Observations on normal-fault scarp morphology and fault system evolution of the Bishop Tuff in the Volcanic Tableland, Owens Valley, California, U.S.A.

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ABSTRACT

Mapping of normal faults cutting the Bishop Tuff in the Volcanic Tableland, northern Owens Valley, California, using side-looking airborne radar data, low-altitude aerial photographs, airborne light detection and ranging (LiDAR) data, and standard field mapping yields insights into fault scarp development, fault system evolution, and timing. Fault zones are characterized by multiple linked fault segments, tilting of the welded ignimbrite surface, dilation of polygonal cooling joints, and toppling of joint-bounded blocks. Maximum fault zone width is governed by (i) lateral spacing of cooperating fault segments and (ii) widths of fault tip monoclins. Large-displacement faults interact over larger rock volumes than small-displacement faults and generate larger relay ramps, which, when breached, form the widest portions of fault zones. Locally intense faulting within a breached relay ramp results from a combination of distributed east-west extension, and within-ramp bending and stretching to accommodate displacement gradients on bounding faults. One prominent fluvial channel is offset by both east- and west-dipping normal faults such that the channel is no longer in an active flowing configuration, indicating that channel incision began before development of significant fault-related geomorphic features. The channel thalweg is “hanging” with respect to modern (Q1) and previous (Q2) Owens River terraces, is incised through the pre-Tahoe age terrace level (Q4, 131–463 ka), and is at grade with the Tahoe age (Q3, 53–119 ka) terrace. Differential incision across fault scarps implies that the channel remained active during some of the faulting history, but it was abandoned between Q2 and Q3 time, while faulting continues to the present day.

INTRODUCTION

The Basin and Range Province is a broad region of western North America that has experienced Cenozoic crustal extension manifested by extensional faulting, formation of graben and half-graben basins separated by narrow mountain ranges, and volcanism (Fig. 1, inset map). The province extends 3700 km from Canada through the United States to southern Mexico, with a width of ~800 km (Stewart et al., 1998). Timing of basin formation, fault activity, andesite-rhyolite volcanism, and the transition from diverse andesite-rhyolite volcanism to basaltic or bimodal volcanism all generally become younger from east to west across the Basin and Range Province (Stewart, 1998). Some of the most active deformation in the Basin and Range Province is occurring in eastern California in a region referred to as the Walker Lane belt—an area of interconnected normal and strike-slip faults that includes Miocene to present large-magnitude crustal extension and tectonic exhumation of the eastern Walker Lane (Stewart, 1983; Snow and Wernicke, 2000; Ferrill et al., 2012b). Owens Valley marks the western edge and most active portion of the Walker Lane belt along the eastern Sierra Nevada range (Dixon et al., 2000, 2003; Oldow, 2003; Fig. 1). In addition to the active extension, the eastern Sierra Nevada is also the site of extensive Quaternary felsic volcanism, with the largest volcanic event represented by the eruption of the Bishop Tuff and associated magma chamber collapse to form the Long Valley caldera (Fig. 1). The eruptive and cooling history of the Bishop Tuff exposed in the Volcanic Tableland in northern Owens Valley, California, has been studied extensively for eight decades (e.g., Gilbert, 1938; Bate- man, 1965; Sheridan, 1970; Ragan and Sheri- dan, 1972; Hildreth, 1979; Hildreth and Mahood, 1986; Wilson and Hildreth, 1997, 1998, 2003).

In recent years, the structurally controlled morphology of the top of the welded ignimbrite and fault scarps that offset this horizon (Fig. 1) have been the subject of study as a natural laboratory for understanding normal faulting processes (e.g., Dawers et al., 1993; Dawers and Anders, 1995; Pinter, 1995; Pinter and Keller, 1995; Willemse et al., 1996; Willemse, 1997; Ferrill et al., 1999b; Ferrill and Morris, 2001; Sims et al., 2005; Phillips and Majkowski, 2011; Lovely et al., 2012). Superb exposure, relatively rapid and active deformation, and the deposition and subsequent partial erosion of the Pleistocene-age Bishop Tuff as a key marker horizon make this an excellent field laboratory for investigations into the structure and neotectonic evolution of an actively forming continental transtensional basin.

Faulting in the Volcanic Tableland has been investigated to: (i) characterize normal fault displacement-length scaling relationships for faults and fault segments (Dawers et al., 1993; Dawers and Anders, 1995); (ii) test linear-elastic boundary element modeling of fault displacement patterns and related fault block geometry (Willemse et al., 1996; Willemse, 1997; Lovely et al., 2012); (iii) understand the formation, deformation, and breaching of relay ramps and resulting development of fault corrugation (Ferrill et al., 1999b; Ferrill and Morris, 2001); (iv) analyze fluvial channel response to relay-ramp-bound- ing fault displacements (Hopkins and Dawers, 2015); and (v) characterize the development of fault network connectivity (Sims et al., 2005). In

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addition, investigations have been performed to understand fault interaction, fault zone deformation processes, and influences of these processes on permeability within the Bishop Tuff and the underlying nonwelded tuff and volcaniclastic sediments (Ferrill et al., 2000, 2009; Evans and Bradbury, 2004; Dinwiddie et al., 2006, 2012; McGinnis et al., 2009).

Brittle deformation—in the form of cooling joints, fumarolic joints, and faults—of volcanic deposits generally begins coincident with emplacement, welding, and cooling (e.g., Dunne et al., 2003). Lateral thickness variations of an ignimbrite deposit can cause differential compaction where thicker sections remain hotter for longer periods and compact more than thinner sections because of their higher total heat capacity. Thickness variations create a lateral stress gradient that can result in brittle deformation, such as extension fracturing or normal faulting (e.g., Hildreth, 1983; Keith, 1991). For the general case of an ignimbrite deposit, Sheridan (1970) proposed that vertical joint formation could be driven by differential compaction between the depocenter, where thick deposits would experience more compaction, and the basin margins, where thin deposits would experience less compaction. Although the Bishop Tuff was likely subject to some of the effects of differential compaction and basin geometry, the most apparent meso- to macro-scale structural fabric is that of a large number of normal faults accommodating predominantly east-west extension. In this paper, we document fault segment interaction and linkage, strain localization and geometry within relay ramps, fault scarp morphology, and potential timing of fault zone development. These observations provide new insight into faulting processes within the Volcanic Tableland.

BACKGROUND ON STRUCTURAL AND TECTONIC SETTING OF THE VOLCANIC TABLELAND

At the latitude of the Volcanic Tableland, Owens Valley has the overall geometry of a graben. Crustal-scale range-front normal faults are present on both sides of the valley—the Sierra Nevada range-front fault system (a left-stepping, en echelon array) to the west and the White Mountain fault to the east (Fig. 1). Based on topographic relief of the bounding range fronts and thickness of basin fill above basement, each of these faults has on the order of 3 to >4 km of normal throw (Bateman, 1965; Lueddecke et al., 1998; Phillips and Majkowski, 2011).

The Volcanic Tableland is composed of the 758.9 ± 1.8 ka Bishop Tuff (van den Bogaard and Schirnick, 1995; Sarna-Wojcicki et al., 2000), which overlies reworked volcaniclastic and other sedimentary deposits. The Bishop Tuff is mechanically layered with air-fall deposits and variably welded ignimbrites that are interpreted to have been emplaced in less than 1 wk (Wilson and Hildreth, 1997). The entire Bishop Tuff sequence can be considered as one heterogeneous growth unit within the overall sedimentary and volcanic fill sequence of the basin (Hollett et al., 1991; Phillips and Majkowski, 2011). Most of the Volcanic Tableland is capped by a welded ignimbrite, unit Ig2E of Wilson and Hildreth (1997), which represents one of two major ignimbrite units in the Volcanic Tableland. Nine associated Bishop air-fall units also have been identified and characterized (Wilson and Hildreth, 1997).

Formation of the Owens Valley graben is generally thought to have begun in the Pliocene.
Epoch, with initiation between 2 and 6 m.y. ago (Bachman, 1978; Pinter and Keller, 1995; Lueddecke et al., 1998; Phillips and Majkowski, 2011). The deformation history recorded by faults since Bishop Tuff emplacement at 759 ka therefore represents only 13%–38% of the temporal extensional history of the basin. As noted by Wilson and Hildreth (1997), the White Mountain fault (e.g., Ferrill and Morris, 1999) would have been subsequently buried by the ignimbrites. These preexisting structures are important because they are likely to have been reactivated during post–Bishop Tuff extension, influencing the present-day fault pattern (Bateman, 1965). Faults in the Volcanic Tableland range in displacement from <1 m to ~15 m. Faults are dominantly north-south striking (Fig. 2, inset), with approximately equal numbers of east-dipping and west-dipping faults, which were mapped using side-looking airborne radar data (see Ferrill et al., 1999b). Present-day topography of the eastern part of the Volcanic Tableland shows increasing dip or rollover into the Fish Slough fault, the largest fault cutting the Volcanic Tableland with a maximum throw of ~145 m (Ferrill et al., 1999b). The footwall of the west-dipping Fish Slough fault is tilted east toward the White Mountain fault (Phillips and Majkowski, 2011) and is itself cut by numerous east- and west-dipping (subordinate) normal faults. The overall architecture of the Volcanic Tableland, including the eastward tilt or rollover, and the associated crustal graben (Ferrill et al., 1999b), is interpreted to be the product of outer-arc extension (Pinter, 1995) as a hanging-wall response to slip on the listric Chalk Bluff fault (Phillips and Majkowski, 2011) and is cut by numerous east- and west-dipping (subordinate) normal faults. The overall architecture of the Volcanic Tableland, including the eastward tilt or rollover, and the associated crustal graben (Ferrill et al., 1999b), is interpreted to be the product of outer-arc extension (Pinter, 1995) as a hanging-wall response to slip on the listric White Mountain fault (e.g., Ferrill and Morris, 1997; Morris and Ferrill, 1999).

Owens Valley and the Volcanic Tableland are in the rain shadow of the Sierra Nevada range. The Owens River, which drains the eastern Sierra Nevada, has cut three terraces above the modern Owens River floodplain (Bateman, 1965; Pinter et al., 1994). These terraces have been correlated to glacial moraines in the eastern Sierra Nevada, where dating of glacial stages has been performed (Pinter et al., 1994). Although faults are best exposed cutting the welded Bishop Tuff across the Volcanic Tableland surface, faults also are present cutting younger erosional Owens River terraces south of the southern escarpment of the Volcanic Tableland along Chalk Bluff (Pinter, 1995; Pinter and Keller, 1995).

Analysis of faulting of these terraces provides incremental constraints on the history of tilting and faulting. Topographic analysis by Sheehan (2007) showed that displacements and related extension magnitudes are greatest for faults cutting the Bishop Tuff, and progressively less for the younger Owens River terraces. Owens River terraces have been tilted to the east and faulted south of Chalk Bluff (Bateman, 1965; Pinter and Keller, 1995). Bateman (1965) analyzed orientations of present and past Owens River floodplain terraces and found that (i) the present floodplain of the Owens River slopes east at an average rate of 5.7 m/km (30 ft/mi), (ii) the lowest terrace (Q2 or Tioga age, 9.8–26 ka; according to Pinter et al., 1994) slopes east at an average rate of 7.6 m/km (40 ft/mi), and (iii) the next higher middle terrace (Q3 or Tahoe age, 53–119 ka; according to Pinter et al., 1994) slopes east at an average rate of 13.2 m/km (70 ft/mi). An older and higher terrace (Q4 or pre-Tahoe age, 131–463 ka; according to Pinter et al., 1994) has also been tilted eastward, as has the paleochannel network on the Volcanic Tableland. Pinter and Keller (1995) used angular terrace and paleochannel tilt magnitudes to estimate rates of eastward tilting in the range of 3.5–6.1°/m.y. This eastward tilting likely reflects continued slip on, and hanging-wall tilting toward, the basin-bounding White Mountain fault. Chalk Bluff, which trends approximately east-west, developed as an erosional escarpment due to lateral cutting and downcutting by the Owens River flowing in the gradient direction east across the valley toward the active range-front fault along the White Mountains.

In addition to evidence for long-term post–Bishop Tuff fault displacement, there is also clear evidence that the fault system remains active. For example, slip on several faults in the eastern part of the Volcanic Tableland was triggered in association with the 21 July 1986 M 6.2 Chalfant earthquake (Cockerham and Corbett, 1987; Smith and Priestley, 2000). The Chalfant earthquake was centered at a depth of 11.5 km on the White Mountain frontal fault zone (Cockerham, 1986). The earthquake produced two systems of surface ruptures that were mapped for a distance of 4 km after the earthquake (Lienkaemper et al., 1987). One set of ruptures was mapped for a distance of 10.5–13.2 km along an azimuth of 350° along the White Mountain frontal fault zone southeast of the epicenter of the Chalfant main shock (black lines, Fig. 2). Slip indicators along these ruptures suggested primarily right-lateral strike-slip displacement with a small dip-slip (normal) component, maximum displacement on individual ruptures of 4.6 cm, and maximum summed displacement of 11.1 cm over the width of the zone (for details, see Lienkaemper et al., 1987).

The other system of surface ruptures associated with the 1986 Chalfant earthquake developed 5–15 km west of the White Mountain frontal fault zone within the east-dipping part of the Volcanic Tableland (black lines, Fig. 2). This system of ruptures was distributed over a zone as much as 10 km wide, trending generally north-south for a distance of ~30 km. These ruptures showed dilation perpendicular to the fracture surface and in some cases vertical offset. Dilational fractures in many cases showed drainage of sand into the fractures. Fractures were observed to have apertures of as much as 1–2 cm, summing to at least 5 cm and possibly as much as 15 cm (Lienkaemper et al., 1987). While no direct evidence of strike-slip motion was observed, it could not be ruled out because of the nature of the loose material and the lack of definitive piercing points or slip indicators. Fractures tended to form in left-stepping arrangements, with some occurring on mapped fault scarps and others occurring between mapped faults (Lienkaemper et al., 1987).

**METHODOLOGY**

Since 1996, we have conducted field and remote-sensing investigations of faulting in the Volcanic Tableland. These investigations have included mapping of faults from side-looking airborne radar data, mapping on foot using real-time kinematic differential global positioning system (RTK-DGPS) surveying, interpretation and analysis of airborne light detection and ranging (LiDAR) data (acquired by Chevron Corporation), analysis of digital elevation model (DEM) data and aerial photography, analysis of low-altitude and low-sun-angle aerial photography, and undertaking our own field observations. For the RTK-DGPS mapping, the horizontal and vertical position accuracy was 20 cm and 60 cm, respectively, for two rover units. A third rover unit yielded 2 cm and 6 cm horizontal and vertical position accuracy, respectively. Structure contour and slope maps were produced using the irregular network of field data and interpolated to a uniform grid spacing of 5 m. A denser data set with smaller data spacing was collected for the southern relay ramp using the third rover unit to enable analysis of small-displacement faults (1 m to 5 m throw). These data were used to generate a contour map of the southern relay ramp with 1 m grid spacing; the small faults in the relay ramp were then mapped and measured in the field using a calculated slope map as the base map, and fault throws were measured in the field using a handheld measuring tape. Although much of this work has served to confirm the work of many others, we have reached a number of new interpretations that are the focus in this paper.
Figure 2. Volcanic Tableland map with faults (red traces are west-dipping faults; blue traces are east-dipping faults) mapped from side-looking airborne radar data (cf. Ferrill et al., 1999b), channels (dashed black lines; Pinter and Keller, 1995), 1986 Chalfant earthquake ruptures (black lines; Lienkaemper et al., 1987), and fault rose diagram based on mapped fault traces (inset at bottom left). Star represents epicenter of Chalfant earthquake. The approximately east-west escarpment north of the Owens River is the erosional Chalk Bluff escarpment. Universal Transverse Mercator (UTM) zone 11, North American Datum (NAD) 83.
Formation of cooling joints early in the deposition of the Bishop Tuff created a preexisting fracture framework that is blocky and irregular (Sheridan, 1970). Fault scarps in the capping unit of the Volcanic Tableland appear rubbly rather than as smooth, planar, slickensided fault surfaces. During field mapping, we observed that many of the faults lose displacement laterally to fault tips, but that monoclinal folds persist beyond the fault tips. In some locations, the top of the welded Bishop Tuff is exposed, and cooling joints are visible (Figs. 3A and 3B). Along Chalk Bluff where the Bishop Tuff is exposed in profile, welding gradation is visible from poorly welded near the base to moderately welded in the cap-rock unit. The poorly welded portion of the Bishop Tuff generally lacks cooling joints, whereas cooling joints are consistently present in the cap-rock unit, with patterns ranging from polygonal network to two nonorthogonal sets described by Sheridan (1970) as conjugate fracture sets. Where the Chalk Cove fault cuts the lower part of the poorly welded Bishop Tuff, the fault zone is extremely narrow and widens upward with a tectonic fracture zone (Dinwiddie et al., 2006). With increasing monocline dip in the cap-rock unit (Fig. 3C), and with the appearance of an actual fault gap, cooling joints that parallel the overall fault trend separate, and cooling-joint-bounded blocks tilt and topple (e.g., Adhikary et al., 1997). Because these fault scarps are developing at the ground surface, in lithified rock with cooling joints already present, the fault rupture in the welded cap rock forms by reactivation of nearly ubiquitous preexisting cooling joints. Precursor monoclinal folding dilates the preexisting cooling joints, and there is little frictional contact between blocks, resulting in limited areas of frictional sliding where slickenlines would be likely to develop (Fig. 3C). Consequently, slickenlines are extremely rare on exposed faults in the Volcanic Tableland. We conclude that these faults scarp in the welded ignimbrite never existed as simple topographic offsets, like those typically assumed in diffusion modeling of slope degradation developed for fault scarp in unconsolidated sediment (e.g., Avouac, 1993; Avouac and Peltzer, 1993); thus, such diffusion models (even accounting for differences in cohesion) are not directly applicable to the Volcanic Tableland fault scarps.

Faults in the underlying nonwelded Bishop Tuff show development of a narrow fault zone, with cataclasis and slickenline development consistent with shear failure in rock without a preexisting fracture fabric to reactivate (Evans and Bradbury, 2004; Dinwiddie et al., 2006, 2012; McGinnis et al., 2009). Propagating as shear fractures through an unfractured medium, these faults developed as thin cataclastic shear zones with associated grain cataclasis and pore collapse. Although the rubbly fault scarp appearance on the Volcanic Tableland may appear at first to be a function of weathering, we conclude that the faults instead initiated and are continuing to develop as subaerial rubbly scarps as part of the normal progression of deformation.

Fault systems in the Volcanic Tableland that have throws of 20–145 m are all made up of multiple fault segments. Upon close inspection, it is evident that many individual fault segments themselves are the result of linkage of multiple closely spaced fault segments (Fig. 4). The scarps contain evidence of minor breached relay
ramps (Fig. 4). Fault growth by segment linkage is the norm rather than the exception, and faults in the Volcanic Tableland at all observable scales show evidence of segmentation or linkage of segments (Figs. 4A and 4B).

Faults nucleate, grow by propagation, intersect, and coalesce to form larger faults that are composites of numerous linked segments, and this is borne out by natural examples (Dawers and Anders, 1995; Ferrill et al., 1999a, 1999b; Ferrill and Morris, 2001; Manighetti et al., 2001, 2015; Soliva and Benedicto, 2004; Childs et al., 2009) and physical analog modeling (e.g., Ackermann et al., 2001; Sims et al., 2005; Schlaggenhauf et al., 2008; Wyrick et al., 2011; Fig. 4C). An inherent aspect of this process is fault segmentation, often with a relatively consistent sense of stepping that produces an en echelon arrangement of segments. These segments may originate as separate en echelon faults or en echelon branches from single faults. Large normal faults develop by breakthrough of relay ramps between overlapping fault segments. This breakthrough can occur by (i) curved lateral propagation of overlapping fault tips or (ii) connecting fault formation (Peacock and Sanderson, 1994; Childs et al., 1995; Ferrill et al., 1999b). In the Volcanic Tableland, linkage of segments occurs most commonly by curved lateral propagation (Ferrill et al., 1999b). The resulting fault surfaces formed by linkage of segments tend to be corrugated, with cusps and asperities located within or at the top and/or bottom of breached relay ramps (Ferrill et al., 1999b), with more pronounced corrugations being produced where the angle of intersection between linking faults approaches 90°, and where the length of the ramp-breaching fault becomes large. Relay ramps experience deformation related to both regional extension and local deformation caused by displacement gradients (lateral and vertical) on overlapping normal faults that bound the ramp. Displacement gradients along overlapping faults produce (i) vertical-axis rotation responding to fault heave gradient (clockwise rotation in left-stepping fault arrays; counterclockwise in right-stepping arrays); (ii) tilting in the fault strike direction accommodating fault throw gradient, perhaps with a component toward or away from the footwall related to fault dip change at depth and/or fault block deformation (i.e., monocline formation); and (iii) relay-ramp extension parallel to fault strike (Ferrill and Morris, 2001).

The Fish Slough fault has the largest displacement of faults cutting the Bishop Tuff in the Volcanic Tableland with a maximum throw of ~145 m (Ferrill et al., 1999b). This fault system also contains some of the most dramatic fault system structures in the Volcanic Tableland, including the steepest relay ramp dips, and, because of the large displacement, a more-evolved stage of fault system development (Fig. 5A). To explore the fault system in detail, an RTK-DGPS system was used to produce high-resolution topographic maps of the top of the Bishop Tuff welded unit to map the fault blocks and fault gaps of the Fish Slough fault system (Ferrill et al., 1999a). Because of the relative lack of weathering and erosion of the capping unit of the Bishop Tuff, these maps represent structure contour maps of the fault system (Fig. 5B) and were used to calculate slope maps (Fig. 5C). Higher-resolution data were used as the base map of the southern relay ramp, for analysis of upthrown and downthrown cutoffs of normal fault scarps with faults with <1 m to >5 m throw (Fig. 6). The mapping demonstrated that many of the faults lose displacement laterally to fault tips, but that monoclinic folds persist beyond the fault tips.

**DEFORMATION WITHIN A RELAY RAMP**

Based on the geometry of the Fish Slough fault system (Figs. 5B, 5C, 6, and 7A), including hanging-wall and footwall cutoff lines and the associated fault blocks, we interpret the following sequence of deformation:

1. Formation of left-stepping, en echelon array or en echelon branching fault (Fig. 7B).

2. Increasing displacement gradient along overlapping fault tips producing clockwise vertical-axis rotation (accommodating lateral fault heave gradient), ramp tilting (accommodating lateral fault throw gradient), and small-scale faulting within relay ramps (Fig. 7B).

3. Curved lateral propagation leading to fault linkage and breakthrough along the upthrown (footwall) fault traces, resulting in a corrugated fault surface with cusps and corrugations along fault segment intersections (Fig. 7C). Based on larger displacement on the through-going fault at the top of the two breached relay ramps, we interpret that the northern ramp broke through first, after ~50% of the total (present-day) fault displacement, followed by breakthrough of the southern ramp at ~80% of the total fault displacement.

4. Cutoff of southern cuspatte corrugation, which smoothed the fault surface and produced a triangular prism-shaped fault block (cusp faulted from footwall) between the top of the relay ramp and footwall (Fig. 7D).

Curved fault tips or connecting faults— inconsistent with the regional fault trend—developed in the ideal fault orientation within the local stress field at the time of their formation associated with the propagating fault tips. Once fault linkage occurred, curved fault segments that formed in ideal orientations within locally perturbed stress fields prior to fault linkage were no longer well oriented for slip in the regional stress field. Shortcut faults developed to sever asperities, straighten the fault trace, and smooth the fault sliding surface, and sometimes led to osculating fault zones (Fig. 7).

Detailed mapping of the southern breached relay ramp revealed that the overall shape of the ramp includes a northern domain that contains a south-plunging anticline represented by a south-dipping western portion and a southeast-dipping eastern portion, and a southern domain that is a gentle south-dipping homocl ine. The northern domain of the relay ramp is cut by 35 normal faults (Fig. 6). These faults include 21 east-dipping faults (antithetic to the relay-ramp-bounding faults) and 14 west-dipping faults (synthetic to the relay-ramp-bounding faults), with maximum throws ranging from <1 m to 4 m (Fig. 6). These faults include segmented fault systems with relay ramps, breached relay ramps,
and horsts and grabens. Dominant fault strike within the relay ramp is north-south, which (i) parallels the dominant fault trend in the Volcanic Tableland; (ii) parallels the axis of the fold within the northern domain of the relay ramp; (iii) is consistent with an outer-arc extension bending strain along the fold axis; and (iv) is oblique to the ramp-bounding faults but in an appropriate orientation to accommodate localized ramp extension in response to displacement gradients developed on the overlapping ramp-bounding faults (Ferrill et al., 1999b; Ferrill and Morris, 2001). Faulting within this relay ramp is more intensely developed, in terms of fault spacings and displacements, compared with faulting in the tableland across Fish Slough to the west. Thus, the faulting within the ramp is not typical of the Volcanic Tableland, but is unique to the breached relay ramp. Although deformation within relay ramps is consistent with the “regional” extension direction, the anomalous spacing and displacement indicate that these are unique to the ramp, accommodating (i) distributed regional extension between overlapping fault tips to balance displacement deficits on the overlapping and fault tips; (ii) localized ramp extension associated with tilting and vertical-axis rotation on the ramp-bounding faults; or (iii) outer-arc extension bending strain along the fold axis within the ramp, with the fold being the product of slip on a listric fault beneath the ramp and possibly monoclinal folding associated with fault propagation. It is possible that one or all of these mechanisms was/were active within the relay ramp as it was developing and ultimately breached. Because of the lack of kinematic indicators in the form of slickenlines or other incremental deformation indicators, it is not possible to reach a unique interpretation from the final structural geometry.

ANALYSIS OF FAULTED CHANNELS

Channels incised into the welded Bishop Tuff on the Volcanic Tableland surface were recognized and discussed by Bateman (1965), and mapped by Pinter and Keller (1995). Bateman (1965) described three channels crossing the Volcanic Tableland that have been faulted. These channels are represented by distinctive, V-shaped notches along Chalk Bluff. The middle of these three channels, the Happy Boulders Channel, is particularly well exposed and is displaced down to both the east and west across faults, beheading the drainage source for this channel in its present configuration. Bateman (1965) noted that the channels generally trend from northwest to southeast and reflect an early Pleistocene drainage pattern down the primary southeastward dip of the Volcanic Tableland surface. The long northwest to southeast channels are cut and offset by normal faults; however, we observed that other channels drain into or out of fault-controlled depressions, apparently reacting to the structurally controlled surface morphology of the Volcanic Tableland and indicating formation or activity during or after faulting.

An airborne LiDAR data set was collected in 2008 over an 8.1 × 8.9 km area in the southeast...
Faulting in the Volcanic Tableland

Figure 6. Detail of faulting within Fish Slough fault relay ramp. Black lines with bar and ball are west-dipping faults, and red lines with bar and ball are east-dipping normal faults. Cross section, W-E, illustrates bending strain within the relay ramp. V:H—vertical:horizontal.
part of the Volcanic Tableland with 2 m pixel resolution (Lovely et al., 2012); this data set is shown as a gray-scale shaded relief map in Figure 8, and as a three-dimensional model in perspective view in Figure 9 (data provided for research purposes courtesy of Chevron Corporation). The LiDAR data show several geomorphologic interrelationships between faulting and fluvial geomorphologic features that provide different relative timing constraints for faulting with respect to fluvial processes (Figs. 8 and 9). These include (i) a north-south–striking, east-dipping normal fault that cuts and offsets a paleo–Owens River terrace (Q3; Pinter et al., 1994) in the southeast corner of the Volcanic Tableland by 4–4.5 m down to the east (Fig. 8C); (ii) an incised channel that drains southeastward from a structurally controlled closed depression in the crested graben (Figs. 8D and 9C); (iii) an incised channel that drains from the west into a structurally controlled closed depression (graben) near the southern edge of the Volcanic Tableland (Figs. 8B and 9B); and (iv) an incised channel that has been offset by east- and west-dipping faults, i.e., the Happy Boulders Channel (Figs. 9, 10, and 11).

Faulted stream channels provide ideal piercing points at the intersection of the channel axis with the fault plane to determine fault displacement direction in the fault plane. The fault offsets of channels consistently show normal displacement across faults, ranging from 1 to ~30 m (Figs. 11A and 11B). Within the resolution of the data, there is no evidence of strike-slip displacement across any of the faults.

Analysis of the Happy Boulders Channel profile demonstrates that the channel is dropped to the west and to the east across faults (Figs. 11A, 11B, and 11C). This indicates that fault displacement ultimately outpaced channel incision. A comparison of channel-axis topographic profiles with topographic profiles collected on top of the welded horizon immediately adjacent to the channel along both the north and south margins on the Volcanic Tableland surface shows that channel incision depth is variable (Figs. 11A and 11B). The most dramatic examples are (i) associated with Chalk Bluff and (ii) associated with two east-dipping normal faults (faults 3 and 4 in Fig. 11B). Along the length of the mapped channel, the channel depth with respect to the adjacent welded tuff horizon is nowhere deeper than 10 m, except where the channel approaches Chalk Bluff. Over a channel distance of 1300 m, the channel deepens from 1 m to 25 m at Chalk Bluff (Fig. 11B). The other case where the channel depth locally expresses a significant change is associated with two east-dipping normal faults with throws of 30 m where the channel is offset (Figs. 11A

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**Figure 7.** (A) Three-dimensional model of Bishop Tuff cut by southern Fish Slough fault system viewed looking down and to the northeast (vertical exaggeration = 2x). (B–D) Interpreted sequence of development of the southern Fish Slough fault system from original en echelon arrangement of faults. (B) Connection of fault segments by curved lateral propagation along upthrown fault surface. (C) Continued displacement and (D) cutoff of southern footwall cusp. Figure is after Ferrill et al. (1999a).
and 11B). At both scarps (faults 3 and 4), the channel depth in the lower two thirds of the scarp approaches 0 m and at the top of the scarp approaches 10 m. This observation suggests that incision was able to keep pace with the first 10 m of fault displacement, after which fault displacement continued without commensurate channel incision. Thus, the amount of channel incision varies from fault block to fault block along the profile, which indicates that channel incision was active during the period of faulting, and incision was able to keep pace with faulting for some period.

**DISCUSSION**

**En Echelon Faulting**

En echelon faulting is often attributed to strike-slip tectonics, but both natural observations and analog modeling experiments reveal that this pattern can also be due to (i) fault growth during simple extension or contraction, (ii) change in displacement direction with time, or (iii) reactivation of a preexisting fault by oblique displacement (e.g., Clifton et al., 2000; Davis et al., 2005; Otsuki and Dilov, 2005; Giba et al., 2012). Evidence from the faulting in the Volcanic Tableland indicates that faults have normal dip-slip displacement rather than strike-slip displacement (Dawers et al., 1993; this study). There is, however, evidence from regional GPS-based crustal displacement studies that indicates an overall transtensional deformation regime for Owens Valley and the surrounding region (Reheis and Dixon, 1996; Dixon et al., 2000, 2003; Oldow, 2003; Phillips and Majkowski, 2011). The Chalfant Valley earthquake (1986, M 6.4; see Fig. 2 for location), which is interpreted to have occurred on the White Mountain frontal fault and to have had 11 cm of right-lateral slip (Lienkaemper et al., 1987; Phillips and Majkowski, 2011) and created two sets of surface ruptures (Lienkaemper et al., 1987), attests to the partitioning of strike-slip and extensional deformation between different components of the Owens Valley graben system. We attribute the majority of the fault displacement in the Volcanic Tableland (as distinct from the larger Owens Valley system) to outer-arc extension of the half-graben geometry of the present-day Owens Valley floor. This geometry is primarily the product of the normal-slip component of regional displacement and is essentially decoupled from the strike-slip component, which is confined to the White Mountain frontal fault. It is likely that these faults formed by initiation or reactivation along a structural element oblique to (rather than perpendicular to) the active extension direction.

**Figure 8.** (A) Shaded relief map of Volcanic Tableland derived from light detection and ranging (LiDAR) data (Universal Transverse Mercator [UTM] zone 11, North American Datum [NAD] 83) showing locations of B, C, and D to illustrate details of geomorphologic features showing relationships among faults, terrace deposits, and channels. Numbers along margins are northing and easting in meters. (B) Channel crossing three east-dipping normal fault scarps descending from west to east into graben. (C) Fault scarp cutting and offsetting (down to the east) terrace by 4.25 m. (D) Channel draining closed depression in crestal graben. This drainage occurs at topographic low point (between footwall highs) between southern tip of west-dipping normal fault to the north and northern tip of east-dipping normal fault to the south.
Figure 9. (A) Oblique perspective view of light detection and ranging (LiDAR) image of the southeastern part of the Volcanic Tableland, illustrating faults, fumarole mounds, grabens and half grabens, and incised channels. This digital elevation model spans the entire domain shown in Figure 8A. (B) Three-dimensional perspective view of channel draining into graben-controlled topographic depression illustrated by Figure 8B. (C) Three-dimensional perspective view of channel draining closed depression in crestal graben that is shown in Figure 8D. Vertical exaggeration = 4x.
Fault Zone Width

Fault zones are generally localized zones of deformation that may include rotation of layering, fracturing, cataclasis, dilation, dissolution, and mineral precipitation. Fault zones typically include a fault core (a narrow zone where displacement is concentrated) and a surrounding damage zone that is typically wider, representing distributed deformation (e.g., Caine et al., 1996; Evans et al., 1997; Mitchell and Faulkner, 2009; Faulkner et al., 2010, 2011). Research over the last three decades has explored the question of whether fault zone width grows proportionally as a simple function of displacement (Robertson, 1982; Evans, 1990; Shipton et al., 2006; Wibberley et al., 2008; Childs et al., 2009). In practice, determining the width of a damage zone can be difficult because the basis for definition varies with deformation mechanisms, for example, deformation dominated by fractures (e.g., Faulkner et al., 2011; Savage and Brodsky, 2011) versus deformation dominated by synthetic dip and conjugate faults (e.g., Ferrill et al., 2011). Along a given fault and comparing faults with similar displacements, fault zone widths commonly vary over several orders of magnitude (Evans, 1990; Shipton et al., 2006). Geodetic and seismologic data can indicate significantly larger damage zone widths than geologic evidence alone (Cochran et al., 2009).

Based on detailed investigation of fault network displacement patterns and rupture history on faults, Nicol et al. (2010) concluded that fault interactions are ubiquitous such that no fault can be considered completely isolated from all other faults. According to Groshev (1988, p. 1341), “A fault zone is a tabular region across which the displacement parallel to the zone is appreciably greater than the width of the zone.” For a structure to be recognized as a fault zone by this definition, it must meet the criterion of having displacement in excess of the fault zone width. Consequently, plots of fault zone thickness or width versus displacement will tend to inherently illustrate increasing thickness with increasing displacement, and this is true of the early investigations of these relationships (Otsuki, 1978; Robertson, 1982; Evans, 1990). Studies that show relatively monotonic increase in fault zone width with displacement are rare and narrowly limited in scope (Shipton and Cowie, 2001; Flodin et al., 2005).

Fault zone width or thickness is strongly controlled by the early history of fault nucleation and evolution during propagation and linkage of fault segments (Wibberley et al., 2008; Childs et al., 2009). The width of a normal-fault–related fold is in many cases the primary control on maximum fault zone (or damage zone) width. This monocline width is determined at the onset of folding ahead of the propagating fault related to the mechanical stratigraphy—in particular, thickness and competence of the involved layers—rather than a simple function of fault displacement (Ferrill et al., 2007, 2012a).

Another primary determinant of fault zone width is the spacing between overlapping fault segments (or width of relay structures; Childs et al., 2009; de Joussineau and Aydin, 2009) and “splays” (e.g., Perrin et al., 2016) that commonly cooperate to define a fault zone. As displacement increases, it may tend to localize into a narrower zone, straightening and smoothing the fault surface by severing asperities that are poorly oriented for slip in the ambient stress field (Engelder, 1978; Power et al., 1988; Childs et al., 1996). This has been referred to as “strain softening” behavior (Gray et al., 2005). The width of the segmented fault array is established early, and therefore this control on damage zone width is not linearly related to fault displacement. However, analog modeling and field observations show that fault systems develop in displacement versus length space along a stair-step path rather than a linear self-similar path, due to periodic jumps in fault length when fault segments link (Ferrill et al., 1999b; Childs et al., 2009; Wyrick et al., 2011). With increasing displacement, fault zones with increasingly wide separation begin to cooperate and link (Ferrill et al., 1999b; Childs et al., 2009). Thus, fault zone width also grows along a stair-step trajectory with respect to displacement, with the widening steps again associated with cooperation and linkage (Childs et al., 2009).

In the moderately welded Bishop Tuff “cap rock” in the Volcanic Tableland, the fault core is obscured by the rubblized fault scarp and is not clearly exposed. The fault core, however, is exposed in fault exposures within the underlying poorly welded Bishop Tuff (Dinwiddie et al., 2006). As described earlier in this article, the rubbly character of fault scarps in the
Volcanic Tableland is the product of (i) upward fault propagation toward the surface, which develops a monocline in moderately welded ignimbrite cap rock; (ii) presence and reactivation of cooling joints in the capping moderately welded ignimbrite; (iii) fault initiation as numerous closely spaced en echelon fault segments that bound relay ramps; (iv) segment linkage to breach relay ramps and straighten or smooth fault surfaces; and (v) toppling and sliding of cooling-joint-bounded blocks of welded ignimbrite. Although toppling and sliding of blocks are expected only at the ground surface, other damage zone elements observed on the Volcanic Tableland are expected to occur at depth, including faulted monocline (with reactivated, dilated cooling joints), breached relay ramps, relict fault tips complete with their associated damage zones, and shortcut faults that sever fault asperities.

Recognizing that large-displacement faults grow from smaller-displacement, segmented fault systems, it is clear that fault zone widths in the Volcanic Tableland are established during the early stages of fault system development, prior to the establishment of a through-going fault system. As displacement continues to accumulate, fault displacement will tend to localize into a narrow, smoother plane that is closer to being ideally oriented in the active stress field, and that is progressively easier to reactivate, i.e., exhibiting strain softening behavior.

This observation of increasing fault zone localization is in contradiction to some published interpretations that fault zone width grows in proportion to fault displacement (see review by Evans, 1990), and this is well documented in the Navajo Sandstone, where the primary fault zone process was cataclastic shear band development (Shipton and Cowie, 2001). Grain cataclasis led to decreases in grain and pore size, and increases in grain contact area, leading to an overall strain hardening behavior. With increasing displacement, shear bands would form and lock up, and a new shear band would form in close proximity. This “strain hardening” behavior leads to progressive widening of the fault zone with increasing displacement, followed ultimately by strain softening behavior and localization of a narrow fault core within the strain-hardened material. Previously, de Joussineau and Aydin (2007) and de Joussineau et al. (2007) documented significant complexity with respect to fault zone width associated with strike-slip faulting in porous sandstone. This complexity is likely related to a combination of factors, including the strain hardening behavior and progressive fault zone widening that are characteristic of relatively small-displacement faults represented by cataclastic fault zones developed in porous sandstone (Antonellini and Aydin, 1995; Flodin et al., 2005; Shipton and Cowie, 2001; Torabi and Berg, 2011).

With increasing fault system displacement, more widely spaced (separated) fault segments can link by relay ramp breaching, resulting in episodic jumps in fault zone width. This observation is consistent with the conclusions of Childs et al. (2009), i.e., that normal fault zone thickness is strongly controlled by the initial spacing.

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**Figure 11. Happy Boulders Channel:** (A) Channel thalweg profile and adjacent north rim and south rim profiles derived from light detection and ranging (LiDAR) data where available (0–3200 m distance along channel), and digital elevation data further northwest (3200–7400 m distance along profile). (B) Enlargement of LiDAR-derived channel thalweg profile and adjacent north rim and south rim profiles, compared with levels of Owens River terraces Q1, Q2, Q3, and Q4 from Pinter et al. (1994). Note that downstream end of incised channel grades into Q3. (C) Three-dimensional perspective view of channel extracted from LiDAR data showing Chalk Bluff and scarps of faults 1, 2, 3, and 4.
between fault segments within a segmented array. Similarly, data presented by de Joussineau and Aydin (2009) from strike-slip fault systems show that step width grows in proportion to maximum fault offset.

Timing of Faulting

The timing of formation of the faults cutting the Bishop Tuff in the Volcanic Tableland is important to understanding Quaternary deformation history of northern Owens Valley. Analysis of crosscutting relationships shows that faults deform the moderately welded pyroclastic deposits; thus, faulting occurred between the time of welding (i.e., 759 ka) and the present. The faulted channels provide additional constraints on the timing of faulting in the Bishop Tuff.

A number of the basic observations regarding channels help to constrain the timing of channel formation with respect to faulting: (i) Channels are incised into welded tuff; (ii) channels are cut by faults; (iii) some channels show differential incision across fault scarps; (iv) some channels were deflected by relay ramp development; and (v) many channels reflect the shape (local inclination direction) of the faulted welded tuff horizon in response to the structural morphology, draining down structurally controlled dip (e.g., Hopkins and Dawers, 2015).

These observations provide the following important relative timing constraints: (i) The observed channels formed after deposition and welding of the Bishop Tuff, although channels may have initiated in nonwelded upper ignimbrite facies; (ii) the channels were active, at least locally, during the same time frame that the fault system has been active; (iii) in some locations and relatively recently, the rate of fault slip has exceeded the rate of channel incision, leading to beheading and abandonment of channels; and (iv) this beheading and abandonment could be due to tectonic or climatic factors, including increased fault slip rate and/or decreased rate of incision due to reduced precipitation and runoff.

To provide timing constraints on channel incision and faulting, we compared the Happy Boulders Channel thalweg profile with the Quaternary terraces along Chalk Bluff that were previously mapped by Pinter et al. (1994). We produced a channel profile using the LiDAR data where coverage exists and the U.S. Geological Survey 30 m DEM beyond (west of) the LiDAR coverage. This channel profile shows that the channel steepens within the V-shaped notch and abruptly steepens southward toward the channel intersection with the modern Owens River floodplain terrace (mapped by Pinter et al. [1994] as Q1, representing a 0 ka surface). The channel does not cross the floodplain and intersect with the active Owens River, and there is no evidence to support a fault scarp parallel to Chalk Bluff to explain the break in slope.

This incised channel can be traced for >7 km across the Volcanic Tableland (Fig. 11A) and ultimately deepens and grades into the elevation of the Q3 Owens River terrace (Fig. 11B), which has been linked to Tahoe-age glaciation in the Sierra Nevada range dated at ca. 53–119 ka (Phillips et al., 1990). This observation leads to the interpretation that this channel was active at ca. 53–119 ka. This channel, however, is no longer in an active configuration for flow, having been cut and vertically offset by west- and east-dipping normal faults that have beheaded the drainage source for the lower part of the channel. The channel is differentially incised across several fault scarps and fault blocks. Offset on both east- and west-dipping faults without commensurate incision provides evidence for fault slip subsequent to 53–119 ka. Some of these fault scarps do not show evidence of differential incision, suggesting that they propagated across the channel after it was incised rather than during incision.

Other channels show different timing with respect to faulting, for example, two channels that are short in length (~1 km; see Figs. 10B, 10D, 11B, and 11C) and reflect essentially present-day fault-controlled structural morphology of the welded Bishop Tuff. Presenting present-day local structural dip direction and draining toward structurally controlled topographic lows, these channels appear to postdate faulting. The fault scarp that cuts and offsets (throw of 4–4.5 m based on airborne LiDAR data) the Q3 terrace at the southeast corner of the Volcanic Tableland provides clear indication of relatively recent activity on the fault (Figs. 8D and 11A). Collectively, these observations and data demonstrate that the Volcanic Tableland fault system has continued to be active through the Pleistocene.

CONCLUSIONS

The rubbly character of fault scarps in the Volcanic Tableland is the combined product of (i) upward fault propagation toward the surface, which develops a monoline in moderately welded ignimbrite facies; (ii) presence and reactivation of cooling joints in the capping moderately welded ignimbrite; (iii) fault initiation as numerous closely spaced en echelon fault segments bounding relay ramps; (iv) segment linkage to breach relay ramps; and (v) toppling and sliding of cooling-joint-bound blocks of welded ignimbrite.

Fault damage zone width is largely governed by the (i) development and lateral spacing between cooperating en echelon fault segments and (ii) widths of fault tip monoclines. Consequently, fault zone width is developed early and at low displacement. As fault displacement increases, segments link by curved lateral propagation of fault tips, developing locally enhanced small-scale faulting in relay ramps, and corrugated fault surfaces with relict fault tips attached to the hanging wall, footwall, or both.

Channels on the Volcanic Tableland include a spectrum of relative ages with respect to faulting, including channel initiation prior to faulting, activity and differential incision during faulting, and development relatively late with respect to fault system development in response to structurally produced geomorphology of the Volcanic Tableland surface. Similar complexity of timing relationships between channels and faulting has been described from the grabens of Canyonlands National Park in southeastern Utah (Trudgill, 2002). One of the major abandoned channels in the Volcanic Tableland was inspected in detail and appears to have continued to have been active through Tahoe-age glaciation, based on the observation that the present channel is “hanging” with respect to modern (Q1) and previous (Q2) Owens River terraces, is at grade with the Tahoe-age (Q3, 53–119 ka) Owens River terrace, and is incised through the pre-Tahoe terrace level (Q4, 131–463 ka).

Faulting within the largest breached relay ramp along the southern segment of the Fish Slough fault zone corresponds to a fold axis within the ramp, which we interpret to have developed as fault block deformation to accommodate bending strain above a listric portion of the underlying fault segment, although faults also may exhibit a distributed component of east-west extension and localized ramp stretching to accommodate relay ramp steepening between two en echelon dip-slip normal faults.

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REFERENCES CITED

Ackermann, R.V., Schlee, R.W., and Withjack, M.O., 2001. The geometric and statistical evolution of normal fault systems: An experimental study of the effects of mechanical layer thickness on scaling laws: Journal of Structural Geology, v. 23, p. 1803–1819, doi:10.1016/S0191-8141(01)00298-1.

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