocks in the Harcuvar (excluding the Harquahala Mountains) and Whipple metamorphic core complexes can be treated as part of a single, structurally intact block that did not undergo significant dismemberment by normal faulting. The areally extensive exposures of crystalline rocks that make up the Transition Zone physiographic province northeast of the study area can also be treated as a single, unextended fault block. The Tote Creek, Poachie, and Artillery mountains represent the tapered end of a largely unextended Transition Zone block. Thus estimates of extension along Transects 2, 3, and 4 should all be similar.

The direction of extension in the Harcuvar and Whipple metamorphic core complex is indicated by the N55°E±3° orientation of detachment-fault corrugation axes (Figure 5) and
the N50°E±10° orientation of mylonitic lineations (average mylonitic-lineation trend from each of 10 studies falls within this range). Detachment-fault corrugation axes have a more narrowly constrained orientation than mylonitic lineations and are probably a better indicator of extension direction. Displacement oblique to the larger corrugations seems unlikely because of the large flexural strains that upper plate rocks would have to undergo during movement.

Ranegras Plain Transect (Transect 1)

The direction of extension in the Ranegras Plain area is suggested by several features: (1) dip directions of tilt blocks, (2) slickenside striations on the Plomosa fault, and (3) mylonitic lineations in the footwall of the Plomosa fault. These features are all consistent with a N45°E±15° direction of extension.

In the northwestern Ranegras Plain area, total extension is estimated to be 23±8 km, which is the sum of estimates of extension along three segments of a transect from the southern Plomosa Mountains to the house rise (Figure 1A, 1B, and 1C in Figure 3, Table 1, and Appendix). Most of this extension occurred by movement on the Plomosa fault, and is apparent in part because upper plate rocks include highly tectonized Paleozoic and Mesozoic strata (location B in Figure 3) that were displaced approximately 13 km to the


**TABLE 1. Estimates of Extension**

<table>
<thead>
<tr>
<th>Transect</th>
<th>Segment</th>
<th>Segment Length (km)</th>
<th>Magnitude of Extension (km)</th>
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<tbody>
<tr>
<td>1 (Plomosa)</td>
<td>A (S. Plomosa)</td>
<td>11</td>
<td>6.0±4</td>
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<td></td>
<td>B (Cen. Plomosa)</td>
<td>13</td>
<td>13±2</td>
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<tr>
<td></td>
<td>C (N. Plomosa)</td>
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<td>4±2</td>
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<td></td>
<td>TOTAL</td>
<td>32</td>
<td>23±8</td>
</tr>
<tr>
<td></td>
<td>D (Ranegras)</td>
<td>25</td>
<td>&lt;25</td>
</tr>
<tr>
<td>2 (Buckskin)</td>
<td>A (Bouse Hills)</td>
<td>16</td>
<td>15±1</td>
</tr>
<tr>
<td></td>
<td>B (Buckskin)</td>
<td>34</td>
<td>33±1</td>
</tr>
<tr>
<td></td>
<td>C (Date Creek)</td>
<td>24</td>
<td>18±6</td>
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<tr>
<td></td>
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<tr>
<td>3 (Harquvar)</td>
<td>A (Harquvar)</td>
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<td>55±10</td>
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<tr>
<td></td>
<td>B (Date Creek)</td>
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<td>4 (Whipple)</td>
<td>A (Whipple)</td>
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<td></td>
<td>B (Crossman)</td>
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<td>8±3</td>
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<td></td>
<td>C (Sacramento)</td>
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<td>12±8</td>
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<td></td>
<td>D (Hualapai)</td>
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<td></td>
<td>F (Big Sandy)</td>
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<td>6±3</td>
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<tr>
<td></td>
<td>TOTAL</td>
<td>105</td>
<td>71±19</td>
</tr>
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</table>

*Represents distance between endpoints of transect.

northeast from their lower plate equivalents in a Mesozoic thrust zone in the Plomosa Pass area (location A in Figure 3). In another transect to the southeast (segment 1D), extension must have been less than about 28 km because starvation of more than 25 km of extension along northeast dipping normal faults places weakly to moderately metamorphosed and deformed Paleozoic and Mesozoic strata in the core of the Granite Wash Mountains [Reynolds et al., 1989] on top of even less metamorphosed and deformed Paleozoic and Mesozoic strata in the New Water Mountains [Sherrold and Koch, 1987]. Although thrust juxtapositions of higher grade Mesozoic metamorphic supracrustal rocks over lower grade rocks are possible, such juxtapositions of Mesozoic strata have not been recognized in the region [Reynolds et al., 1986c, 1988]. Our estimate of 20±5 km of total Tertiary extension in the Ranegras Plain area represents the range of extension encompassed by estimates from both transects 1A-1C and 1D.

The tilted Tertiary strata on the west side of the Bouse Hills represent the tilted and eroded footwall of the breakaway zone for the Buckskin-Rawhide detachment fault. Equivalent hanging wall rocks are now in the subsurface beneath Date Creek basin east of the eastern Buckskin Mountains and were originally separated from lower plate rocks in the Bouse Hills only by strata that are now preserved in klippen or were eroded away. South to southwest dipping Tertiary strata in the Artillery and Poachi mountains project down to the southern boundary of the subsurface beneath Date Creek basin and are inferred to be unstruck in the subsurface by the northeast dipping Buckskin-Rawhide detachment fault. Total displacement of Tertiary strata and underlying crystalline rocks, originally close to each other in the breakaway zone, is estimated to be 60±5 km (sum of displacements along transect segments 2A, 2B, and 2C in Figure 7, Table 1, and the Appendix).

**Harquvar Transect (Transect 3)**

A southwest dipping sequence of Miocene volcanic and sedimentary rocks in the Bullard Peak area on the south flank of the eastern Harquvar Mountains contains a conglomerate and a breccia unit that is interpreted to have been displaced 55±10 km northeastward relative to lower plate rocks (transect segment 3A, Figure 7; Table 1; Appendix) [Reynolds and Spencer, 1983]. The crystalline rocks of the eastern Date Creek Mountains represent the edge of the Transition Zone physiographic province and are approximately 25 km northeast of the Bullard Peak area. We estimate that 12±1 km of extension has occurred within this transect segment (transect 3B), for a total of 67±17 km of extension along transect 3.

**Whipple Transect**

The major offset feature in the Whipple-Mohave mountains region to the northwest of the Harquvar metamorphic core complex is a dike swarm in the lower plate of the Whipple detachment fault (Chambers Well dike swarm of Davis et al. [1980, 1982]) that is interpreted to be represented in the upper plate by a dike swarm in the Mohave Mountains (transect segment 4A in Figure 7) [Nakata, 1982; Howard et al., 1982]. Restoration of 43 km of displacement along a N55°E transport vector is necessary to bring the easternmost exposures of the Mohave Mountains dike swarm above the easternmost exposures of the Chambers Well dike swarm (Table 1; Appendix; see also Davis and Lister [1988]). The N55°E direction is based on the orientation of axes of corrugations in the Whipple Mountains [Poulton, 1961]. If displacement was more northerly, as suggested by the N45°E±10° mylonitic lineations in the Whipple Mountains [Davis and Lister, 1988], slightly greater minimum displacement would be indicated. Additional extension occurred at structurally higher levels in the Mohave Mountains [Howard et al., 1982] and areas farther east (Figure 7; Table 1; Appendix). Total extension along this transect is estimated at 71±10 km.

**Summary**

Of the three estimates of total extension between the Transition Zone and the lower plate rocks on the western flanks of the Harquvar and Whipple metamorphic core...
complexes, the estimate for the Buckskin transect (66±8 km) is the best constrained and is entirely within the uncertainties in estimated extension for the other two transects. An additional 20±5 km of extension occurred in the Ranegras Plain area. If this amount is added to the estimate for extension along the Buckskin transect, extension between the southern Pima Mountains and the Transition Zone totals 86±13 km. The belt of extension through the Ranegras Plain area may die out northwestward west of the Turtle and Whipple mountains. If it does, then the unextended area along the southwest ramp of the wrappie tilt-block domain must have undergone 10°-15° of clockwise rotation.

TECTONIC RECONSTRUCTION

Major pre-Tertiary geologic features in west central Arizona and adjacent southeastern California include the Maria fold and thrust belt, which is an east-west trending belt of generally south-vergent folds and thrusts [Reynolds et al., 1986; Laubach et al., 1989; Spencer and Reynolds, 1990b], and the thick sequence of Jurassic (?) and Cretaceous clastic sedimentary rocks (McCoy Mountains Formation of Harding and Coney [1985]; see also Stone et al. [1987]) that form the structurally lowest levels of the Maria fold and thrust belt. Restoration of Tertiary extension in the Ranegras Plain area brings exposures of the McCoy Mountains Formation in the Granite Wash Mountains [Reynolds et al., 1989] to a position adjacent to similar rocks in the New Water Mountains [Sherrod and Koch, 1987], and thus increases the linearity of the belt of outcrops of the McCoy Mountains Formation (compare Figures 8 and 9). The reconstructed Maria fold and thrust belt (Figure 9) is more linear and narrow than it is today (Figure 8) partly because thrust-fault segments in Tertiary fault blocks that are now above the Buckskin-Rawhide detachment fault were restored to their pre-Tertiary position [Spencer and Reynolds, 1990b]. In addition, the mylonitic lower plate of the Harcuvar metamorphic core complex is restored to beneath the Transition Zone or beneath the slightly faulted rocks of the eastern Pocah Mountains (see also Reynolds and Spencer [1990] and Weddle [1985]).

CROSS-SECTION EVOLUTION

Initial Shear Zone Geometry

Southwest tilted Tertiary strata in the western Bouse Hills are interpreted as the tilted footwall block of the Buckskin-Rawhide detachment fault adjacent to the point where the fault initially reached the Earth's surface (initial breakaway region). Lower plate rocks in the Buckskin Mountains contain Tertiary mylonitic fabrics approximately 16 km to the east of the tilted strata. The initial Buckskin-Rawhide detachment fault must have cut downward from the Earth's surface to depths of mylonitization over this 16 km distance, which is measured across the now rotated and flattened lower plate but represents original downdip distance along the fault.

Tertiary mylonitic fabrics below the Buckskin-Rawhide detachment fault are not restricted to areas of Tertiary

![Map showing major tectonic features of study area. Dashed line with bar and ball is southwesternmost normal fault of Whipple tilt-block domain. Restoration of Cenozoic extension shown in Figure 9 was done by restoring 66 km of N55°E extension (represented by three long arrows) between the Transition Zone and mylonitic rocks below detachment faults. All structures and rocks that are to the southwest of the mylonitic rocks in Figure 9 were also restored this amount except for two segments of thrust faults in the upper plate of the detachment fault in the Buckskin and Rawhide mountains which were restored smaller amounts relative to the Transition Zone. Additional restoration of rocks southwest of the normal fault that marks the southwest boundary of the Whipple tilt-block domain was done by moving the southern area 20 km to the northeast (short arrow) with rotation about point A.](image)
magnetism but must have formed at depths where regional temperatures were sufficient for plastic deformation of quartz. Lower plate rocks in the Buckskin and Rawhide mountains have consistently yielded Miocene K-Ar and 40Ar/39Ar dates on biotite [Spencer et al., 1989b; Richard et al., 1990] and Miocene fission track dates on apatite and zircon [Bryan and Naeser, 1987]. The association of young thermochronometric dates with mylonitic rocks suggests that mylonitization occurred at temperatures above approximately 300°C. The 300°C isotherm would be at a depth of approximately 15 km in a low geothermal gradient of 20°C/km and at approximately 6 km depth in a high geothermal gradient of 50°C/km. For a planar fault to reach these depths over a down dip distance of 16 km would require fault dips of 65° to 22°. Average dip of the initial Buckskin-Rawhide detachment fault was probably within this range.

The lack of major changes in rock type or metamorphic grade from southwest to northeast across the mylonitic lower plate in the Buckskin and Rawhide mountains suggests that the ductile shear zone that was the precursor to the Buckskin-Rawhide detachment fault had a gentle dip. If the shear zone had a steep initial dip, lower crustal granulites or even mantle peridotites would presumably be exposed in the eastern part of the lower plate. The increasing intensity of mylonitization toward the northeast is the only presently known lithologic indication of greater structural depths toward the northeast. Late Cretaceous and early Tertiary K-Ar and 40Ar/39Ar hornblende dates from the Rawhide and eastern Harcuvar mountains indicate that the eastern part of the lower plate was above the 450°C-500°C isotherm during the mid-Tertiary [Shackelford, 1980; DeWitt and Reynolds, 1990; Richard et al., 1990]. A maximum temperature difference at opposite ends of the Buckskin-Rawhide mountains of 150-200°C within a geothermal gradient of 20°-30°C/km corresponds to a 3-10 km maximum differential in pressure and temperature across the ranges and indicates that initial, average, shear zone dip was not more than about 16°.

The initial cross-sectional form of the Buckskin-Rawhide detachment fault and its down dip continuation as a ductile shear zone is thus constrained by the following: (1) the average dip of the brittle, upper crustal part of the shear zone was between approximately 22° to 65°, and (2) below the brittle-ductile transition the ductile shear zone dipped less than about 16°. A broadly listric shear zone form seems virtually certain (Figure 10). These geometric constraints allow a fault dip of greater than 30° where the fault intersects the brittle-ductile transition, which is consistent with seismological studies of normal-fault dips at depths of earthquake nucleation [Jackson, 1987].

An alternative interpretation is that the lower plate was derived from a reservoir of highly volatile deep crust, and exhumation of low-viscosity crust from this reservoir was accommodated above by slip on a moderately to steeply dipping mylonitic shear zone. Major penetrative strain would accompany such an exhumation process. Leucocratic granitoid sills are present in many areas of the Harcuvar metamorphic core complex, including a single sill-like intrusion that extends over much of the central and eastern Harcuvar Mountains [Drowes et al., 1990]. In the Harcuvar Mountains, the granitoids are interpreted as a single comagmatic intrusive suite that includes the 78 Ma [DeWitt and Reynolds, 1990] Tank Pass granite in the western Harcuvar and Granite Wash mountains. Host rocks have yielded pre-Tertiary hornblende 40Ar/39Ar plateau dates [DeWitt and Reynolds, 1990]. Unlike their highly foliated Proterozoic host rocks, the granitoids in the central and eastern Harcuvar Mountains are typically unoriented except for mylonitization at their upper and lower margins. The lack of evidence of penetrative deformation in these Cretaceous granitoids and their moderate mid-Tertiary temperature probably precludes this alternative exhumation model.

The initial cross-section geometry of the two major normal shear zones in the Plomosa Mountains (Plomosa fault and breakaway fault in Figure 3) is not as well constrained as for the Buckskin-Rawhide detachment fault. The 45°-60° southwestward dip of Tertiary strata in the Plomosa Pass area [Stoner, 1985] suggests that the Plomosa fault and breakaway fault have been similarly rotated and thus formed as high-angle normal faults. Initially, the two major normal faults in the Plomosa Mountains could have been
approximately parallel to the listric Buckskin-Rawhide detachment zone (Figure 10), but it is also possible that they had significantly different geometry, either dipping more shallowly and merging with the mylonite zone exposed in the Buckskin and Rawhide mountains or dipping more steeply and diverging with depth from the Buckskin-Rawhide detachment zone.

**Incision and Excision**

Detailed geologic studies in the nearby Whipple Mountains and ranges farther north have led to recognition of two processes that played a significant role in the evolution of some detachment faults. Incision is the process whereby a new splay of a detachment fault cuts downward through the lower plate and transfers a slab of lower plate rock to the upper plate. Excision is the process whereby a new splay of a detachment fault cuts upward through the upper plate and transfers a slab of upper plate rocks to the lower plate [Davis and Lister, 1988; see also John, 1987]. In both processes, the old splay of the detachment fault becomes largely or entirely inactive.

Although both processes were inferred to be important in the Tertiary structural evolution of the Whipple Mountains, we infer that incision was not important in the Harcuvar complex.

Fault blocks and slivers incised from the lower plate should be present above the Buckskin-Rawhide detachment fault if incision had been an important process during detachment faulting. Lower angle faults in the lower plate in the eastern Buckskin and Rawhide mountains are discontinuous, do not juxtapose highly contrasting rock types, and do not have significant displacement compared to the detachment fault [Shackelford, 1989a; Bryant and Wooden, 1989]. Similar lower plate faults have been recognized in the Whipple Mountains [Davis et al., 1983]. The Buckskin-Rawhide detachment fault is discontinuously exposed for approximately 24 km in the direction of displacement, yet upper plate fault blocks and slivers of lower plate rock types have not been recognized. Upper plate crystalline rocks in the Buckskin and Rawhide mountains are composed primarily of reddish brown granite-granodiorite rocks that do not resemble the commonly epidote- and chlorite-bearing, gneissic and mylonitic lower plate rocks. In addition, the upper plate crystalline rocks commonly are depositionally overlain by sedimentary and volcanic rocks deposited during early detachment faulting [Spencer and Reynolds, 1989c]. These upper plate crystalline rocks were thus at the Earth's surface early in the history of detachment faulting and represent shallow crustal levels characteristic of the top of the upper plate rather than the mid-crustal levels where lower plate rocks underwent plastic deformation and, during uplift, retrograde metamorphism. Incision was thus not an important process during much or all of the history of detachment faulting in the Buckskin and Rawhide mountains. In contrast to Davis and Lister's [1988] interpretation that the Whipple detachment fault is only the youngest manifestation of a complex, evolving detachment system, the Buckskin-Rawhide detachment fault was a long-lived feature, during extensional faulting, with respect to the lower plate.

Upper plate normal faults that are truncated downward by the Buckskin-Rawhide and Whipple detachment faults indicate that excision was an important process in both areas [Davis and Lister, 1988; Lister and Davis, 1989; Spencer and Reynolds, 1989c]. The excised slabs that contain the presumably listric, downdip continuations of the upper plate normal faults are not presently exposed and are presumably in the subsurface to the southwest. Stepwise migration of the breakaway fault to previously active upper plate normal faults is similar to incision because it transfers upper plate rocks to the lower plate, but unlike incision, it is not
accompounded by formation of new faults. In both processes, upper plate fault blocks can be left stranded on a progressively lengthening, inactive segment of the detachment fault [Buck, 1988; Hamilton, 1988; Wernicke and Axen, 1988].

The hanging walls of listric or low-angle normal faults within the upper plates of detachment faults may be displaced down and onto the footwalls of detachment faults. Such a process could occur during extension within the upper plate or during stepwise breakaway migration. In the Buckskin and Rawhide mountains, much of the upper plate is composed of synextensional sedimentary and volcanic rocks that did not even exist when detachment faulting began. The detachment fault, with respect to the upper plate, is thus a young structural feature in the history of detachment faulting. This is similar to the interpretation of the Whipple detachment fault by Davis and Lister [1988], although they emphasize low-angle excision, rather than extension of the upper plate and stepwise breakaway migration, as the primary causative process.

The process of excision appears to have affected only the tapered end of the upper plate, and to have affected primarily Tertiary strata deposited in an extensional basin above this tapered end. In other words, only a small but now well-exposed volume of the total upper plate was affected by excision tectonics. It is possible that a larger volume of rock, including a substantial component of crystalline rock, was transferred from the upper plate to the lower plate near the breakaway but is now concealed by surficial deposits southwest of the Buckskin Mountains [Spencer and Reynolds, 1989c], but there is no trace of such an excised sliver in the mouse mires. We conclude that neither incision nor excision was very significant in terms of volume of rock affected, and these processes can be largely ignored in constructing regional evolutionary cross sections of the Buckskin-Rawhide detachment system (Figure 10).

Monoclinal Flexure

The lack of significant extension within the Artillery and Pasoche mountains and the absence of an extensive array of crystalline fault blocks above the Buckskin-Rawhide detachment fault, especially in the eastern and central Buckskin Mountains, indicate that the upper plate did not undergo major changes in form during displacement. The only significant modification occurred at the tapered end of the upper plate which was affected by southwest tilting and, for a fairly small volume of rock, extreme extension. Displacement of the largely rigid upper plate of the Buckskin-Rawhide detachment shear zone, with its broadly listric underside, thus required major flexural deformation of the lower plate (Figure 10). A monoclinal flexure of the lower plate must have followed the tapered end of the upper plate as the upper plate was displaced (relatively) to the northeast. Following uplift and flexural deformation, the lower plate formed a regionally subhorizontal, generally low-relief surface. This migrating monoclinal style of deformation has been proposed elsewhere [Buck, 1988; Hamilton, 1988; Wernicke and Axen, 1988].

Timing of Extension

The Plomosa fault and the structurally lower normal fault that underlies all of the northern Plomosa Mountains were rotated to gentle dips and probably became inactive before much or most of the movement on the Buckskin-Rawhide detachment fault. Basin formation had begun in the northern Plomosa Mountains, Bouse Hills, Buckskin Mountains, and the Artillery Mountains by about 24 Ma [K-Ar data from Eberly and Stanley [1978] and R. Miller, written communication, 1987, 1990], probably as a result of initial movement on the Plomosa and Buckskin-Rawhide detachment faults. Tilting had largely ended in the Bouse Hills by the time 20 Ma volcanics were erupted [Eberly and Stanley, 1978; Spencer and Reynolds, 1990a], but continued in the southernmost Bill Williams and, presumably, Buckskin and Rawhide mountains until after deposition of a 16 Ma basalt [Spencer et al., 1989b]. Lower plate rocks in the Buckskin Mountains cooled through the approximately 300°C argon blocking temperature of biotite at 13 to 13 Ma [Spencer et al., 1989b; Richard et al., 1990] and were presumably undergoing rapid uplift and denudation at this time. On the basis of these data, we infer that extensional faulting began at about 23–25 Ma in the Plomosa Mountains and Bouse Hills and on the Buckskin-Rawhide detachment system, ended in the Plomosa Mountains and Bouse Hills by about 20 Ma, and continued until after 15 Ma in the Buckskin and Rawhide mountains. Approximately 80 to 90 km of extension occurred at an average rate of 8 to 9 mm per year over the approximately 10 m.y. period of extensional deformation in the study area.

The Plomosa detachment fault and structurally lower faults, and their down dip continuations as dextral shear zones, have been warped and rotated in part by isostatic uplift associated with displacement on the structurally higher Buckskin-Rawhide and correlative detachment faults. Warping and rotation may have been so severe that the normal shear zones were tilted over large areas back toward their breakaways and their movement was terminated. Sequential termination of movement on imbricate detachment faults, with structurally highest faults remaining active longest, is thus indicated.

Postdetachment Geomorphology

Basaltic volcanic rocks and locally occurring felsic volcanic and clastic sedimentary rocks, dated at 9-15 Ma and deposited in and around the northern Rawhide and western Buckskin mountains, were largely buried and partially preserved the immediately postdetachment landscape (Figure 11). The crests of lower plate anticlines are commonly at greater elevations than the base of preserved postdetachment strata. This is not due to warping after deposition of postdetachment strata because these strata are not warped, but was an original feature of the immediately postdetachment landscape that is consistent with evidence for subaerial exposure of the lower plate late during detachment faulting. The level of structural exposure represented by the surface upon which Miocene postdetachment strata were deposited is similar to present levels of exposure, a conclusion reached earlier by Davis et al. [1980, p. 122].

Elevations in the Whipple-Buckskin-Rawhide region range between about 200 and 600 m above sea level over an area of greater than 2000 km² and are similar to those of surrounding parts of the Basin and Range province. In contrast to tectonically active parts of the Basin and Range province, the geomorphic maturity of the study area is indicated by gentle, smooth slopes of alluvial fans, pediments above some granitoid rocks, and the absence of internal drainage. Incised drainages are
Simplified geologic map of the central and western Buckskin Mountains and adjacent areas showing smoothed contours of the minimum elevation of the surface upon which postdetachment basalts and clastic sediments were deposited. Contours intersect topographically highest areas such as ridge tops and therefore display the minimum elevation of the postdetachment surface. K-Ar dates indicate the age of postdetachment deposits (9-15 Ma), tilted upper plate volcanic rocks (one date at 16 Ma), and cooling of lower plate crystalline rocks (two biotite dates at approximately 15 Ma). K-Ar data are from Armstrong et al., [1976], Suneson and Lucchita [1979], Shackelford [1988], Davis et al., [1992], Reynolds et al., [1986a], and Spencer et al., [1998b]. The 15.1 Ma date from the western edge of the map is a weighted mean average of three dates (see Long and Rippeteau [1974] for weighting method).

SPECCULATIONS ON CRUSTAL MECHANICS

Migrating Monocline and Flexural Strength

The migrating monocline style of deformation in the lower plate requires that the lower plate have low flexural strength. If the lower plate had a teetsonically significant flexural strength, it would first have resisted concave upward bending as it was drawn upward beneath the upper plate, and then would have resisted concave downward bending and later flattening to its present subhorizontal form. The result of significant resistance to bending would be to retard lower plate uplift, thereby forming a deep basin at the tapered end of the upper plate that could be fitted by sedimentation or by a distorted fault-block array derived from the upper plate. However, the thickness of upper plate sediments and distended fault blocks in the Buckskin and Rawhide mountains is trivial compared to the magnitude of footwall upur. Significant resistance to bending of the lower plate in the breakaway region would tend to preserve a large mountain front adjacent to the trace of the breakaway fault, but a mountain front at the breakaway zone has not been preserved. The presence of abundant catastrophic debris-avalanche deposits in the lower parts of many Tertiary sedimentary sequences in the study area is probably due to substantial relief produced during early normal faulting and may reflect high lower plate flexural strength that characterized only the earliest phase of extension.
The low flexural strength of the lower plate probably had several causes. High lower plate temperatures during uplift would result in low flexural strength by allowing ductile creep at shallow depths [Brace and Kohlstedt, 1980]. Mylonitic lower plate rocks were initially at midcrustal temperatures and would retain some of their heat during uplift, especially if uplift was rapid [England and Jackson, 1987]. Magmatism further elevated temperatures and reduced flexural strength. Flexural slip along weak zones within lithologically layered lower plate rocks may have accommodated flexure and further reduced flexural strength. Flexural strength reduction by these mechanisms is supported by Miocene K-Ar, 4Ar/39Ar, and fission track cooling ages from the lower plate, and by the presence of mid-Tertiary intrusions and subhorizontal faults within the lower plate [Shackelford, 1989; Bryant and Wobrock, 1989; Spencer, 1989]. Flexural slip on normal faults, exposed in the Plumas Mountains, that dip beneath the Buckskin-Rawhide detachment fault possibly reduced the flexural strength of the lower plate of the Buckskin-Rawhide detachment fault in its breakaway region. Such a strength reduction may have been necessary for complete denudation in the breakaway region; upper plate synformal keels and sedimentary basins cover lower plates in most other breakaway regions [e.g., Spencer, 1984; John, 1987] where lower plates may not have been underlain by deeper active normal faults that could have accommodated flexural slip.

The migrating monocline style of deformation depicted in Figure 10 requires not only low flexural strength of the footwall, but also that the deeper middle to lower crust behaved as a low-viscosity fluid that flowed into the area beneath the migrating monocline and mechanically decoupled the upper crust from the stronger mantle lithosphere [Block and Rydter, 1990]. The low viscosity of the deeper crust either was caused by heating of pre-existing crustal rocks by mid-Tertiary magmas [e.g., Gans, 1987; Spencer and Reynolds, 1989b] or reflects voluminous magmatic underplating and deep intrusion [Thompson and McCarthy, 1990].

Regional Elevations and Isostasy

The areally extensive Tertiary mylonitic rocks in the Harcuvar and Whipple metamorphic core complexes overlie the root zone of thrusts in the east-west trending Cretaceous Maria fold and thrust belt to the south. This spatial relationship suggests that isostatic uplift of the buoyant crustal root (downwarping protruding Moho bulge) of the Maria fold and thrust belt was responsible for the widespread exposures of Tertiary mylonitic rocks [Spencer and Reynolds, 1990b]. The apparent influence of a crustal root on vertical movements in the upper crust requires that deep-crustal viscosity be high enough to support tectonically significant horizontal pressure gradients throughout most or all of the 10 m.y. period of detachment faulting and magmatism. If lower crustal viscosity was too low, uplift of a Moho root would cause rapid lateral flow in the deep crust with rapidly diminishing influence on the upper crust.

In contrast to the above inference of a tectonically significant, moderate deep-crustal viscosity, migration of short-wavelength monoclinal flexures of lower plates requires a highly mobile, low-viscosity substrate. These contrasting inferences about deep crustal behavior may be consistent with each other because of the different length scales of deep-crustal flow for each process. The low viscosity of the deep crust during extension may have allowed rapid isostatic equilibration at horizontal length scales of 5 to 15 km, but this viscosity may have been sufficient to support regional horizontal pressure gradients that were significant at length scales of 20 km. The viability of this mechanism is supported by numerical analysis that indicates that the time period for equilibration of horizontal pressure gradients by horizontal channelized flow is proportional to the square of the length scale of equilibration [Krusse et al., 1991].

Detachment-Fault Corrugations

Extension-parallel corrugations of detachment faults have been recognized for many years [e.g., Rehrig and Reynolds, 1980], but there has been controversy regarding their origin. Corrugations above largely nonmylonitic crystalline rocks farther north in the lower Colorado River trough have been interpreted as reflecting original fault geometry and not folding of a planar fault [Spencer, 1983; John, 1987]. This interpretation is based on part on the irregular, nonmonoclinal form of the detachment faults in these areas. Field relationships in the Whipple Mountains suggest that corrugations of the detachment fault formed after arching of lower plate mylonitic foliation and lithologic layering and that the corrugated form of the detachment fault was somewhat discordantly superimposed upon the arched lower plate fabric but crudely mimics its form [Davis and Lister, 1988].

The arched form of Cretaceous stuff in the eastern Harcuvar and eastern Buckskin mountains, which is mimicked by the form of mylonitic foliation and the detachment faults, may be the result of folding during the period of rapid lower plate uplift and low flexural strength. Folding apparently occurred without corresponding folding of adjacent nonmylonitic lower plate rocks in the Granite Wash and western Harcuvar mountains [Reynolds et al., 1989]. The Granite Wash and western Harcuvar mountains had greater flexural strength during mid-Tertiary extension because they had cooled through K-Ar mica blocking temperatures tens of millions of years before extension [Rehrig and Reynolds, 1980] and were thus within the strong, brittle upper crust at the time of extension.

Folding of only the initially deep, low flexural strength part of the lower plate can be accounted for by the following model: The isostatic normal fault that became the Buckskin-Rawhide-Bullard detachment fault system had an original trace at the Earth's surface that was characterized by segmentation and irregularity, as do modern normal faults [e.g., Jackson and White, 1989]. The irregular fault passed downward through the brittle-ductile transition into a more planar, gently dipping ductile shear zone. As ductile middle-crustal rocks, initially beneath or within a planar ductile shear zone, were displaced from beneath the upper plate, they deformed to conform to the irregularities at shallower crustal levels on the underside of the relatively rigid upper plate. Neither the upper plate nor shallow lower plate rocks were folded, but lower plate rocks at depths and temperatures sufficient for mylonitization were folded to conform to irregularities in the fault surface beneath the tapered end of the upper plate. Thus the corrugations of mylonitic lower plate rocks reflect the type of irregularities and segmentation that characterize active normal faults [John, 1987; Jackson and
White, 1989), but the corrugations are folds that formed during extension and are not original features of the lower plate. Folding was not accomplished by shortening perpendicular to fold axes; rather, it represents an increase in the surface area of the top of the lower plate as it was stretched to conform to an irregular surface.

An implication of this model is that the lower plate folds are a kind of tape recorder of extension direction. As each part of the ductile lower plate is drawn up beneath the grooved, tapered end of the upper plate, the extension direction is recorded by the orientation of fold axes in that part of the lower plate as it cools and becomes stronger. If this interpretation is correct, the NS5°E±3° orientation of corrugation axes in the Harcourt metamorphic core complex accurately represents the direction of extension. The greater range of mylonitic lineation orientations (NS0°E±10°) is possibly due to nonrigid ductile deformation before or during folding that reflects the influence of other unknown processes.

CONCLUSION

Miocene extension along a transect through the Whipple tilt-block domain in west central Arizona was accommodated at upper crustal levels by movement on three imbricate normal faults, the structurally highest of which evolved into the Buckskin-Rawhide detachment fault. Most of the extension was accommodated by movement on the Buckskin-Rawhide detachment fault, and this movement uncovered extensive areas of crystalline rocks with Tertiary mylonitic fabrics. Estimates of displacement on the three faults indicate that an area at the Earth’s surface that was originally only 10 to 20 km wide is now approximately 100 km wide. Restoration of fault movement realigns Mesozoic tectonic features.

The hanging wall of the Buckskin-Rawhide detachment fault and its downdip continuation as a ductile shear zone had an original, broadly listric form that does not appear to have changed significantly during extension. Various types of constraints on the cross-sectional evolution of the Buckskin-Rawhide detachment fault require that, as the lower plate was displaced beneath the upper plate, it flexed to conform to the curved underside of relatively rigid upper plate and flattened at near-surface levels as it was uncovered. The low regional surface relief of the denuded lower plate apparently reflects short-wavelength isostatic equilibration that occurred during extension. Folds below the Buckskin-Rawhide-Bullard detachment fault system have axes parallel to extension direction and are interpreted to have formed during extension as weak lower plate rocks conformed to the irregularities along the tapered end of the strong upper plate.

The inferred structural evolution of the study area requires very low flexural strength of the lower plates of detachment faults and low viscosity of the deeper crust during extension. Such crustal mechanical characteristics may be restricted to areas of high heat flow. Thrust belts, in contrast, are typically supported by the substantial flexural strength of underthrust lithosphere (e.g., McNutt et al., 1988). Horst and graben topography with large range fronts and deep basins may characterize areas of extension where flexural strength and deep-crustal viscosity were greater than they were during mid-Tertiary extension in west central Arizona. Even in west central Arizona, the deep crust apparently had sufficient viscosity to support horizontal pressure gradients over distances of 50-100 km that promoted regional uplift of overlying metamorphic core complexes. We conclude, however, that structural styles of Miocene extension in west central Arizona may be an example of extensional deformation in a crustal mechanical environment of unusually low crustal flexural strength and low-viscosity deep crust.

SEGMENT A: MAMMOTH ANGUS AND EXTENSION

Ranger Plain Transect (Transect 1, Figure 3)

Segment 1A. Strata at the base of the tilted Tertiary section in the lower part of the transect show an ENE-SW to NE-SE to southeasterly project downdip into a northeast-dipping normal fault. The exposed basal depositional contact is now approximately 11 km from the buried trace of the underlying fault to the southwest. A major, poorly constrained variable relevant to extension estimates concerns the amount of concealed pre-Tertiary rock beneath volcanic rock in tilt blocks along this 11-km-wide transect segment. If there is much concealed pre-Tertiary rock in tilt blocks, extension may not have been great. The absence of a negative residual Bouguer gravity anomaly over the half graben [Lyon et al., 1980], which contains primarily exposed volcanic and clastic sediments at exposed levels, suggests that underlying crystalline rocks are not at great depth, but it is not known if underlying crystalline rocks are below or above the master normal fault. In addition, the location of the buried trace of the breakaway fault is not well constrained but is projected into the transect from the southeast where it is exposed. On the basis of these considerations, we estimate 0.2-1.0 km total extension for this segment. The estimated range of 0.2 to 1.0 km of extension encompasses virtually all of the uncertainties outlined above.

Segment 1B. Two fault blocks above the Plomosa detachment fault contain multiply deformed and metamorphosed Paleozoic and Mesozoic strata (locations B in Figure 3) that strongly resemble tectonized rocks along a steeply dipping thrust zone in the Plomosa Pass area (location A in Figure 3) [Stoneham, 1985; Scarbrough and Meader, 1989; Duncan, 1990]. The 13 km distance between the northern and the two fault blocks and the thrust fault itself projected laterally to the transect is interpreted to represent normal displacement on the Plomosa detachment fault. Uncertainties in projecting the thrust along alluvium lead to an estimated magnitude of extension along this segment of the transect of 0.5±0.5 km.

Segment 1C. Normal faults exposed in the northern Plomosa Mountain [Scarbrough and Meader, 1989; Plomosa, 1990] that project beneath the western Boise Hills are estimated to have accumulated 4.2±2 km of extension.

Segment 1D. See text.

Buckskin Transect (Transect 2, Figure 7)

Segment 2A. The base of the steeply tilted Tertiary section in the western Boise Hills (location C in Figure 3) is an eroded segment of the breakaway of the Buckskin-Rawhide detachment fault. Mylonitic crystalline rocks in the southern Buckskin Mountains are exposed 16 km northeast of this contact, and no klippen are preserved between these two points. The absence of klippen indicates that the lower plate was completely overturned by 16 km of displacement on the detachment fault or that the klippen were left behind but are now eroded away or assimilated by a Miocene pluton in the western Boise Hills. We estimate that now-eroded or assimilated klippen could have been derived from a sliver at the trailing edge of the upper plate that was as much as 2 km thick and that detachment-fault displacement necessary to initiate unroofing of the mylonites could have been 2 km less than 16 km. Total displacement on this segment is thus estimated at 13±1 km.

Segment 2B. Lower plate rocks are exposed over a distance of 34 km along this segment of the transect in the southern Buckskin and Rawhide Mountains. We estimate that slivers of pre-Tertiary rock comprising up to 2 km of the upper plate could have been left stranded on the lower plate (but are now largely eroded away). This yields an estimate of 33±2 km as the displacement necessary to uncover the mylonitic lower plate.

Segment 2C. Along this transect segment the Buckskin-Rawhide detachment fault dips northward and to the north beneath the southwestern end of Buckskin Basin and the eastern edge of Rawhide Mountains. The north to southwest dipping base of the Tertiary stratigraphic sequence in the Poachie and Antilip Mountains projects downdip into the detachment fault. The point where the transect crosses the truncated contact at the base of the Tertiary section is interpreted to mark the offset, upper plate equivalent of the basal Tertiary contact in the western Boise Hills. The transect-parallel distance (approximately downdip) along the detachment fault from the edge
of exposed lower plate rocks in the eastern Buckskin Mountains to this point of intersection is estimated at 14.4 km and represents extension due to detachment-fault displacement. An additional 4.2 km of extension occurred along normal faults in the southern Poachie Mountains. Total extension on this segment of the transect is thus estimated at 18.6 km.

**Harcover Trench (Transect 3, Figure 7)**

Segment 2A. Sedimentary breccias in a giant Tertiary stratigraphic sequence above the Buckskin detachment fault are quite common along the eastern edge of the Buckskin Mountains and Little Harquahala mountains (west of dotted line on Figure 7). These consist of breccia and conglomerate that is cemented by alteration of the upper plate rocks. The breccias were probably deposited by catastrophic debris avalanches and were probably not deposited more than 10 km from their source area. The Buckskin detachment is thus at least 45 km displaced from the Buckskin detachment fault, which is consistent with a previous estimate [Reynolds and Spencer, 1985].

**Segment 3B.** The Buckskin fault dips southwestward at the interface between the Buckskin Mountains and Little Harquahala mountains (west of dotted line on Figure 7). The amount of extension that has occurred on other faults, if any, that is completely concealed beneath Buckskin detachment fault at the end of the Harcover Mountains is poorly constrained. The Buckskin detachment fault to the east of the Harcover Mountains is poorly constrained. The Buckskin detachment fault to the east of the Harcover Mountains is poorly constrained. The Buckskin detachment fault to the east of the Harcover Mountains is poorly constrained. The Buckskin detachment fault to the east of the Harcover Mountains is poorly constrained. The Buckskin detachment fault to the east of the Harcover Mountains is poorly constrained. The Buckskin detachment fault to the east of the Harcover Mountains is poorly constrained. The Buckskin detachment fault to the east of the Harcover Mountains is poorly constrained.

**Whipple Trench (Transect 4, Figure 7)**

Segment 4d. A dike swarm in the Mohave Mountains in the upper plate of the Whipple detachment fault is the probable offset equivalent of the lower plate Chambers Well dike swarm in the western Whipple Mountains [Howard et al., 1982; Nakata, 1982; Davis and Lister, 1988]. Restoration of 43 km of displacement on the detachment fault places the eastern edge of the dike swarm on the eastern edge of the Buckskin detachment fault. Concluded dikes to the west of the Mohave Mountains would require greater displacement. Precise realignment is not possible because of the following: (1) the base of the Mohave Mountains dike swarm, where it is truncated by the detachment fault, is not exposed, (2) the eastern edge of the Mohave Mountains dike swarm is not exposed, and (3) the Chambers Well dike swarm fans upward and therefore does not have vertical sides. An estimate of 45±5 km proximity encompasses the range of likely offsets.

**Segment 4b.** Additional extension is represented by displacement on low-angle normal faults that are structurally higher than the Whipple detachment fault (but may merge with it in the subsurface). The Cosumnes Peak low-angle normal fault, exposed as a thrust fault on the south side of the Mohave Mountains [Howard et al., 1990], has 4.2 km of displacement [Howard et al., 1982].

**Segment 4c.** Bedrock is concealed beneath Quaternary surficial deposits between the Mohave and Hualapai mountains, but is moderately to highly elevated where it is exposed along strike to the southwest between the Bill Williams and McCracken mountains [Sanzeno, 1980; Lucchitta and Sunesen, 1989] and on the north flank of the Buckskin and Rawhide Mountains [Shackelford, 1989a, b; Spencer and Reynolds, 1989]. Bedrock in both these areas is tilted approximately 20° to the west [Howard et al., 1982]. Extension in both of these areas could be the lateral equivalent of extension above the Cosumnes Peak fault and need not entirely represent additional extension. Considering these uncertainties we estimate that 4 to 20 km of extension has occurred within the approximately 25 km wide swath of concealed bedrock between the Mohave and Hualapai mountains.

**Segment 4d.** There are no known Tertiary normal faults within the south central Hualapai Mountains (north extended). The Hualapai Mountains are separated from the Transition Zone by the Big Sandy Valley, which formed by Tertiary extension (the timing of extension is not well constrained) [Scarborough and Wilt, 1979]. Reconnaissance geologic mapping has not revealed any normal faults on the floors of Big Sandy Valley, probably because they have been buried. Drilling and modeling of gravity data indicate that bedrock is greater than 1500 m deep (maximum depth of drill holes) in some of the valley [Lease, 1981; Oppenheimer and Sunesen, 1980]. Crossing of Precambrian bedrock on opposite sides of the valley, and the Cretaceous(?)/Miocene(? crustal rocks at the edge of the west side of the valley that has no offset equivalent on the east side, indicates that total extension was significantly less than the 12 to 16 km width of the sediment-covered valley floor. Three kilometers of extension is the minimum extension needed for a 49° fault dip) to produce the greater than 3 km difference in elevation between the crest of Hualapai Mountains and the minimum depth to bedrock beneath the Big Sandy Valley. If fault dip is more gentle or there are faults on both sides of the valley, extension could be much more, perhaps as much as 9 km. We thus estimate that extension has been 5.2 km.

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