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Recency of Faulting and Neotectonic Framework in the Dixie Valley Geothermal Field and Other Geothermal Fields of the Basin and Range

Principal Investigator:

Steven G. Wesnousky

Center for Neotectonic Studies

University of Nevada

Reno, NV 89557, USA

steve@seismo.unr.edu

Key Personnel/Co-PI:

S. John Caskey

Department of Geosciences

San Francisco State University

1600 Holloway Ave.

San Francisco, CA 94132, USA

caskey@sfsu.edu

Co-PI:

John W. Bell

Nevada Bureau of Mines and Geology

Mackay School of Mines and Geology

University of Nevada

Reno, NV 89557, USA

EXECUTIVE SUMMARY

We studied the role that earthquake faults play in redistributing stresses within in the earths crust near geothermal fields. The geographic foci of our study were the sites of geothermal plants in Dixie Valley, Beowawe, and Bradys Hot Springs, Nevada. Our initial results show that the past history of earthquakes has redistributed stresses at these 3 sites in a manner to open and maintain fluid pathways critical for geothermal development. The approach developed here during our pilot study provides an inexpensive approach to 1) better define the best locations to site geothermal wells within known geothermal fields and 2) to define the location of yet discovered geothermal fields which are not manifest at the surface by active geothermal springs. More specifically, our investigation shows that induced stress concentrations at the endpoints of normal fault ruptures appear to promote favorable conditions for hydrothermal activity in two ways: 1) Nearby fault ruptures may induce aftershocks associated with small amounts of slip on macro fractures along the fault and within the damage zone of the fault, thereby producing open fractures without producing large stress drops on the fault; and 2) when the geothermal field is adjacent to fault rupture endpoints, decreases in fault-normal stress may be large enough to produce significant increases in fracture dilatancy, thereby increasing hydraulic conductivity. We conclude that an understanding of the spatial distribution of active faults and the past history of earthquakes on those faults be incorporated as a standard tool in geothermal exploration and in the siting of future boreholes in existing geothermal fields.

PROJECT SUMMARY

INTRODUCTION

It is widely accepted that geothermal activity is largely associated with areas of active faulting. Figure 1 illustrates that the spatial correlation between thermal springs and late Quaternary faults in Nevada is quantitatively robust. It is commonly accepted that faults play a role in this association by sometimes forming conduits for upward movement of ground water. Similarly, in geothermal reservoirs, it is thought that it is necessary to have recent fault movements in order to maintain open fractures and permeability. However, the question of how the mechanics of the faulting process may play a causative role in the factors that influence geothermal production remains poorly understood. A better understanding of the faulting processes that create and maintain reservoir permeability can potentially reduce geothermal exploration costs if it can be used to more accurately target successful wells. With this objective in mind we have initiated detailed field studies of active faults in the Dixie Valley, Beowawe, and Bradys geothermal production fields to characterize the distribution and recency of fault movements. To evaluate how the mechanics of the faulting process may influence the geothermal environment, we examine the redistribution of stresses associated with the most recent identifiable fault ruptures in the vicinity of the geothermal fields.

Figure. 1. Number of thermal springs as a function of distance to the nearest late Quaternary fault.

METHOD AND FAULT MODELS

We use the approach of Reasenber and Simpson (1992) and Stein *et al.* (1992), based on the methodology of Okada (1992), to calculate the static stress changes resulting from displacements on individual fault segments (i.e. dislocations) embedded in an elastic half space. Individual fault ruptures are identified from aerial photos and field verification. Average displacements on faults are based on profiles and field measurements of vertical offsets across fault scarps. The values of static stress changes are then used to calculate the Coulomb Failure Function (CFF) (failure stress changes) on the faults associated with the geothermal fields, where CFF is defined as

$$CFF = Dt_s + m(Ds_n),$$

where Dt_s is the static shear stress change (positive in the direction of fault slip), Ds_n is the normal stress change (positive for tensile stress changes), and m is the coefficient of static friction. Possible changes in pore fluid pressure are not considered in the stress calculations. Stress changes moving a fault closer to failure are reflected by positive values of CFF. CFF values are calculated for an assumed dip slip sense of shear (i.e. rake = -90°). In order to evaluate the relative contributions of shear and normal stress changes to CFF values, we determined CFF for $m = 0.0$ (where only Dt_s affects CFF) and $m = 0.40-0.75$ (consistent with laboratory rock mechanics experiments (Byerlee, 1978)). For Dixie Valley calculations, normal stress contributions were determined by subtracting the shear stress changes from the total CFF values.

RESULTS

Active faulting in the vicinity of the Dixie Valley Geothermal Field

"The Dixie Valley Geothermal Field (DVGF) is, in many respects, the classic Basin and Range geothermal reservoir developed along an active major range-front fault" (Benoit, 1995). It is located along the southeast-dipping Dixie Valley fault and within the Stillwater Seismic Gap (Wallace and Whitney, 1984); a 45 km-long section of the fault between the 1915 (M7.7) Pleasant Valley and 1954 (M6.8) Dixie Valley earthquake rupture zones to the north and south, respectively (Figure 2). Within the Stillwater Gap, Wallace and Whitney (1984) recognized discontinuous Holocene fault scarps along the Stillwater range front, and they interpreted this to mean that the gap has experienced at least one large Holocene earthquake. Our field studies of these surface ruptures indicate that ruptures to the north and south of the DVGF are likely of different ages and that the geothermal field lies within a ~ 10 km gap with respect to these Holocene earthquake ruptures. For example, profiles of fault scarps north of the DVGF are generally more degraded with shallower scarp slope angles compared with those to the south, indicating an older age of formation (Figure 3). Fault scarps to the south, with few exceptions, exhibit steep scarp slope angles, essentially at angle of repose ($\sim 30^\circ$) for unconsolidated alluvial fan gravel deposits. Our interpretation of different ages for these scarps is in agreement with previous, more detailed scarp profile analyses of Pearthree (1990) for the Dixie Valley region. There is no evidence for Holocene fault ruptures along the 10 km portion of the range front in the vicinity of the DVGF. It is possible that the lack of scarps in the DVGF is a result of non-preservation due to erosion and/or burial of scarps. However, several landslides and fan surfaces along the range front interpreted to be pre-Holocene in age are apparently not faulted, indicating that this portion of the range front has not ruptured at the surface in the Holocene.

Figure. 2. Generalized fault map of the Stillwater Gap area showing the distribution of historic 1915 and 1954 ruptures (bold red lines), Holocene ruptures (bold orange lines) and Quaternary faults (thin black lines). Base map (modified after Stewart and Carlson, 1978) shows distributions of bedrock (gray) and Quaternary alluvium (yellow). Map also shows the area of the Dixie Valley geothermal field (DVGF).

Figure. 3. Representative fault scarp profiles across Holocene fault ruptures north and south of the Dixie Valley geothermal field. Profile locations are shown on Figure 2. Scarp slope angles are shown below profiles.

Age and extent of Holocene faulting south of the geothermal field

Holocene fault ruptures south of the DVGF overlap with the 1954 Dixie Valley rupture zone by 22 km and can be traced northward for a distance of 45 km, to ~5 km south of the DVGF. The interpreted older Holocene scarps can be traced from ~4 km north of the DVGF to the Sou Hills (Figure 2). Vertical offsets across the fault ruptures south of the DVGF average ~2-3 m and reach a maximum of 5-6 m a few kilometers south of Dixie Comstock Mine. Based on rupture length and average slip estimates, the moment magnitude of the late Holocene earthquake is estimated to be about $M_w 7.3$.

The age of fault ruptures south of the DVGF was previously constrained from field relations Bell and Katzer (1990). Faulted alluvial fan deposits in the Stillwater Gap area contain or overly Mazama tephra (~7 ka) at two locations (Caskey et. al, 1996; work in preparation). Unfaulted fan deposits along the southern portion of the rupture zone contain Turupah Flat tephra (1.5 ka) (Bell and Katzer, 1990). Hecker (1985) used morphologic scarp dating techniques and 27 scarp profiles along the Holocene rupture trace between The Bend and Dixie Comstock and determined a range of mean scarp ages of 2.6-2.9 ka with a more conservative age range of 1.1-7.5 ka. Similarly, Pearthree (1990) using 24 scarp profiles determined a range of mean scarp ages in this area at 2.4-3.0 ka. Although these more tightly bracketed, late Holocene age estimates are consistent with the youthful expression of the scarps, the age of faulting has previously not been as tightly constrained by direct stratigraphic and structural relations.

More definitive constraints on the age of faulting south of the geothermal field come from structural and stratigraphic relations exposed in an exploratory trench excavated across a lateral spread graben on the piedmont near Dixie Comstock mine (this study) and from new radiometric ages on faulted sinter deposits along the range front fault several kilometers to the north Lutz et al., 2002, Lutz and Caskey, unpublished data). The large lateral spread graben near Dixie Comstock (Figure 4) is interpreted as a lateral spread formed by seismically-induced liquefaction of subsurface, water-saturated sediments – most likely pluvial lacustrine deposits. The liquefaction features in this area are just part of the nearly continuous, 30 km-long zone of lateral spreading that extends from The Bend area to about 10 km north of this site (discussed further below). These seismically induced features almost certainly formed during the $M_w \sim 7.3$ earthquake that ruptured along the range front fault. For this reason, an exploratory trench was excavated across the lateral spread graben near Dixie Comstock. The hope was to find a well-stratified and well-preserved record of pre- and post-seismic deposition that would enable us to determine the age of faulting. In this sense, the lateral-spread graben is used as a proxy indicator for the age of faulting at the range front.

Trench results

The trench was excavated across part of the 45 m-wide graben, and exposed a 9 m-thick section of Qfm alluvium truncated by the lateral-spread head scarp on the west side of the graben (Figure 5). Depositionally overlying the head scarp is a large colluvial wedge that extends and thins eastward across the width of the exposure. The wedge shows distinct eastward-inclined layering where it laps onto the head scarp. A 3 m-thick section of graben fill overlies the colluvial wedge. The graben fill consists primarily of well-stratified silty material with thin intertongues of gravel colluvium that thicken toward the scarp. The well-stratified silty deposits likely represent eolian material that has been slightly reworked or modified during wet periods. The silty deposits are

Figure 4. Low-sun-angle photograph of the Stillwater gap trench site area. The trench was excavated across a large lateral-spread graben. Note location of cross section A-A' (referred to in text and in Figure 9) across the compressed zone on the playa margin.

Figure 5. Trench log across part of the lateral-spread graben south of Dixie Comstock (Figure 4). The age of the earthquake that produced the lateral spread is bracketed between 3.7 ka, a bulk radiocarbon age of a paleo-root zone in deposits inferred to be Qfm fan gravel beneath the large colluvial wedge (sample SGRC101), and 2.0 ka, the radiocarbon age of organic material sampled from the easternmost exposure of the colluvial wedge (sample SGRC1). The structure shown beneath the trench log is conceptual, but based on similar structure exposed in a trench across a large lateral-spread graben located about 10 km to the north.

overlain by ~1 m of alluvial gravel that is probably derived from the deep incisions into Qfm alluvium adjacent to the trench.

The fine grained graben fill consists of numerous burn layers with locally abundant charcoal and a 2-3 cm-thick Mono Craters tephra layer that extends across the entire section of graben fill and laps onto scarp-derived colluvium to the west. Radiocarbon ages from four of the burn layers range from 1.8-1.0 ka, with all ages in correct stratigraphic sequence. The 1.0 ka date is from charcoal sampled just below the tephra layer indicating an age for the tephra that is similar to other Mono Craters tephra found in this region. Organic-rich sediment sampled from the distal, westernmost exposure of the colluvial wedge yielded an age of 2.0 ka. A hand excavation at the bottom of the trench exposed a 0.5 m section of moderately oxidized gravel with distinct, partially decomposed roots, some of which are up to 1 cm thick. These partially decomposed roots are absent in the overlying colluvial gravel. We interpret the oxidized gravel in the lowermost part of the trench as Qfm alluvium that is buried beneath scarp derived colluvium. A bulk sample from the buried Qfm alluvium, which contained some of the partially decomposed roots, yielded a radiocarbon age of 3.7 ka. Based on these interpreted stratigraphic relations and radiocarbon ages, the earthquake that produced the lateral spreading occurred after 3.7 ka (age of the buried Qfm alluvium) and before 2.0 ka (age of scarp-derived colluvium).

A more recent study by Lutz et al. (2002) along the range front fault, 15 km to the north, provides additional constraints on the age of faulting in this area. A radiocarbon age from faulted alluvium, which is interbedded with opaline sinter, yielded a radiocarbon age of 3.4 ka, thus providing a slightly lower maximum age of faulting than determined at the trench site. More recently, a sample of faulted pollen-rich opaline sinter yielded a radiocarbon age of ~2.5 ka (Lutz and Caskey, unpublished data).

In summary, if we combine the constraints from the trench and the faulted opaline sinter found along the range front fault to the north, then the age of the most recent fault movement along the fault south of the DVGF is constrained to have occurred between 2.0 and 2.5 ka. This age range is consistent with the morphological scarp ages (2.4-3.0 ka) determined by Hecker (1985) and Pearthree (1990).

Age of faulting north of the geothermal field

Presently, we have no absolute age control on Holocene scarps north of the DVGF. However, Pearthree's (1990) profile analyses indicate that these fault ruptures may be several thousand years older than those south of the DVGF. The observations indicating that scarps north and south of DVGF appear to represent different aged events gives us confidence that the ~10 km gap we observe between Holocene ruptures may be real rather than a consequence of non-preservation.

Lateral-spreads and lateral spread-induced contraction along the piedmont and playa margin south of the DVGF

Debate has recently arisen as to the nature of the extensive, structurally disrupted zone along the piedmont zone between the Bend and the DVGF. This debate came to light at a workshop on Dixie Valley Geothermal Research the Dixie Valley Geothermal field sponsored by the Department of Energy at the Desert Research Institute, Reno, Nevada, June 12. Whether these features are faults (of primary tectonic origin) as argued by David Blackwell (personal communication) or lateral spreads (formed as secondary, surficial features during seismic

shaking) as argued herein, is important for understanding the hydraulic models of the Dixie Valley fault system.

An extensive, structurally disrupted zone exists along the fan piedmont-basin floor boundary between the Bend and the DVGF for a distance of about 30 km (Figure 6). Although large scarps along this zone have been interpreted as fault scarps formed by primary tectonic fault slip (Blackwell, personal communication), we alternatively interpret these features as lateral-spread headscarps formed by seismically-induced liquefaction and shear failure of shallow, unconsolidated, water saturated sediments which allowed down-slope (basinward) movements of overlying relatively cohesive alluvium. Such movements are akin to landslide features and have been extensively described and well documented elsewhere in areas affected by seismically induced liquefaction (e.g., Obermeir, 1987). The lateral spread interpretation of the features in Dixie Valley is based on: 1) the scalloped-shaped nature of the scarps locally (Figures 7 and 8); 2) the distribution of fragmented blocks or mounds that emanate from the head scarp in raft-like fashion (Figures 7 and 8); 4) the close proximity of the disrupted zone to the playa margin where the water table is shallow; 5) soft-sediment (liquefaction) features (e.g., convoluted bedding and sand dikes) expressed locally on cleaned exposures of the disrupted sediments; 6) the graben form of the structures and lack of net vertical offset across the graben (e.g., Figure 9); and 7) the presence of contracted sediments along the playa margin found directly downslope from the piedmont graben (Figures 7 and 8). The presence of the contracted or compressed zones along the playa margin apparently accommodated the down-slope, movement of relatively cohesive material from the lateral spread zones as the material bulldozed its way into the soft, unconsolidated playa margin sediments.

The “compressed zone” along the playa margin is particularly strongly developed downslope from the piedmont at Dixie Hot Springs (Figures 7 and 8) where lateral spreading appears to be the most extensive. Near Dixie Comstock (Figure 4), the compressed zone is not as dramatically expressed as it is at other locations. However, at this location the playa margin is easily accessible, and excellent exposures reveal structural and stratigraphic relations that allowed us to constrain the amount of shortening across the compressed zone. This constraint could then be compared to the structural relations and possible amount of spreading across the piedmont graben found directly upslope.

At the edge of the playa, lacustrine strata are severely buckled up along folds and imbricate “thrusts” that parallel the edge of the playa across a width of ~100 m. Cleaned exposures across the 100 m transect (Cross section A-A’, Figure 9b) revealed 14 different exposures of deformed Mazama ash. The ash bed, at most locations, dips to the east indicating that the strata is repeated along multiple west-vergent “thrusts” and folds. The ash bed is overlain almost everywhere by a thick, well-indurated salt layer that has apparently given rigidity to the lake sediments and allowed for the development of the somewhat regular “fold and thrust belt” geometry of the compressed zone. Bedding attitudes at each of the 14 tephra exposures, a total station profile accurately marking each exposure, and a hand dug pit that revealed the depth to the Mazama ash east of the deformed zone provided constraints for a detailed, reconstructable cross section along the transect (Figure 9b). The cross section was constructed in such a way as to minimize the amount of shortening represented by the folds and thrusts. Although the cross section does not represent a unique solution for the observations, it appears to be reasonable.

Figure. 6. Map of the Stillwater seismic gap region showing extensive zone of paleo-liquefaction formed during the 2.0-3.4 ka earthquake, and locations of Stops 2-4 and 3-2. 1915 Pleasant Valley and 1954 Dixie Valley earthquake ruptures are shown as thick black lines. Holocene ruptures along the Stillwater Range front are shown as thicker gray lines. The 2.0-3.4 ka fault ruptures overlap with the 1954 ruptures by about 20 km and extend for a total length of 46 km to just north of the playa. It is unknown whether the Holocene ruptures at the north end of the valley formed during the 2.0-3.4 ka event. Other late Pleistocene faults are shown as thin black lines (After Dohrenwend et al., 1996). Hatched areas within the liquefaction zone show areas where 1954 liquefaction also occurred (Caskey et al., 1996).

The cross section indicates a minimum of 27 m of shortening across the compressed zone at this location. It stands to reason that the amount of shortening across the compressed zone should approximately balance with the amount of down-slope spreading across graben directly upslope from the playa. Although lateral spreads formed across the entire area between the playa margin and the large graben at the trench site, the greatest spreading probably occurred across the largest graben. At the trench site, the graben width is ~43 m. The 27 m of (minimum) shortening across the playa margin suggests that a significant portion of the graben width was produced by actual down-slope movement of the “detached” alluvium (Figure 9a). The areal extent of severe ground failure related to liquefaction along the 30-km-long disrupted zone is estimated at 20-30 km².

Stress Models

The possibility that the DVGF lies between the endpoints of recent earthquake ruptures has important implications for the redistribution of stresses and their influence on geothermal activity in the DVGF. Our model of stress changes associated with Holocene fault ruptures in the Stillwater Gap (Figure 4) suggests that the Dixie Valley fault and parallel fractures in the vicinity

Figure. 7. Low-sun-angle aerial photograph of the Dixie Hot Springs area showing lateral spread paleoscarps developed on Qfm fan deposits and folded and thrust imbricated lacustrine deposits (compressed zone) along western margin of the playa. These features formed in response to liquefaction during the 2.0-2.5 ka Mw~7.3 earthquake that ruptured along the range-front fault.

Figure. 8. Low-sun-angle aerial photograph of the area immediately south of the Dixie Hot Springs area showing lateral spread paleoscarps developed on Qfm fan deposits and folded and thrust imbricated lacustrine deposits (compressed zone) along western margin of the playa. These features formed in response to liquefaction during the 2.0-2.5 ka $M_w \sim 7.3$ earthquake that ruptured along the range-front fault.

Figure. 9. a) Conceptual cross section across the lateral-spread graben south of Dixie Comstock. b) Cross section A-A' across the playa-margin compressed zone located down-slope from the lateral spread graben in a) (see Figure 1 for location). Cross section is constrained from attitudes of 14 repeated beds of 6.9 ka Mazama ash exposed across the zone of folded and thrust-imbricated lacustrine sediments and a test pit excavated into the playa floor east of the disrupted zone (note east is to the left). The deformed beds are interpreted to be the result of accommodation of the down-slope movement of relatively cohesive Qfm fan deposits above a liquefied zone of water-saturated lacustrine sand. Cross section A-A' suggests approximately 27 meters of shortening across the zone. (Note that shortening was calculated from a non-vertically exaggerated version of the cross section). The graben width is about 43 meters, suggesting that much of that width represents actual down-slope movement of the detached alluvium east of the graben. Note that both figures are shown at the same horizontal scale. The two cross sections are separated by a down-slope distance of about 300 m.

of the DVGF have experienced large positive stress changes (>10 bars). The plot of CFF calculations on a horizontal plane at 5 km depth (Figure 10) shows that the large stress changes in the DVGF are due to its proximity to the endpoints of Holocene fault ruptures to the north and south. Stress changes on an inclined grid, oriented 040° 50° SE and projected to down-dip lengths of 6 km (Figure 11), show stress calculations resolved on the Dixie Valley fault plane as well as proximal, fault-parallel fractures. The three panels in Figure 11 represent the total CFF values (Figure 11a), CFF contributions from shear stress changes (Figure 11b), and CFF contributions from normal stress changes only (Figure 11c). Also shown on Figure 11 are locations and depths to the geothermal reservoir for the northern- and southernmost producing wells within the DVGF (Wells 45-33 and 76A-7, respectively). The models show that CFF values within the DVGF largely result from a decrease in normal stresses (i.e. increase in tensile stresses) (compare Figure 11b and 11c), although a significant increase in shear stress is also evident. The DVGF production area lies above the deepest part of the lobe of decreased normal stress. The model may have important implications for recent borehole studies of in situ stress and fracture conductivity (e.g. Barton et al., 1996; Hickman et al., 1997), as well as the dynamics of fluid flow within the DVGF (discussed below).

An examination of CFF values in the Stillwater Gap associated with slip on historic ruptures along the west-dipping 1915 (M7.7) Pleasant Valley and the east-dipping 1954 (M6.8) Dixie Valley rupture zones (Figure 2) yields similar results for the DVGF, although the stress change values from historic slip are considerably lower in the geothermal field (Figure 12). This is due to the greater distance of the geothermal field from the endpoints of the historic fault ruptures. The plot of CFF calculations, on a horizontal plane at 8 km depth (Figure 12), shows that failure stress along the Dixie Valley fault has everywhere been enhanced in the Stillwater Gap, with most of the fault showing an increase of stress of 1-3 bars. As expected, much larger stress changes occur at the north and south ends of the Gap (>5 bars) near the rupture endpoints. In the fault plane views of the Stillwater Gap segment of the fault (Figure 13) it is clear that stress changes from the 1915 rupture zone are more pronounced in the shallow crust. This is because the Dixie Valley fault and the Pleasant Valley ruptures dip oppositely, and down dip portions of the Dixie Valley fault become increasingly farther from the 1915 ruptures. The fault plane views also show that the historic stress changes in the DVGF result from nearly equal contributions of increased shear stress and decreased normal stress. Although smaller in value to those stress changes associated with Holocene ruptures, the sense of stress change in the DVGF from historic earthquakes is very similar, and therefore would contribute likewise to the state of stress and hydraulic conductivity along the fault.

Mapping along fault scarps in the piedmont area south of the Sou Hills (Figure 2) revealed a small, localized 0.10-0.15 m-high scarp in recent alluvial deposits. The well-preserved nature of such a subtle feature indicates that the scarp is probably historic, and therefore, likely formed during the nearby 1915 earthquake. It may not be a coincidence that the scarp lies within the pronounced lobe of CFF values that result from slip along the 1915 ruptures (Figure 13). The likelihood that the scarp represents a previously unrecognized 1915 rupture appears to demonstrate that stress changes associated with the 1915 earthquake were large enough to trigger fault slip in northern Dixie Valley.

Figure. 10. Coulomb failure stress changes in the DVGF associated with Holocene fault ruptures along the Dixie Valley fault. Calculations are shown on a horizontal grid at 5 km depth, for friction coefficient for 0.75 and resolved on 040° 50° SE oriented normal fault planes (same as Dixie Valley fault in the DVGF) along a -90° rake (dip slip direction of slip). White dislocations represent active slip surfaces. For the fault segments south of the DVGF, we assign values of dip slip of 1.56 m, 2.12 m, and 3.92 m (south to north) based on field measurements of vertical offset across fault scarps. Similarly we assign a dip slip value of 1.96 m for the dislocation north of the DVGF. The transparent dislocation represents the section of the Dixie Valley fault that lies between the Holocene rupture segments.

Figure. 11. Coulomb failure stress changes from Holocene fault ruptures resolved on the Dixie Valley fault in the area of the DVGF for down-dip distances of 6 km. Views are perpendicular to the fault plane as viewed from the southeast. Figures show contributions of CFF from; A) changes in shear and normal stress, B) changes in shear stress only, and C) changes in normal stress only. Stress changes are resolved along a -90° rake (dip slip direction of slip). The map trace of this segment of the Dixie Valley fault is shown above the plots for reference. 76A-7 and 45-33 denote locations of the northern- and southern-most production wells in the DVGF

Figure. 12. Coulomb failure stress changes in the Stillwater Gap associated with slip on the 1915 Pleasant Valley and 1954 Dixie Valley earthquake ruptures (shown as white dislocations). Slip on the dislocations is constrained from detailed field studies of fault slip (Caskey et al., 1996; Wallace, 1984). Transparent dislocation represents the Stillwater Gap portion of the Dixie Valley fault. Calculations are shown on a horizontal grid at 8 km depth, for a friction coefficient of 0.75 and resolved on $038^{\circ} 50\text{SE}$ oriented normal fault planes (same as Dixie Valley fault in the Stillwater Gap) along a -90° rake (dip slip direction of slip).

Figure. 13. Coulomb failure changes from historic fault ruptures resolved on the Dixie Valley fault in the Stillwater Gap for down-dip distances of ~ 19 km. Views are perpendicular to the fault plane as viewed from the southeast. Figures show contributions of CFF from; A) changes in shear and normal stress, B) changes in shear stress only, and C) changes in normal stress only. Stress changes are resolved along a -90° rake (dip slip direction of slip). Map trace of the Stillwater Gap segment of the fault is shown above the fault plane view for reference.

Beowawe Geothermal Field

The Beowawe Geothermal Field (BGF) lies in a more complex structural setting than the DVGF and, at present, has not been field investigated with respect to fault activity to the extent of the Stillwater Gap. However, aerial photo analysis and field reconnaissance mapping during Summer 1999 have resulted in a first order assessment of the distribution and relative ages of the most recently active faults in the vicinity of the BGF.

The Beowawe facility is located within Whirlwind Valley, a small reentrant valley that sits within the Shoshone Range (Figure 14). The west flanks of the Shoshone Range, Dry Hills, and the Cortez Range are each bounded by active faults. The east boundary of Whirlwind Valley is marked by a sharp escarpment which is clearly produced and controlled by the Malpais fault zone. However, it is only along the northern portion of the escarpment that we have observed fault scarps in Quaternary alluvium. These appear to be late Pleistocene to early Holocene-aged scarps based on the estimated age of the youngest alluvial fan deposits offset along the fault. The southernmost limit of observed alluvial scarps along the Malpais fault is approximately 3 km northeast of the town of Beowawe, where we measured vertical offsets of 0.7 m across the scarp. Curiously, the apparent southern limit of surface faulting corresponds to the location of local hot springs along the fault.

Figure. 14. Location map of the Beowawe geothermal field (BGF) showing the traces of the most recent fault ruptures along the west flanks of the Shoshone Range and the Dry Hills, and the northern Malpais fault. These fault ruptures are used for the stress analysis of the BGF (Figure 9).

Structural relations along the southern portions of the Malpais fault in the BGF were most recently investigated by Benoit (1995), who similarly reported an absence of recent scarps. The age of surface faulting along the southern section of the fault appears to predate Quaternary landslide deposits and adjacent bedrock slump blocks approximately 5 km west of the geysers. These landslide deposits are apparently not faulted; an observation supported by mapping in the area by Smith et al. (1979). Future efforts will focus on placing firmer constraints on the ages of the oldest, unfaulted Quaternary deposits along the southern portion of the Malpais fault, and the age of surface faulting along the northern portion of the fault.

Structural and stratigraphic relations along the west flank of the Shoshone Range (Figure 14) indicate multiple Quaternary faulting events. We interpret the most recent event as being early to mid-Holocene in age based on the estimated age of the youngest, faulted alluvial fan deposits at the mouth of Mt. Lewis Canyon. In this area, older fan deposits exhibit progressively greater vertical offset, which attests to repeated Quaternary fault movements. In the same fashion, the Dry Hills fault (Figure 14) exhibits multiple Quaternary displacements, the youngest of which clearly shows Holocene surface expression.

In summary, aerial photo analysis and preliminary field studies show that fault ruptures have been most recent along the Shoshone, Dry Hills, and northern section of the Malpais fault (Figure 14), with no evidence for late Quaternary fault movements in the BGF and elsewhere along the southern Malpais fault. Our model of stress changes associated with these most recent fault ruptures (Figure 15) suggests that the Malpais fault in the vicinity of the BGF has experienced a significant increase in failure stress on the order of 1-2 bars at 3 km depth (Figure 15a). The positive CFF values result from slip on both the Shoshone fault and the northern section of the Malpais fault. Slip on the Dry Hills fault results in negative CFF values along the southern Malpais fault. CFF values calculated for depths of 6 km yield similar results, although these calculations are not included herein.

A comparison of Figures 15a and 15b shows that CFF calculations in the BGF for a friction value of zero (Figure 15b) are less than those for a value of friction of 0.4 (Figure 15a). This indicates that the positive CFF values are a result of both increases in shear stress and decreases in normal stress. The sense of stress changes is similar to the changes in the DVGF. However, the stress changes in the BGF are considerably smaller owing to the greater distance between the geothermal field and the most recent fault ruptures.

Bradys Geothermal Field

Mapping by Bell (1984) indicates that the Bradys Geothermal field lies at the endpoint of Holocene fault ruptures (Figure 16). These ruptures lie within the pluvial Lake Lahontan basin, and are therefore required to be younger than the most recent latest Pleistocene occupation of the lake. We are currently conducting detailed field studies to more completely assess the distribution and age of surface faulting, and have yet to analyze stress changes associated with the fault ruptures. Our preliminary studies indicate that the southern extent of ruptures is accurately portrayed by Bell (1984), but that the fault ruptures may extend farther to the north. These relations would indicate that the Bradys area lies in an area of pronounced stress change owing to its close proximity to a recent fault rupture endpoint. From our previous models of the DVGF we anticipate that these stress changes would result from both increases in shear stress and decreases in normal stress resolved parallel to normal slip on the fault.

Figure. 15. Coulomb failure stress changes at 3 kms depth in the Beowawe geothermal field (BGF) associated with the most recent fault ruptures (white dislocations) in the area. We assign 1.0 m of dip slip on each of the rupture segments along the Shoshone Range, Dry Hills, and northern Malpais fault based on field measurements of vertical offset across the faults. Black dislocation represents the southern Malpais fault. Calculations are for friction coefficient for; A) 0.40, and B) 0.0. Stress is resolved on planes oriented 245° 60° NW (same as southern Malpais fault) along a -90° rake (dip slip direction of slip).

Figure. 16. Location map of the Bradys geothermal field showing previously mapped Holocene fault trace (base map modified after Bell, 1984).

DISCUSSION AND CONCLUSIONS

Our models for the Dixie Valley geothermal field indicate that induced stress changes near the endpoints of recent fault ruptures appear to promote favorable conditions for geothermal production. The stress changes result in an increase in failure stress on unruptured, along-strike portions of the fault. These stress changes are also characterized by significant decreases in fault-normal stress that are particularly enhanced in the upper few kilometers of the crust. Our models of Holocene stress changes in Dixie Valley, suggest the area between the northern- and southernmost geothermal production wells coincides with the portion of the fault most strongly affected by tensile stress changes. Stress changes in the geothermal field induced by the 1915 and 1954 earthquakes are in the same sense as in the Holocene stress model. However, these changes are smaller because of the greater distance between the geothermal field and the historic rupture endpoints. Both the Holocene and historic stress changes we calculate are consistent with results from recent investigations of in-situ stress and fracture permeability in the Dixie Valley geothermal field (e.g. Barton et. al, 1996; Hickman et al., 1997). These studies show fractures parallel to the fault zone are both critically stressed for failure and hydraulically conductive.

Our preliminary assessment of the neotectonic framework of the Bradys geothermal field indicates that this production area lies at the endpoint of a Holocene earthquake rupture, a setting that is analogous to the Dixie Valley geothermal field. We anticipate that stress models for the Bradys area will yield similar results as for the Dixie Valley geothermal field. The Beowawe geothermal field lies in a more complex structural setting. However, our model for late Pleistocene-Holocene stress changes along the southern Malpais fault shows both increases in shear stress and decreases in normal stress in the Beowawe geothermal field.

In summary, we find that the Basin and Range geothermal production fields we have investigated do not appear to lie along the most recently active portions of faults. Our studies show that induced stress concentrations at the endpoints of fault ruptures promote conditions for fracture permeability and geothermal production. . This may be accomplished in two ways: 1) Nearby fault ruptures may induce aftershocks associated with small amounts of slip on macro fractures along the fault and within the damage zone of the fault, thereby producing open fractures while maintaining high failure stress conditions on the fault; and 2) If the geothermal field is close enough to fault rupture endpoints, then it is conceivable that decreases in fault-normal stress may be large enough to produce significant increases in fracture dilatancy, thereby increasing hydraulic conductivity. The observations would suggest that in regions of high heat flow, areas along normal faults which act as persistent rupture segment boundaries may also be areas of enhanced hydrothermal activity. Studies of normal fault segmentation show that fault trace salients (i.e. convex fault strike changes such as at the Dixie Valley geothermal field) and fault trace steps appear to act as the most persistent fault rupture boundaries (Crone and Haller, 1991; Machette et al., 1991).

Our studies illustrate that a detailed understanding of the neotectonic framework and the mechanics of faulting processes are fundamental to developing conceptual models for controls on the state of stress and fracture permeability in geothermal fields. We feel that this integrated approach has enormous potential as a geothermal exploration tool.

Our continuing efforts will focus on constraining better the absolute ages and distribution of fault ruptures in the Dixie Valley, Beowawe, and Bradys areas and on investigating the recent fault history and stress changes for geothermal production fields elsewhere in the Basin and Range province.

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CONFERENCE PAPERS, ABSTRACTS, OTHER PUBLICATIONS, AND PRESENTATIONS UNDER THE AWARD

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