Deformation associated with a continental normal fault system, western Grand Canyon, Arizona

Phillip Resor
Wesleyan University, presor@wesleyan.edu

Follow this and additional works at: https://wesscholar.wesleyan.edu/div3facpubs
Part of the Earth Sciences Commons, and the Environmental Sciences Commons

Recommended Citation

This Article is brought to you for free and open access by the Natural Sciences and Mathematics at WesScholar. It has been accepted for inclusion in Division III Faculty Publications by an authorized administrator of WesScholar. For more information, please contact anelson01@wesleyan.edu, jmlozanowski@wesleyan.edu.
Deformation associated with a continental normal fault system, western Grand Canyon, Arizona

Phillip G. Resor†
Department of Earth and Environmental Sciences, Wesleyan University, Middletown, Connecticut 06459, USA

ABSTRACT
Reverse-drag folds are often used to infer subsurface fault geometry in extended terrains, yet details of how these folds form in association with slip on normal fault systems are poorly understood. Detailed structural mapping and global positioning system (GPS) surveying of the Frog Fault and Lone Mountain Monocline in the western Grand Canyon demonstrate a systematic relationship between elements of the normal fault system and fold geometry. The Lone Mountain Monocline, which parallels the Frog Fault, is made up of two half-monoclinal flexures: a hanging-wall fold in which dips gradually increase toward the fault over ~1.5 km reaching a maximum dip of 25° and a footwall fold in which dips decrease away from the fault over ~0.5 km from a maximum of 12°. The highest dips associated with folding are found where throw on the Frog Fault and antithetic faults have been variably termed reverse-drag, monocline, deformation, modeling.

INTRODUCTION
Deformation associated with normal faults is responsible for many of the familiar structures common to areas of extension, including antithetic faults, synthetic faults, and reverse-drag folds. The development, growth, and interaction of these structures affects patterns of crustal deformation (e.g., Melosh and Williams, 1989; Willems, 1997) and basin architecture and evolution (e.g., Galloway, 1987; Vendeville and Cobbold, 1988). A thorough understanding of extensional fault-fold systems is important for studies of seismic hazard and earthquake triggering (Bruhn and Schultz, 1996; Ofoegbu and Ferrill, 1998), estimation of hydrocarbon trap volume (Tearpock and Bischke, 2002), and prediction of subseismic reservoir and aquifer fracturing (Laubach et al., 2000; Maerten et al., 2002). Although there have been numerous studies of normal faults and their associated deformation, many questions remain unresolved.

One fundamental question that remains unanswered is the process by which hanging-wall bedding is warped down toward a normal fault to form a fault-parallel half-monocline. These folds have been variably termed reverse-drag, rollover, or fault-bend folds (see Schlische, 1995; Janecke et al., 1998) for thorough reviews of extension-related folding. Conceptual models relate the development of reverse-drag with changes in fault geometry (Laubscher, 1956). Hamblin (1965) stated what has become the most widely accepted conceptual model for reverse drag wherein a hanging-wall fold develops in kinematic response to a fault with decreasing dip with depth (listric profile). The common occurrence of hanging-wall folds in extending deltas where faults have listric profiles associated with salt or shale detachments (Jackson et al., 1994; Morley and Guerin, 1996) has often been cited as support for this conceptual model. The presence of reverse-drag folds has thus been used as evidence to support the interpretation of listric normal fault geometry at depth when the fault geometry is unknown (Shelton, 1984), and a number of authors have developed kinematic models of normal fault deformation that relate the geometry of hanging-wall strata to fault shape through geometric constructions (Dula, 1991; Gibbs, 1983; Groshong, 1994; Medwedeff and Krantz, 2002; Verhall, 1981; White et al., 1986; Xiao and Suppe, 1992). These constructions require listric fault geometry to generate a hanging-wall fold.

In contrast, mechanical models of normal fault deformation suggest that reverse-drag folds may form in association with planar faults (Bott, 1997; Grasemann et al., 2004; King et al., 1988; Reches and Eidelman, 1995; Willems et al., 1996). In these models, the development of fault-related folds occurs as a natural response to fault slip over a limited vertical extent (e.g., limited by upper-crustal thickness for large faults or stratigraphic layering for smaller faults) and does not require a curved fault or drag as a folding mechanism (Barnett et al., 1987; Schlische, 1995). Mechanical models predict not only the development of hanging-wall half-monoclines, but also the development of an upward flexure in the footwall with beds dipping away from the fault. Most kinematic models, on the other hand, suppose a rigid footwall (White et al., 1986; Xiao and Suppe, 1992). Studies of near-planar, outcrop-scale normal faults (Grasemann et al., 2004; Gupta and Scholz, 1998) reveal deformation patterns consistent with mechanical models, including both hanging-wall and footwall folds. Therefore, although reverse-drag folds may form in association with listric faults, it is clear that listric faults are not required to generate these folds.

These two classes of models may lead to radically different interpretations of subsurface...
fault geometry and modes of extension based on the same pattern of hanging-wall folding. Kinematic models lead to the interpretation of listric fault geometry and detached extension, while mechanical models can produce a similar pattern of displacement with planar faults and thus imply thick-skinned extension. Since the subsurface geometry of many faults is unknown, a more complete understanding of extensional fault-fold relationships is important for understanding patterns of extensional strain and will contribute toward resolving the long-standing debate over the relative role of low-angle and high-angle faults in accommodating upper-crustal extension (see Axen, 2007, for a recent review).

This study specifically addresses the relationship between fault and fold geometry and the mechanisms by which folding occurs in association with crustal-scale normal faults by revisiting Hamblin’s (1965) field area in the western Grand Canyon. The Frog Fault and associated smaller faults as well as the 3D geometry of folded Esplanade Formation strata were mapped using GPS surveying techniques. The Frog Fault was chosen over the nearby better known Hurricane Fault because of the quality of exposure of the monocline and normal faults. The location of the Frog Fault system between the Parashant and Grand Canyons has led to exposure at nearly the same level in the Esplanade Formation over a length of ~10 km, with natural cross sections exposed in the canyons. The area allows direct observation of fault geometry over >800 m of vertical section as well as exploration of the relationship between fault and fold geometry along strike within a relatively simple fault system. The resulting 3D data set shows systematic relationships between fault and fold geometry that are consistent with mechanical models of normal fault-related deformation and suggest that fold geometry may be strongly controlled by secondary structures, namely the presence of synthetic or antithetic faults.

**STRUCTURAL SETTING**

The western Grand Canyon (Fig. 1) provides an excellent opportunity to study processes of normal faulting and fault-related deformation. The Colorado River in the western canyon cuts through a series of subparallel normal faults of moderate offset that mark the transition between the Colorado Plateau and the Basin and Range tectonic provinces. Due to the incision of the Grand Canyon, these faults can be observed in the field over more than a kilometer of vertical section from the Kaibab limestone, cropping out at plateau levels, to Precambrian rocks of the inner gorges. Fault throw is much less than canyon incision and thus both hanging-wall and footwall rocks may be directly observed. The relatively low strains and excellent exposures present an opportunity to investigate processes of continental extension without the large rotations and crosscutting fault patterns typical of much of the Basin and Range province (e.g., Proffett, 1977).

The study area is located along the Frog Fault between the Grand and Parashant Canyons ~5 km west of the Hurricane Fault and ~100 km south of Saint George, Utah (box in Fig. 1). Previous workers (Billingsley and Wellmeyer, 2003; Huntoon and Billingsley, 1981; Wernick et al., 1997) mapped the area at scales of 1:48,000 or smaller, but no detailed studies of the faults and folds exist. The fault system consists of the north-northwest-striking, westward-dipping Frog Fault (Frog Fault of Huntoon and Billingsley, 1981), and a series of smaller offset synthetic and antithetic normal faults (Fig. 2). In the study area, the Frog Fault is exposed from the Kaibab Formation to the Redwall Formation, a stratigraphic section that includes limestone, sandstone, siltstone, and shale.

Geological evidence suggests that the presently exposed normal faults of the western Grand Canyon largely formed at or near Earth’s surface. The onset of normal faulting is generally inferred to be Miocene or younger (Amoroso et al., 2004; Fenton et al., 2001; Stewart and Taylor, 1996). The presence of Paleocene-Eocene gravels (Elston and Young, 1991) on nearby plateau surfaces and ca. 75 Ma apatite fission-track cooling ages in the Grand Canyon (Dumitru et al., 1994) indicate that the region was eroded to near present-day plateau levels by late Eocene. An integrated analysis of uplift and erosion history for the Colorado Plateau (Pederson et al., 2002b) estimated that, with the exception of the canyons, the area has had less than 500 m of erosion since ca. 30 Ma.

Quaternary basalts (Fig. 1) drape many of the fault scarps and are in turn faulted themselves. These relationships allow for estimation of slip rates along the two major faults in the area. 40Ar/39Ar ages of offset basalt flows yield slip-rate estimates of 94–107 m/Ma for the Toroweap Fault over the last 500–600 ka and ca. 100–250 m/Ma for the Hurricane Fault over the last 200–850 ka (Amoroso et al., 2004; Karlstrom et al., 2007; Pederson et al., 2002a). Amoroso et al. (2004) and Fenton et al. (2001) interpret the consistency of slip-rate estimates over time spans ranging from ca. 15 ka to ca. 1 Ma as evidence that the rates were relatively constant since fault initiation at 3–5 Ma. Historic seismicity (e.g., the 1992 Saint George, Utah earthquake, Mw 5.6) indicates that the area remains tectonically active.

---

Figure 1. Shaded-relief location map of western Grand Canyon. The study area is outlined in the hachured box. Quaternary faults from U.S. Geological Survey (USGS) database. Quaternary basalts from Brown et al. (2000)—UT; Turner et al. (1991)—NV; and Hirschberg and Pitts (2000)—AZ. The bold dashed line is the approximate boundary between the Colorado Plateau and Basin and Range tectonic provinces.
Figure 2. Geologic and structural map of study area. Formation contacts modified from Huntoon and Billingsley (1981). Faults from this study. Bedding attitudes with asterisks (*) are from Huntoon and Billingsley (1981); all others are new data. Bold black outline is boundary of area of surface model presented in Figure 6. Bold gray box shows area of Figure 11. Coordinate system: Universal Transverse Mercator (UTM) zone 12 N, World Geodetic System (WGS) 1984 datum. Grid unit: meters. Pm—Permian; Penn—Pennsylvanian; Miss—Mississippian; Dev—Devonian.
One of the most striking features of the normal fault systems in the western Grand Canyon is their association with monocline flexures in the hanging walls (Fig. 3). These folds might be more properly termed half-monoclines, with bed dips increasing toward the associated normal faults. Hamblin (1965) proposed that these folds formed in response to slip on the parallel normal faults. Alternatively, it has been suggested that the monoclines of the western Grand Canyon are inherited structures formed during the contraction of the Laramide Orogeny (e.g., Huntoon, 2003). In this model the monoclines would have formed as forced folds over basement reverse faults which have subsequently been reactivated as normal faults. Normal faulting may have modified the preexisting structures, but the monoclines are viewed as largely the product of an earlier shortening event (Huntoon, 2003). The presence of Laramide structures in the region and outcrop-scale thrusts in the western Grand Canyon, as well as the coincidence of Laramide drainages with present-day normal faults is presented as evidence that the monoclines of the western canyon are Laramide in age and related to crustal shortening (see Young, 2001, and references therein). Although there have been detailed studies of the Kaibab monocline in the eastern Grand Canyon (Cooke et al., 2000; Reches, 1978; Tindall and Davis, 1999), a fold that is clearly associated with a basement reverse fault, there have been no published detailed structural studies of the monoclines in the western Grand Canyon. Stewart and Taylor (1996) noted that there are likely both Laramide and extension-related folds in the region and that to some degree each fault and fold relationship should be considered separately.

### Quantifying the 3D Structure

Modern mapping technology and geographical information systems (GIS) facilitate the construction of 3D geologic databases to describe the geometry of kilometer-scale geologic structures with spatial resolution (data density) of less than 10 m and precisions on the order of 1 m, thus providing important constraints for understanding the development of these structures through numerical models (Maerten et al., 2001). In this study, detailed structural mapping and global positioning system (GPS) surveying were employed to create a 3D structural model of the Lone Mountain Monocline and Frog Fault system. Geologic field work focused on mapping the trace and geometry of the Frog and other major faults in the area, mapping and surveying the geometry of well-exposed bedding surfaces in the Permian Esplanade Formation, and describing outcrop-scale deformation.

### Faults

The faults associated with the Frog Fault system were mapped at a scale of 1:6000 (Fig. 2), and structural data were collected to constrain fault geometry and kinematics along ~8 km of strike length. In the southeastern portion of the study area, the fault system is composed of two asymmetric grabens, each associated with a west-dipping master fault and lesser east-dipping faults. Throw on all of these faults decreases to the northwest, and all but the eastmost faults tip out near the northwestern end of the study area. The northwestern and southeastern fault systems intersect in a complex relay zone exposed in Parashant Canyon. In this area, fault slip is somewhat oblique with a consistent left-lateral component of slip on the southern ends of the northern segments. In addition to the main Frog Fault system, a northeast-striking system of discontinuous, small-offset faults is located in the footwall of the Frog Fault. This system of faults is associated with the development of a small graben and local intrusion of mafic dikes (see Billingsley and Wellmeyer, 2003, for dike locations).

The faults in the study area are variably exposed from the Kaibab Formation to the Redwall Formation, a vertical extent of >800 m that includes a variety of lithologies. Fault architecture varies as a function of the host lithology. In sandstone units of the Wescogame, Esplanade, and Coconino Formations, fault surfaces are irregular and appear to have formed through the shearing and linking of preexisting joint surfaces as described by previous workers for faulting in sandstone (Davatzes et al., 2003; Flodin et al., 2003). This faulting process leads to poor fault exposure in the field with highly variable orientations of individual exposed slip surfaces. Faulting in limestone leads to development of cemented breccia with well-developed fault surfaces similar to descriptions of faulted carbonates from other regions (e.g., Jackson and McKenzie, 1999). Exposed fault surfaces in carbonates often exceed 100 m². Kinematic indicators on these surfaces include well-developed

---

**Figure 3.** (A) Oblique aerial photograph of Frog Fault looking southeast with Parashant Canyon in foreground modified from Hamblin (1965). (B) Sketch cross section constrained by global positioning system (GPS) surveying of Esplanade bedding. Pe—Permian Esplanade Formation, MPs—Mississippian-Pennsylvanian lower Supai Group, Mr—Mississippian Redwall Formation.
slickenlines and meter-scale undulations that are generally parallel to slickenline orientations.

Fault orientation data are analyzed to determine average values for fault geometric and kinematic properties as well as to evaluate the role of changing lithology or depth on the geometry of faulting (Fig. 4). A plot of poles to measured fault surfaces (Fig. 4A) shows broad scatter around two primary fault orientations, a dominant west-dipping group with an average strike of 152° and dip of 72° and an east-dipping group with a nearly parallel strike of 331°, but a steeper average dip of 77°. Slip on these faults is nearly pure dip-slip (Fig. 4B) with an average slickenlines rake of 88°. Figure 4C shows a plot of fault dip versus elevation with symbols representing the different formations in the footwall of the fault. The solid symbols are measured in sandstone, and the stars are measured in limestone. There is no clear trend in fault orientation (e.g., a listric geometry) with either host-rock lithology or depth. The local scatter in the data is likely associated with fault zone evolution whereby faults develop through a process of segment linkage that often involves formation of joints oblique to the main fault trend and subsequent shearing of these structures to generate a through-going fault surface (Flodin and Aydin, 2004; Myers and Aydin, 2004; Segall and Pollard, 1983).

Deformation of Bedding

To document the 3D structure of the study area, including fault offsets and fold geometry, bedding surfaces were mapped and surveyed within the upper Esplanade Formation. McKee (1982) subdivided the Esplanade Formation into three units, a lower slope-forming unit (~20 m thick), a main cliff-forming unit (~65 m thick), and an upper cliff-slope unit (~45 m thick). These units contain variable proportions of red mudstone and siltstone interbedded with cross-stratified sandstone. In the western Grand Canyon, the upper unit also includes minor dolomites and interbedded gypsum (McKee, 1982). McKee has interpreted the depositional environment of the upper Esplanade Formation as coastal plain with intervening lagoons. The structural mapping focused on the top of the main cliff-forming unit and the upper cliff-slope unit because of the presence of fairly continuous sandstone layers separated by extensive siltstone layers.

In the study area, the upper cliff-slope unit of the Esplanade Formation is composed of ~50 m of interbedded very fine sand and siltstone that overlies the sand-rich, cliff-forming, lower Esplanade (Fig. 5A). Individual sand packages are 1–10 m thick, continuous over several kilometers, and encased in siltstone; thus the tops of
Figure 6. Data and models of the base of the upper Esplanade Formation (location marked on Fig. 2). (A) Global positioning system (GPS) point locations adjusted vertically to a consistent stratigraphic position. Points are color-shaded by elevation. (B) Oblique view of 3D GoCAD model interpolated from GPS point data with colored contours of structural elevation. White gaps are fault cuts for major west-dipping, normal faults. (C) Structure contour map derived from B. (D) Calculated dips for model elements in B; color-shaded contour interval is 3°. Coordinate system: Universal Transverse Mercator (UTM) zone 12 N, World Geodetic System (WGS) 1984 datum. Grid unit: meters.
beds may be traced continuously across much of the area and easily correlated across major faults. The tops of sandstone surfaces are interpreted as flooding surfaces that were likely near horizontal during deposition. Thin, interbedded marine dolomites and gypsum and the preponderance of siltstone suggest that these sediments were deposited in an environment with relatively low topography, supporting this interpretation. The geometry of bedding surfaces thus can be used to track relative vertical deformation of these markers subsequent to their deposition.

**Surveying of Esplanade Formation Geometry**

Individual, well-exposed bedding pavements within the upper Esplanade Formation were surveyed using differential GPS techniques with a mapping-grade, single-frequency receiver (Fig. 5B). Our method is similar to that described in detail by Maerten et al. (2001). Bedding surfaces were identified from aerial photographs and field inspection and then walked out with a mapping-grade GPS receiver set to record a point location every five seconds. Data acquisition was paused when crossing areas of poor exposure or erosion. Over 24,000 GPS location points (Fig. 6A) were collected primarily on the tops of four sandstone beds (Fig. 5A).

The error in measuring the structural position of a bedding surface at any point is estimated as ~3.5 m. This error can be subdivided into two major sources—GPS measurement error and stratigraphic error. GPS horizontal and vertical precisions (2σ) of ~0.5 m and ~1.5 m, respectively, were estimated by recording a batch of data for a single point over one hour and processing these data in the same manner as the structural data. The precision of individual data points will vary somewhat from this estimate depending on satellite geometry, atmospheric conditions, and local multipath effects (satellite signals reflected off nearby objects). Stratigraphic error was estimated by visual inspection in the field, and includes errors in correlation, stratigraphic variability, and erosion of bedding surfaces. These error sources were minimized by mapping the tops of thinner sandstone units, staying close to the contact wherever possible. By choosing the thinner sands, one minimizes the amount of erosion that can occur before the sandstone breaks up into blocks that can be clearly identified as eroded.

In addition to collecting GPS survey data on Esplanade Formation beds, local orientation data were measured using a field compass for comparison to the GPS-derived surface data (Fig. 2). Cross-stratification within the sandstone layers leads to variability in direct measurements of bedding attitudes while the overall structure appears to be relatively smooth. Thus measurement techniques that average over larger areas, such as GPS surveying, are more likely to capture the broad structural trends.

**Construction of a Model Surface**

The data from bedding surfaces of the upper Esplanade Formation were merged to create a single, idealized, upper Esplanade bedding surface to quantify the three-dimensional geometry of the fold and the throw profiles along the major faults (Fig. 6). Data from individual sand layers were vertically adjusted to account for stratigraphic thickness between layers. Bed thicknesses were determined from local measured stratigraphic sections and estimates from projecting GPS measurements along strike. This approach ignores the effect of dip on the apparent thickness and introduces an additional systematic error of less than 0.8 m. This error was not considered significant enough to warrant more involved correction methods such as an iterative correction based on estimating the local dip.

The merged data were interpolated at 50 m intervals within structural domains (e.g., footwall of the Frog Fault) using simple kriging with a second-order trend surface to create a regular gridded surface (Goovaerts, 1997). De-trending of data is required before kriging to satisfy the requirement of a zero mean (Isaaks and Srivastava, 1989), and has the added benefit of interpolating the general trend through data-poor areas while estimating local variation on this trend where data are more abundant. The resulting fault block grids were further interpolated in GoCAD (Mallet, 1992) to create a consistent faulted surface that covered the entire study area (Fig. 6B). GoCAD uses a discrete, smooth, interpolation algorithm (Mallet, 1989) that simultaneously minimizes the misfit to the data as well as the model roughness and allows for the explicit incorporation of faults. The resulting surface model is constructed of a triangular mesh that is optimized to the data density. The quality of the resulting model is assessed by comparing interpolated model results to field measurements of the elevation and dip of the Esplanade surface (Fig. 7). Regression of the relationship between the model and observed data are highly significant (P < 0.0001) but yield slopes somewhat less than one. These results indicate that for both elevation and dip the structural model slightly underpredicts the measured values. The largest errors are found in areas of poor data coverage or in close proximity to the faults. The root mean square error (RMSE) is 8 m for the elevation data and 6° for the dip data. The error can be attributed to measurement errors in the field data (see previous discussion), the smoothing constraint applied in the interpolation algorithm, and the spatial resolution of the structural model. For example, it was not practical to include all smaller faults because their slip profiles would often be poorly constrained by the data. These faults therefore appear as localized folds, and their effects can be readily seen in the calculated dips (for example, see the northeast-striking fault near the southern end of the map area).

**Results for Esplanade Bedding**

The overall structure at the upper Esplanade level is illustrated in map views of the raw GPS data and interpolated model surface (Fig. 6) and in cross sections extracted from the GPS data (Fig. 8). Consistently lower structural elevations at the northern end of the map area indicate a slight (<1°) regional tilt to the north-northwest (Fig. 6C). The Frog Fault system cuts the Esplanade surface with ~225 m of throw along the southeast segment that decreases to zero at the northwest tips of the northwestern segments (Figs. 6B and 6C). A system of folds, including...
Deformation within a normal fault system, western Grand Canyon, Arizona

Figure 8. Cross sections through merged global positioning system (GPS) data (see Fig. 6A for section lines). The sections were made by projecting a swath of data parallel to the section line onto a vertical plane. Sections A and B project 500-m-wide swaths of data, while sections C–E project 700-m-wide swaths. All sections are shown looking N33W with no vertical exaggeration.

In detail the structural geometry varies systematically along strike. The southeastern portion of the study area has the highest structural relief and greatest bed dips (Figs. 6C and 6D; Fig. 8, sections A–A′, B–B′, and C–C′). In this area the fault system is composed of the main Frog Fault with a ~3-km-long synthetic fault in the hanging wall. The Frog Fault has a fairly constant throw of ~225 m from the southeast edge of the map area until just south of Parashant Canyon, where throw decreases rapidly to zero over ~500 m in association with a well-developed relay with the easternmost of the northwest graben-bounding faults (Fig. 6C). Beds within this relay dip toward the northwest at ~20°. Throw on the synthetic fault varies from zero at its northwest and southeast tips to near 60 m at its center.

The hanging-wall monoclinal is most clearly developed along this section of the Frog Fault system and is associated with 150–200 m of structural relief. Bedding within the half-monocline is progressively tilted toward the fault from near horizontal at distances greater than ~1500 m from the fault to dips of greater than 25° at distances of less than ~500 m. A plot of bedding dip versus distance from the fault (Fig. 9) allows for further analysis of fold shape. Dips are compiled from direct field measurements and from GPS-derived slopes averaged over a 40 m window. Analysis of variance (ANOVA) testing of the dip data indicates that they are significantly (P <0.001) better described by a second-order polynomial fit (R² = 0.72) than a linear fit (R² = 0.57). These results suggest that the fold curvature (approximated by the second derivative) increases toward the fault rather than remaining constant across the fold.

The highest hanging-wall dips (>25°) are associated with the maximum throw on the synthetic fault and are located in the footwall of the synthetic fault. This relationship is clearly illustrated in Figure 8 where the apparent dips in cross section increase from section A–A′, where there is almost no slip on the synthetic fault to section B–B′, where slip is at a maximum on the synthetic fault and then decreases again toward the northwest in section C–C′ as throw on the synthetic fault decreases. Along a line parallel to the synthetic fault, dips increase from ~6° beyond the tips of the fault to ~12° in the hanging wall and >20° in the footwall near the center of the fault. In addition to the hanging-wall fold, a smaller fold is present in the footwall. This fold is best illustrated in section C–C′, where bedding dips increase from flat-lying to 5°–12° over a distance of ~500 m when approaching the Frog Fault from the east.

The structural pattern changes dramatically to the northwest of Parashant Canyon. This portion of the Frog Fault system is composed of two asymmetric grabens with the larger faults dipping...
toward the west (Fig. 8, sections D–D′ and E–E′). With the exception of two well-exposed relay ramps that have steeper northwesterly dips, bedding dips are consistently lower in this part of the fault system with maximum eastward dips of ~6°. The easternmost fault has the greatest throw, which rapidly increases from zero at its southeastern tip to ~200 m in association with the Parashant Canyon relay and then more gradually decreases toward the northwest tip of the fault. The antithetic fault in the eastern graben has ~50 m of throw, where it is well constrained in its central portion and the throw decreases to the northwest of this point. Beds within the graben dip gently to the east; however, poor exposure has prevented better resolution of the details of the structure. The footwall of the easternmost fault shows an upwarp that is particularly clear in section E–E′. In the western graben, the structure is closer to a half-graben with a maximum throw of ~75 m on the west-dipping fault and <25 m of throw on the east-dipping fault. Esplanade bedding is warped downward toward the west-dipping fault in an upper half-monocline with maximum dips of ~6° in the hanging wall of the west-dipping fault.

Figure 10 presents two examples of how variations in fault throw are related to the amplitude of the spatially associated folds. Figure 10A is a panoramic photograph showing the southeastern tip of the hanging-wall synthetic fault as viewed from near the break in slope of the Lone Mountain monocline. The surface in the foreground and the surface on top of the middle ridge are both eroded to the top of the middle cliff-forming unit of the Esplanade. Throw on the synthetic fault increases from zero at the right-hand side of the picture to ~60 m on the left-hand side. Associated with this throw gradient, the footwall of the fault is increasingly uplifted and the hanging wall is further depressed, thus amplifying the overall dips of the Lone Mountain monocline. Figure 10B shows elevation profiles extracted from the Esplanade Formation surface model (Fig. 6) for footwall and hanging-wall cutoffs and the hinge of the Lone Mountain monocline (Fig. 2). Fault throw is the difference between the footwall and hanging-wall cutoff elevations and fold amplitude is the difference between footwall cutoff and fold-hinge elevations. The plot illustrates that fold amplitude and fault throw decrease toward the northwestern tip of the fault system.

Parashant Canyon Cross Section
Parashant Canyon provides a natural cross section through the study area (Fig. 11) and an opportunity to evaluate fold shape over ~300 m of section including sandstone, massive
Deformation within a normal fault system, western Grand Canyon, Arizona

limestone, and interbedded limestone and shale. High-precision GPS measurements were not practical in the inner-canyon environment due to limited sky visibility. Variations in the fold shape with depth in the canyon are thus evaluated by combining GPS measurements of upper Esplanade bedding with profiles derived from digital elevation models (DEM) for the top of the Redwall limestone and a thin limestone within the Watahomigi Formation. The DEM profiles were created by interpreting bedding surfaces on orthorectified aerial photographs and then extracting the elevation information from U.S. Geological Survey (USGS) 10-m DEMs. All horizons were projected onto a planar section striking 55°. The individual horizons were projected up to 1 km along strike which leads to minor error in close proximity to the faults, such as the apparent thinning of the section between the top of the Redwall limestone and the top of the Watahomigi Formations in the immediate footwall of the Frog Fault.

The resulting cross section (Fig. 11B) shows a fold that is remarkably similar in profile throughout this part of the section. The beds are nearly horizontal ~1.5 km to the southwest of the fault in the hanging wall and gradually increase in dip toward the Frog Fault with maximum dips greater than 10°. The data do not extend to within 250 m of the fault where dips measured in the field reach 19°. There is a smaller footwall flexure developed over less than 1 km that appears to be more pronounced at the top of the Redwall limestone. Maximum footwall bed dips are ~6°. The bed-perpendicular thickness between the surveyed layers was estimated by fitting a second-order surface to the raw data and then measuring the thickness perpendicular to the local dip. Using this approach, thicknesses varied by less than 6 m over a 240-m-thick section. This variation is on the order of the error in the DEM, suggesting little to no variation in bed thickness associated with the observed folding.

Outcrop-Scale Deformation

Observations of outcrop-scale deformation provide additional data for understanding fold development. The structures are indicative of strains associated with folding and can thus help differentiate between various models of fold formation. Structures observed in the field suggest two distinct outcrop-scale deformation mechanisms—slip on preexisting joints and development of new joints. These structures are typically associated with measurable changes in the dip of bedding surfaces.

A regionally extensive joint set is observed within sandstone of the upper Esplanade Formation. In the hanging wall, the predominant joint set strikes ~115° and is nearly perpendicular to bedding. In sandstone of around 1 m thickness, these joints have ~1 m spacing. A second joint set strikes nearly due south, also is approximately perpendicular to bedding, and is present in both the hanging and footwall (Welsh, 2005).

Figure 12A presents two lower-hemisphere, equal-area stereoplots of poles to the predominant joint set. The left-hand plot is from a location near the upper hinge of the Lone Mountain Monocline (LM0223), where bedding dips 4° to the northeast. At this location, the joint set has a mean orientation of 297° with an 83° dip toward the northeast. At location LM0227, the bedding dips 17° to the northwest, and the predominant joint set has a mean orientation of 137° and a dip of 82° toward the southwest. The change in orientation of the joint set is consistent with a rigid-body rotation of ~15°, similar to the observed change in the dip of bedding.

Within the hanging-wall fold, joints of the main set are locally reactivated in shear with up to 10 cm of east-side–down slip independent of joint orientation. Localized zones of intense small-scale faulting were also observed (Fig. 12B). In these zones, a complex array of steeply dipping faults accommodates extension of the beds and is associated with an increase in the dip of the bedding of 2°–4° across the zone in the direction approaching the Frog Fault.

Newly formed joints also appear to play a role in accommodating folding of the Esplanade Formation. Both local zones of enhanced jointing within sandstone of the upper Esplanade as well as a systematic increase in joint density within the main cliff-forming unit of the Esplanade Formation were documented in the field. Local zones of intense jointing are associated with changes in bedding dip of ~3°. In these zones, steeply dipping joints are spaced <10 cm apart over a width of several meters (Fig. 12C). The joints cut through intervening siltstone, suggesting that small-scale, layer-parallel slip is not a significant deformation mechanism.

In Parashant Canyon, the main cliff-forming unit of the Esplanade Formation in the hanging wall of the Frog Fault shows a marked increase in fracture frequency toward the fault. This
change in frequency was quantified by interpreting fractures within a 30-m amalgamated sandstone package from a photo panorama of the cliff exposure (Fig. 13). The resolution of the photo panorama (~5 cm) limits the ability to interpret fractures that have not been significantly enhanced by weathering. A modified area method (Wu and Pollard, 1995) was used to calculate fracture frequencies, measured in units of 1/m. A cross-section line was drawn perpendicular to the structural trend, and equal-area rectangles (30 m × 50 m) were projected from the section onto the cliff face. All fractures within each sample box were interpreted, and their total lengths were summed. Fracture frequency was calculated by dividing total fracture length by the area of the sample region. The results are plotted against distance from the fault in Figure 14. The dashed reference line for fracture saturation was calculated for fracture spacing equal to the layer thickness (30 m). In the upper plot, the GPS-derived fold geometry at the top of the Esplanade cliff-forming unit is projected ~1 km along strike from the southeast (see Fig. 10). The plot shows that fracture frequency decreases systematically away from the fault and appears to asymptotically approach a background value of ~0.05 m⁻¹ at a distance greater than 1.5 km from the fault, a distance comparable to the width of the hanging-wall fold.

**Discussion of Structural Observations**

The structural associations described above demonstrate a clear relationship between the geometry of the fault system and that of the folding of the Esplanade Formation. The amplitude of the hanging-wall half-monocline is proportional to the throw on the associated fault(s). Additionally, maximum bed dips associated with folding are directly related spatially to smaller-scale structures. Where synthetic faulting is present in the hanging wall, dips are at a maximum; whereas, where antithetic faulting is present, dips are more moderate. The dip of bedding clearly changes magnitude across the faults, indicating that the normal faulting has significantly modified any preexisting structure. These lines of evidence support the conceptual model that the observed monoclinal structure developed primarily in association with Basin and Range normal faulting rather than Laramide reverse faulting. The tilting of 500–600 ka basalt flows in the hanging wall of the Toroweap fault and variations in river incision rates across hanging-wall blocks (Karlstrom et al., 2007) provides additional evidence that significant tilting of strata has occurred in association with normal faulting in the western Grand Canyon.

**MODELING FAULT-RELATED DEFORMATION**

To understand the development of hanging-wall and footwall half-monoelines better, and to elucidate the role that specific secondary structures may play in fold development, a series of models were executed using 2D elastic dislocation theory (Savage, 1980). Elastic models have been shown to be useful in modeling earthquake deformation (e.g., Johnson et al., 2001; Jónsson et al., 2002; Reid, 1910; Savage, 1980; Thatcher, 1979). Their applicability in modeling geologic structures that accommodate strains greater than a few percent and that develop over geologic time scales is somewhat controversial, but a number of studies have found a compelling similarity between predictions from elastic models and observations of geologic structures associated with faulting (e.g., Bourne and Willemse, 2001; Grasemann et al., 2005; Gupta and Scholz, 1998; Maerten, 2000; Willemse et al., 1996). Although it is unlikely that the elastic stresses associated with the total slip on a large fault are preserved over geologic time, it appears that the distribution of elastic stresses for incremental slip on such a fault play an important role in the development of associated structures.
Furthermore, displacement fields for incremental slip can offer important insights about the distribution of cumulative displacements (Fiore et al., 2007).

The edge dislocation model is particularly well suited for solving inverse problems, where patterns of ground deformation are used to infer unknown fault parameters, due to its mathematical simplicity. Dislocation models for the present analysis were implemented in Matlab following theory outlined by Freund and Barnett (1976) and Steketee (1958a, 1958b). The base model includes a single edge dislocation, a dipping fault of constant slip over a finite down-dip extent or height, within a homogeneous isotropic elastic half space (Fig. 15A). Although the assumption of constant slip leads to a nonphysical condition at the lower fault tip, the surface displacement patterns predicted by the dislocation model are only marginally different from those predicted by a more realistic crack model (Pollard and Segall, 1987; Rudnicki and Wu, 1995) with a constant stress drop and slip that tapers elliptically to zero at the lower fault tip (Fig. 15B). For fault parameters similar to those used in the following analyses of the Frog Fault, 200 m of slip on a 60°-dipping fault extending 1000 m down dip, the maximum difference in surface displacement between the dislocation model and a crack model, with the fault height scaled by a factor of 4/π to maintain the total moment (slip integrated over height), is less than 10 m or 5% of the slip on the fault (Fig. 15B).

For further comparison, if the crack displacements are assumed to be the true ground displacements, an inversion using the dislocation model yields the correct dip and height (scaled by π/4) but underestimates slip by ~10%.

I have chosen mechanical modeling rather than more commonly applied kinematic approaches for several reasons. First, mechanical models include explicit consideration of the mechanical properties of the rock and the forces that cause the deformation. Mechanical models “balance” material and forces in 3D using conservation of mass and momentum rather than balancing area or line length using ad hoc geometric assumptions. Second, mechanical models naturally incorporate footwall and hanging-wall folding that is consistent with field observations. Kinematic models require prior knowledge and modification of standard methods to incorporate footwall deformation (Groshong, 1994). Third, mechanical models are naturally extendable to explore issues such as the effect of secondary faults on the deformation. Incorporation of secondary structures in kinematic models requires additional ad hoc assumptions (e.g., Song and Cawood, 2001). Finally, in mechanical models, the kinematics of folding and deformation adjacent to the fault follow from the conservation laws rather than being presupposed. Mechanical models thus predict heterogeneous stress and strain fields that facilitate the interpretation of the distribution and orientation of off-fault structures (e.g., King et al., 1994). In contrast, kinematic models typically assume a mechanism, such as inclined shearing, to balance area and thus artificially impose a uniform orientation of incremental strain.

**Forward Modeling: The Role of Secondary Structures**

A series of forward models facilitates exploration of the role that antithetic and synthetic faults play in modifying fold geometry. In each of these models, 200 m of normal slip is prescribed on a 60°-dipping master fault that extends 1000 m down dip. Secondary structures, dipping 60° and extending to 500 m depth, are introduced in the hanging wall 750 m from the master fault and are free to slip in response to the stresses generated by slip on the primary fault. Deformation of the free surface is compared to patterns of deformation observed in the field assuming that the observed folds formed at or near Earth’s surface.

The fold shape predicted by an initial model with a single dislocation (Fig. 16A) is similar to the folds observed in the field (Fig. 8, profile C–C′); in particular, the model predicts not only the hanging-wall half-monocline, but also a smaller amplitude, footwall half-monocline. A synthetic fault introduced into the model (Fig. 16B) slips 22 m in response to slip on the master fault and locally enhances the tilting of the surface (Fig. 16D). As discussed previously, a similar pattern is observed in association with...
the synthetic fault in the southeast portion of the map area (Fig. 8, profiles A–A′, B–B′, and C–C′). Introduction of an antithetic fault (Fig. 16C) creates an asymmetric graben with 62 m of slip on the antithetic fault. The tilting of the surface is greatly reduced in both the hanging wall and footwall of the antithetic fault, while the footwall deformation of the master fault remains largely unaffected (Fig. 16D). A similar effect is seen in the field where hanging-wall bed dips are lower in the northwestern half of the map area, where antithetic faults are prevalent (Fig. 8, profiles D–D′ and E–E′). The predicted fault-fold relationships of the elastic models are thus qualitatively similar to the field observations and suggest that the elastic strains associated with antithetic and synthetic faults play an important role in determining fold shape.

Inversion for Fault Parameters

Application of the edge dislocation model in an inverse sense allows for estimation of unknown fault parameters from observed patterns of deformation. Fault dip, height, and slip can thus be estimated from data available at the surface. The best-fit single dislocation model for the southeast (Fig. 8, profile C–C′) and northwest (Fig. 8, profile D–D′) portions of the study area. The best-fit single dislocation model for profile C–C′ estimates 188 m of slip on a 52°-dipping fault that extends 1300 m down dip (Fig. 18A). The root mean square residual (RMS) of this model is 7.7 m. Adding a synthetic dislocation reduces the model residual by 23% to 5.9 m; however, it yields the unlikely result that the synthetic fault with 100 m of slip extends to 1200 m, while the master fault with 276 m of slip extends to only 600 m. The preferred model (Fig. 18B) thus includes the additional constraint that the synthetic fault height be equal or less than the master fault. This model estimates 183 m of slip on a 66°-dipping master fault that extends to 1000-m depth and 45 m of slip on a 50°-dipping synthetic fault that extends to 1000-m depth. The preferred model results in a 6.5 m residual, somewhat greater than the estimated data error (3.5–4.3 m). The greatest systematic misfit occurs in the block between the master and synthetic fault where tilting is underpredicted. GPS profile D–D′ is modeled with three dislocations.

Figure 14. Top: Global positioning system (GPS)-surveyed fold profile at the top of the Esplanade cliff unit (see Figs. 11 and 13). Bottom: Graph of fracture frequency plotted versus distance from the Frog Fault. For both plots, the Frog Fault is located 600 m to left of the y axis.

Figure 15. Comparison between dislocation and crack models. (A) Sketch of model setup. In the dislocation model, the fault surface (black dashed line) extends for a limited down-dip distance into an elastic half space. Slip on the dislocation, illustrated by the dark gray box, is constant. An equivalent crack model (solid gray line) has the same slip at the surface but tapers toward the tip in an elliptical profile (light gray). The crack model extends deeper into the crust by a factor of 4/π to maintain an equivalent moment (slip*depth). (B) Modeled surface displacements for equivalent crack (solid gray line) and dislocation (dashed black line) models. See text for details of model parameters.

Figure 16. Predicted surface displacements for dislocation models. (A) Single 60°-dipping dislocation extending 1000 m down dip with 200 m of slip. (B) Same as model A but with 60° synthetic fault extending 500 m down dip. Synthetic fault slips 22 m in response to slip on the master dislocation. (C) Same as model A, but with 60° antithetic fault extending 500 m down dip. Antithetic fault slips 62 m in response to slip on the master dislocation. (D) Vertically exaggerated surface displacement profiles for outlined area in A–C.
Deformation within a normal fault system, western Grand Canyon, Arizona

Figure 17. Model sensitivity analysis. (A) Plot of surface deformation patterns for 60°-dipping faults with 200 m of normal slip and variable down-dip extents. (B) Plot of surface deformation for faults with 200 m of normal slip extending to 1000 m depth with variable dip. (C) Fault height-dip misfit space for single dislocation model of global positioning system (GPS) profile C–C′ (Figs. 8 and 18). Contours are 2 m of root mean square (RMS) residual.

Figure 18. Best-fit 2D inverse models to profiles through global positioning system point measurements of upper Esplanade surface (Fig. 8). All models shown at 1:1 (top) and with 3x vertical exaggeration (bottom) (A) Single dislocation model fit to profile C–C′. (B) Two-dislocation model fit to profile C–C′. (C) Three dislocation model fit to profile D–D′.

Discussion of Model Results

Forward models of normal fault-related deformation using elastic dislocation theory yield qualitatively similar deformation patterns to those observed in association with slip on the Frog Fault. Specifically, the models predict the development of both hanging-wall and lesser footwall half monoclines. Additionally, the incorporation of secondary (smaller) synthetic and antithetic faults in the dislocation models modifies the results in a manner similar to the patterns of deformation observed in the field. Dips are at a maximum in association with synthetic faulting and a minimum in association with antithetic faulting.

Inverse modeling illustrates that the observed patterns of deformation can be fit quantitatively to a first order by deformation associated with planar faults. Systematic residuals are associated with unmodeled structures or with closely spaced faults. The Esplanade beds between closely spaced faults appear to have acted in a more block-like fashion than the elastic models predict (Fig. 18) and may indicate more efficient relaxation of elastic stresses in fault-bounded blocks. Westaway and Kusznir (1993) suggested that similar apparently rotated blocks may have in fact formed through vertical shearing rather than rotation. In the case of the Frog Fault, however, the continuity of marker beds in the field suggests that the rotation is not entirely accomplished by pervasive shearing at a high angle to bedding.

The best-fit models (Fig. 18 and Table 1) have throw values that are within 15–20 m of observed throw and fault dips that are within the range of observed values, although often lower than values observed on a given fault segment. The models consistently yield 1000–1500-m down-dip extents for the major faults. These results imply ~1300–1900-m down-dip extents for equivalently scaled crack models. These heights appear anomalously low in comparison to segment lengths of 6–8 km and a total fault system length of >20 km (Huntoon and Billing-sley, 1981). Using the equivalent crack model values results in fault aspect ratios (length/height) of ~4:1 for individual segments and >12:1 for the entire Frog Fault system. Fault slip to height ratios of ~0.1 are also anomalously large in comparison to typical fault slip/length ratios of 0.03 (Schlische et al., 1996). Fault slip/length ratios, however, appear normal (~0.03) for individual segments to low (~0.01) for the entire fault system. One possible explanation for
the apparently low fault heights is that the deformation is strongly influenced by the contact between the layered sedimentary rocks and the underlying basement located ~1 km below the top of the Esplanade Formation (Billing and Wellmeyer, 2003). The faults may be or have been layer bound so that propagation occurred largely along strike rather than down dip. Layer-bound faulting leads to high fault aspect ratios and relatively low slip/length ratios (Schultz and Fossen, 2002; Soliva et al., 2005).

Another possible explanation for the apparent slip/height discrepancy is that inelastic deformation mechanisms may have acted to localize strain over geologic time scales (e.g., Bott, 1997; Hassani and Chery, 1996). These processes might reduce the width of the fold, which would lead to an underestimate of fault depth using the elastic model. Studies of crustal-scale flexure associated with normal faults have demonstrated that fold widths are consistently smaller than those predicted by elastic models. These results are often interpreted in terms of an effective elastic thickness, which is typically much less than the thickness of the upper crust (e.g., Stein et al., 1988). In the case of the Frog Fault, the abundant joints in the half-monocline (Figs. 13 and 14) may have weakened the rock and localized bending strains. Previous studies have demonstrated quantitatively how joints may act to increase the compliance of an elastic material and accommodate inelastic strains (Segall, 1984; Walsh, 1965). Layer-parallel slip may also reduce fold wavelength by reducing the effective elastic thickness of the folding layer.

**IMPLICATIONS FOR THE DEVELOPMENT OF REVERSE DRAG**

The widespread application of kinematic models to describe normal fault-related deformation has led to the common assumption that rollover or reverse-drag folds indicate listric fault geometry (e.g., Twiss and Moores, 2007, p. 69). In the case of the Frog Fault and Lone Mountain Monocline, the application of a vertical shear model, for example, predicts a listric fault profile with a detachment at ~3 km. This kinematic model fits the observed hanging-wall fold profile to a similar degree as the single fault mechanical model when constructed with an equal number of model parameters. The kinematic model, however, does not naturally predict the observed footwall deformation or the effect of secondary faults. The planar fault model thus yields a better fit to the Esplanade Formation surface data. This result does not rule out the possibility that the faults are listric or kinked at depth, but rather illustrates that the folds could be generated by planar faults of finite extent, a result that is consistent with field observations of high-angle faults extending over >800 m of vertical section.

Previous studies have proposed that reverse-drag folds may form as a natural consequence of the heterogeneous displacement field around finite faults (Barnett et al., 1987) and demonstrated, using mechanical models of finite faults in an elastic whole space, that reverse drag is expected for marker layers that intersect normal faults at angles greater than 30°–40° (Grasemann et al., 2005). The utility of these models in describing normal fault-related deformation over geologic time scales has been illustrated by comparison to outcrop-scale examples where faults and reverse-drag folds can be observed over their entire along-strike and down-dip extents (Gupta and Scholz, 1998). Models incorporating planar normal faults that rupture the entire elastic upper crust (Bott, 1997; King et al., 1988; Stein et al., 1988) have also been successful at reproducing patterns of deformation over geologic time scales and are consistent with many observations of historic earthquake fault geometry (e.g., Chiaraluce et al., 2003; Collettini and Sibson, 2001; Jackson and White, 1989; Stein and Barrientos, 1985).

The model of an edge dislocation in an elastic half space (Fig. 15A) can be used to derive a general expression for the maximum hanging-wall dip expected in a reverse-drag fold formed due to slip on an isolated planar fault of finite down-dip extent. The results are applicable to folds that have formed at Earth’s surface from previously horizontal layers. In Figure 19 the maximum expected bed dip is expressed as a function of fault dip and fault slip to height ratio. The fault height is scaled by a factor of 4/π, to an equivalent crack model height, to facilitate comparison to natural examples. Reverse-drag folds with the greatest hanging-wall dips are found in association with high slip to height ratios and low fault dips. For average slip/length scaling values (0.03) the expected maximum dip would be ~2° for 60°-dipping faults. Faults may, however, accommodate greater slip in cases where they interact with nearby faults (Willen et al., 1996) or where the fault tip is mechanically constrained (Schultz and Fossen, 2002; Soliva et al., 2005). Slip/height ratios in these cases may exceed values of 0.125 (estimated from data presented in Soliva et al., 2005) and hanging-wall bed dips may exceed ~6° for isolated 60°-dipping normal faults. In the case of the Frog Fault system, the largely isolated western graben fault is associated with ~6° hanging-wall dips. Greater maximum hanging-wall dips may be achieved due to superposition of subparallel faults, as illustrated for the southeastern portion of the Frog Fault, or through inelastic deformation mechanisms such as fracturing and bedding-plane slip.

The results of this analysis provide additional support for the suggestion of Grasemann et al. (2005) that deformation associated with finite faults may be a better general explanation for the development of reverse-drag features observed in association with many normal faults. Inferences of non-planar (listric) fault geometry should generally be supported by statistically significant improvements in the fit to observed patterns of deformation or independent observations. A critical reevaluation of the reverse-drag listric fault connection may have implications well beyond simply understanding specific fault and fold geometries. Kinematic-listric and

**TABLE 1. MODEL AND OBSERVED FAULT PROPERTIES FOR SELECTED GLOBAL POSITIONING SYSTEM PROFILES**

<table>
<thead>
<tr>
<th>Profile</th>
<th>Fault</th>
<th>Height (m)</th>
<th>Modeled properties</th>
<th>Dip (°)</th>
<th>Throw (m)</th>
<th>Observed properties</th>
<th>Dip (°)</th>
<th>Throw (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>C-C†</td>
<td>Frog</td>
<td>1000</td>
<td>183</td>
<td>66</td>
<td>167</td>
<td>67–72</td>
<td>175–200</td>
<td></td>
</tr>
<tr>
<td>D-D†</td>
<td>Synthetic</td>
<td>1000</td>
<td>45</td>
<td>50</td>
<td>34</td>
<td>60–67</td>
<td>50</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Eastern</td>
<td>1500</td>
<td>183</td>
<td>65</td>
<td>166</td>
<td>76–80</td>
<td>175–200</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Antithetic</td>
<td>300</td>
<td>58</td>
<td>70</td>
<td>55</td>
<td>?</td>
<td>50</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Western</td>
<td>1400</td>
<td>76</td>
<td>50</td>
<td>58</td>
<td>70–77</td>
<td>~75</td>
<td></td>
</tr>
</tbody>
</table>

† Model values are for preferred two-fault model.
planar-mechanical models lead to significantly different interpretations of subsurface geometry from the same pattern of reverse-drift folding, a result that may impact estimates of hydrocarbon volume where fault and fold systems are poorly imaged. The two classes of models also lead to significantly different estimates of crustal extension. Kinematic models imply that the entire crust above a detachment horizon has been extended by the observed fault throw for vertical shear models, and higher values for anticlinal inclined shear models (White et al., 1986). Finite fault mechanical models, however, imply that extension is heterogeneous and limited to a finite volume around a fault (Grasemann et al., 2005; Pollard and Segall, 1987). Fault shape is also likely to impact earthquake rupture processes including rupture velocities, slip triggering, and ground-motion amplitudes (e.g., Ofoegbu and Ferrill, 1998).

CONCLUSIONS

Large-scale mapping of the Frog Fault system and deformed Esplanade strata, coupled with modeling of normal fault deformation, has yielded the following insights into the development of the monoclines of the western Grand Canyon and the development of normal fault-related, reverse-drift folds in general. (1) The consistent relationship between normal fault geometry and the vertical displacement field as documented using the hanging-wall and footwall half-monocones strongly suggests that these structures formed during Basin and Range extension rather than being inherited from Laramide contraction. (2) In contrast to results reported by Hamblin (1965) for widely spaced outcrops along the Hurricane fault, the faults in the vicinity of the Frog Fault system dip steeply through more than 800 m of exposed section, and yet a hanging-wall, “reverse-drift” fold is well developed. (3) The reverse-drift folding includes a hanging-wall half-monocone and footwall flexure that parallel the main fault trends, similar to the patterns of deformation predicted by mechanical models of deformation associated with planar faults. (4) There is a consistent relationship between the geometry of the hanging-wall half-monocone and the presence of secondary structures: bedding dips within the folded strata are greater in areas of synthetic faulting and lower in areas of antithetic faulting. (5) Best-fit elastic dislocation models consistently yield ~1 km fault heights that are anomalously small in comparison to fault length and tentatively yield ~1 km fault heights that are anomalous. Best-fit elastic dislocation models consistent with the deformation and lower in areas of antithetic faulting.

Acknowledgments

This work was initiated as part of the author’s Ph.D. dissertation at Stanford University. Support for the author’s studies at Stanford was provided in part by the ARCO Stanford Graduate Fellowship. Partial funding for this research was provided by the Stanford Rock Fault Project, the Stanford School of Earth Sciences Mcgee grants, American Association of Petroleum Geologists Grants-in-Aid, Geological Society of America research grants, and Wesleyan University. Laura Chiarounante, Nick Davatzes, Kaj Johnson, Jordan Muller, Ian Mynatt, and Kurt Sternloff for help with fieldwork. Thanks to Dave Pollard for reviews and discussions that helped in many ways to improve this research. Thanks to an anonymous reviewer, Richard Groshong, and John Fletcher for detailed insightful comments that greatly improved the final manuscript. This research was completed under National Park Service research permit LAME-2001-SCI-0015.

REFERENCES CITED

Bourne, S., and Willemse, E., 2001, Elastic stress control of reverse drag folds in general. (1) The Canyon and the development of normal fault-deformation, has improved the final manuscript. This research was funded by the Stan-


