Analysis of the favorability for geothermal fluid flow in 3D: Astor Pass geothermal prospect, Great Basin, northwestern Nevada, USA

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Keywords: structure; fault permeability; Basin and Range; exploration, 3D modelling, geothermal fluid flow; geothermal potential

Abstract

As geothermal exploration increasingly focuses on blind or hidden systems, precise geologic characterization of the sub-surface at potential development sites becomes essential. Geothermal circulation requires elevated heat, relatively high permeability, and ample fluid flow. Evidence for the collocation of these characteristics occur in areas where geothermal circulation is most likely to occur and where exploration activities should be focused. Employing a 3D geologic framework constructed through integration of many separate datasets, we demonstrate a methodology for analyzing the data types that can be used as proxies for these three key characteristics. This methodology is applied at the Astor Pass geothermal prospect in northwestern Nevada, western USA. Based on geologic structure modeled in 3D, several proxies for heat, fluids and permeability are compared in order to identify areas within the field with the highest favorability for geothermal fluid flow. Geological and conceptual models constructed through these methodologies can be used to develop exploration strategies and subsequently site wells. Such models can be iteratively adapted with newly acquired data, as prospects evolve into mature geothermal developments. If developed prior to expensive drilling programs, these techniques allow for more efficient use of limited drilling budgets, ultimately lowering the risks and costs of geothermal exploration and development.

1. Introduction

Robust geothermal activity in the Great Basin, USA, is a product of both anomalously high regional heat flow (85-90 mW/m²) and locally elevated permeability (Blackwell, 1983; Blackwell et al., 1991; Lachenbruch and Sass, 1977; Wisian and Blackwell, 2004). Fracturing associated with some fault systems provides the permeability pathways for geothermal fluid circulation (Blackwell et al., 1999; Faulds et al., 2006). Many geothermal systems are associated with high density faulting and fracturing at structurally complex fault intersection/interaction areas (Curewitz and Karson, 1997; Faulds et al., 2011b, 2010, 2006; Hinz et al., 2011, 2010,

2008; McNamara et al., 2015; Wallis et al., 2012). Conceptual models of such favorable structural settings (e.g. Curewitz and Karson, 1997; Faulds et al., 2011b, 2010, 2006; Hinz et al., 2011, 2010, 2008) are crucial for locating and characterizing geothermal systems in a regional context.

At the local scale, however, characterizing the specific geologic attributes that control the locations of fluid upflow zones within a geothermal system requires precise 3D analysis. A variety of structural, geophysical, downhole, temperature, and geochemical data sets must be integrated in 3D and interpreted *en masse*, in order to elucidate the details of the local controls on geothermal fluid flow. We have developed an innovative 3D analytical technique for such data integration and for analysis of the favorability for geothermal fluid flow at the scale of individual geothermal fields.

Our technique is applied to Astor Pass, a blind, greenfield geothermal prospect on Pyramid Lake Paiute Tribal lands, located ~80 km north of Reno, Nevada, in the western USA (Figure 1). A variety of geologic, geophysical, and shallow temperature studies have investigated geothermal activity at Astor Pass and the surrounding area (e.g. Anderson and Faulds, 2013; Coolbaugh et al., 2006; Cooper et al., 2012; Eisses et al., 2011; Kratt et al., 2010, 2005; Louie et al., 2011; Siler et al., 2012; Vice, 2008; Vice et al., 2007). Subsequent drilling confirmed the presence of a moderate temperature, blind geothermal system (Faulds et al., 2011a; Reeves et al., 2012).

We synthesize the results of previous studies with new analysis of well cuttings and seismic reflection data to produce a 3D geologic model of the Astor Pass geothermal system. Utilizing this 3D geologic framework, we conduct quantitative structural analyses to evaluate the potential for fluid flow along discrete fault zones and within discrete lithologies. We produce a 3D geothermal favorability map of Astor Pass, identifying specific locations in the subsurface and in 3D that are characterized by the collocation of factors necessary to support geothermal fluid upflow (i.e. elevated heat, permeability, and fluids). This work provides detailed information about the systematics of the Astor Pass geothermal system, produces a conceptual model of the geothermal reservoirs and geothermal fluid flow at Astor Pass, and demonstrates a methodology for quantitative 3D exploration techniques that can be applied for locating prospective fluid flow zones in geothermal systems worldwide.

1.1 Geological context: Geothermal activity in the Great Basin

There are ~430 known geothermal systems in the Great Basin region (Faulds et al., 2011b, 2006). As the majority of geothermal systems within the Great Basin show little or no evidence for a magmatic heat component (Kennedy and van Soest, 2007; Kennedy et al., 1997), the primary controls on geothermal system location are crustal heat input and tectonically mediated permeability. Heat flow is elevated region-wide (Blackwell, 1983; Blackwell et al., 1991; Lachenbruch and Sass, 1977), and abundant Tertiary faulting associated with Basin and Range extension generates fracture permeability throughout the Great Basin (Faulds et al., 2006; Wisian and Blackwell, 2004). Spatial correlations between high-temperature geothermal systems and both Holocene faulting (Bell and Ramelli, 2009, 2007) and elevated strain-rate (Faulds et al., 2012) suggest that active tectonism is an important factor in generating and maintaining permeability pathways necessary for robust geothermal upflow. In many geothermal systems in

the Great Basin, such permeability pathways are locally controlled by high density faulting and fracturing occurring at discontinuities along faults, fault intersections, or fault interaction areas with specific geometries (e.g. Faulds et al., 2011b, 2010, 2006; Hinz et al., 2011, 2010, 2008). These settings host areas of closely-spaced faults and fractures, as well as overall accentuated fracture permeability allowing for geothermal fluid flow (Barton et al., 1998a, 1998b; Curewitz and Karson, 1997; Hickman et al., 1997).

1.2 Pyramid Lake area

The Astor Pass geothermal prospect is located directly northwest of Pyramid Lake in western Nevada (Figure 1). This area lies along the boundary between extensional, Basin-and-Range-style tectonics to the east and transtensional, Walker-Lane-style tectonics to the west (Figure 1 inset). The Walker Lane is a belt of dextral shear through western Nevada and eastern California (Faulds and Henry, 2008; Stewart, 1988) and accommodates ~20% of the motion (~1 cm/yr) between the Pacific and North American plates (Kreemer et al., 2009). Along the boundary between these two domains, the Pyramid Lake area is characterized by strain transfer from northwest-trending dextral shear to west-northwest directed extension (Faulds et al., 2011a, 2006, 2005b). One of these dextral faults, the Pyramid Lake fault, located at the southern end of Pyramid Lake, terminates northward into a complex system of north-northwest to north-striking normal and dextral-normal oblique slip faults throughout the Lake Range, Virginia Mountains, and Terraced Hills, including the Astor Pass geothermal area (Drakos, 2007; Faulds et al., 2005a; Vice, 2008; Vice et al., 2007; Figure 1).

A thick section (~1 km) of middle Miocene volcanic rocks known as the lower Pyramid sequence dominates the geologic section in the Astor Pass area. The lower Pyramid sequence consists of primarily basaltic andesite lavas and associated intrusions, along with lesser ash flow tuffs, intermediate to felsic lavas, rhyolite and andesite domes, clastic sedimentary rocks, and diatomaceous shale (Bonham and Papke, 1969; Faulds et al., 2005a, 2002; Henry et al., 2004). At Astor Pass, the lower Pyramid sequence nonconformably overlies Mesozoic granodioritic basement and is locally overlain by Quaternary alluvial fan, calcium carbonate tufa, and lacustrine deposits.

The Astor Pass geothermal prospect lies in a ~2 km-wide, north-northwest-trending, fault-bounded graben known as Astor Pass. The prospect lies ~5 km to the north-northwest (along fault strike) of Needle Rocks, a culturally protected area of modern sublacustrine geothermal outflow, at the northern end of Pyramid Lake (Figure 1). The Needle Rocks area consists of dozens of Pleistocene-to-modern, calcium carbonate tufa towers (up to 100 m high), precipitated as a result of CO₂ degassing during outflow of thermal waters into cold lake water (Benson, 1994). An ~1800-m-deep well drilled in the 1960s at Needle Rocks encountered ~117° C fluids at depth. Geothermal fluids continue to geyser from the uncapped well today (Coolbaugh et al., 2006; Cooper et al., 2012). Though Astor Pass itself has no modern geothermal expression at the surface, it contains a linear arrangement of Pleistocene tufa deposits up to ~10 m-high (Figure 3), similar to those at Needle Rocks (Figures 1 and 2).

Based on detailed analyses of surface geology, gravity data, seismic reflection data, shallow (2 m depth) temperature data (Figure 2, inset), hyperspectral mineral mapping, geothermometry, and

subsequent exploratory drilling, a blind geothermal system has been identified at Astor Pass (Coolbaugh et al., 2006; Faulds et al., 2011a, 2006; Kratt et al., 2010, 2005; Vice, 2008). Drilling targeted an apparent dilational fault intersection zone between two Quaternary faults, a north-northwest-striking normal fault (fault 1) and a northwest-striking dextral-normal fault (fault 2) within the Astor Pass graben (Figure 2). Three wells with depths of ~558 m (APS#1), ~1315 m (APS#2), and ~1379 m (APS#3) were drilled between 2006 and 2011 (Faulds et al., 2011a). Drilling intersected fault 1 at ~600 m downhole. Geothermal fluids with isothermal temperatures as high as 95°C and relatively high flow rates (~450 gpm) were encountered (Reeves et al., 2012). The following 3D analysis of the Astor Pass geothermal system builds upon these initial studies.

2. Materials and Methods

2.1 3D modeling

A 3D geologic model was built for the Astor Pass geothermal system employing Earthvision software (Dynamic Graphics, Inc.) and methods similar to those established in previous efforts (e.g. Hinz et al., 2013; Jolie et al., 2015, 2012; Moeck et al., 2010, 2009a, 2009b; Siler et al., 2012). Fault geometries were modeled based on 1:24,000 scale mapped surface fault traces (Mayhew, 2013; Vice, 2008; Figure 2), interpretation of sixteen 2D seismic reflection profiles (Mayhew, 2013; Figure 4) reprocessed with advanced techniques specifically designed for fault imaging (Louie et al., 2011), geologic cross-sections (Mayhew, 2013; Vice, 2008), occurrences of slickenlines and/or fault gouge in the well cuttings, and the depths of major zones of lost circulation of drilling fluids in the three wells (Mayhew, 2013; Figure 4). Stratigraphic interpretation was not possible from the 2D seismic profiles, so the locations of subsurface stratigraphic horizons were modeled based on geologic mapping and cross-sections (Mayhew, 2013; Vice, 2008), as well as lithologic analysis of well cuttings from all three APS wells (Mayhew, 2013; Figure 4). All three wells are near-vertical and lie within ~500 m of one another (Figure 4), limiting the amount of subsurface stratigraphic control throughout the Astor Pass area. As a result, the basaltic andesite, basalt and volcanoclastic units of the lower Pyramid sequence were assumed to retain the stratigraphic thicknesses obtained from downhole lithologic analyses into areas where data were absent. Although some lateral variations in discrete stratigraphic units are certain in this volcanic terrain, consistent lateral thickness of the lower Pyramid sequence, as a whole, is reasonable for the Pyramid Lake area (e.g. Bonham and Papke, 1969), as evidenced by published geologic maps in the area (Anderson et al., 2014; Faulds et al., 2002; Henry et al., 2004). Three distinct rhyolitic bodies were interpreted from well cuttings (Figure 4). The modeled extent and morphology of these units were based on the surface extent and morphology of rhyolitic rocks mapped in the Astor Pass area (Figure 2; Vice, 2008).

2.2 Geothermal favorability mapping

We define local areas of high geothermal favorability as those where elevated heat and permeability are collocated with evidence for fluids. For this work, we utilize several proxies for the presence of permeability, heat, and fluids along with direct measurements of temperature and fluid flow as indicators of areas that are favorable for geothermal fluid flow.

2.2.1 Fault zone permeability

Fault zones are commonly characterized as consisting of a core of low-permeability clay gouge, surrounded by a higher-permeability breccia zone that grades into fractured wall rock and into undamaged host rock. All of these domains can have highly variable fracture permeability (Caine et al., 1996; Rawling et al., 2001). Bulk permeability in the breccia zone may be as many as ten orders of magnitude higher than intact rock (Brace, 1980; Norton and Knapp, 1977), and structural meshes produced by interlinked extension, shear, and hybrid fractures in the wall rock accentuates primary lithologic permeability, adding significantly to the permeability of the fault zone as a whole (Anders and Wiltschko, 1994; Sibson, 1996). This fault zone related fracture permeability or secondary permeability (as opposed to primary lithologic permeability) is a critical component of the overall permeability that controls fluid upflow in geothermal systems, especially the fault controlled systems in the Basin and Range province. The presence and character of the fault core, damage zone, and host/wall rock permeability domains, however, can vary significantly between fault zones and in different segments of a single fault zone depending on host rock lithology, offset, total displacement, slip rate, stress conditions, and other factors (Caine et al., 1996; Curewitz and Karson, 1997; Evans et al., 1997).

At Astor Pass discrete fault zones are not exposed in a manner facilitating evaluation of this permeability variation between faults, along faults, or in fault interaction zones. Thus, we begin with the assumption that all fault zones at Astor Pass are potential fluid flow conduits. The discrete 3D modeled locations of fault planes were buffered to a 15 m half-width, representing the total width of fault-zone-related fracture permeability. This corresponds to the distance from the fault core at which micro-fracture density asymptotically approaches zero (Scholz et al., 1993) for a ~2.5 km long fault. Since permeability throughout the geothermal field could not be measured in the field or adequately estimated based on the limited drilling data we utilized three proxies for permeability and fluid flow favorability. Within the modelled fault zones, permeability and fluid flow favorability can be accentuated based on; 1) stress state of the fault, 2) proximity to fault intersections and fault tips, and 3) host rock lithology. Details related to the effect each that these parameters have on permeability and fluid flow favorability are described below.

2.2.2. Critically stressed faults

Fault zone fracture permeability and fluid flow are favored in fault segments that are critically stressed under ambient stress conditions (Barton et al., 1995; Ito and Zoback, 2000; Morris et al., 1996; Sibson, 1996, 1994; Townend and Zoback, 2000; Zoback and Townend, 2001). The tendency of a fault segment to slip (Morris et al., 1996) or to dilate (Ferrill et al., 1999) is one measure of whether faults or fault segments are critically stressed and therefore likely to transmit geothermal fluids. The slip tendency (T_s) of a surface is defined by the ratio of shear stress (τ) to normal stress (σ_n) on that surface:

$$T_s = \tau / \sigma_n \quad (\text{Morris et al., 1996}).$$

Dilation tendency (T_d) is a measure of all the stress components acting normal to a given surface:

$$T_d = (\sigma_1 - \sigma_n) / (\sigma_1 - \sigma_3) \quad (\text{Ferrill et al., 1999}).$$

Slip and dilation tendency values range from one to zero. For values of one, a fault plane is ideally oriented to be critically stressed for slip or dilation under ambient stress conditions. If zero, a fault plane has no resolved stresses and no potential to slip or dilate (Ferrill et al., 1999; Morris et al., 1996).

The stress field utilized in slip and dilation tendency calculations (Figure 5) was calculated using standard methods for analysis of borehole breakouts, the details of which can be found in Zoback (2010) and Zoback et al., (2003). The orientations of the horizontal stress field were determined from stress induced features (borehole breakouts and drilling induced tensile fractures) observed on acoustic and resistivity borehole images from wells APS#2 and APS#3 (Figure 5). The orientations of borehole breakouts indicate an S_{hmin} (minimum horizontal stress) orientation of $093^\circ \pm 12$ (273°) at Astor Pass.

The vertical stress (S_v) magnitude was calculated as a function of depth, with density values taken from density logs in well APS#3. Where no density logs were available, density values were obtained from sonic logs. The relationship between sonic and density values within the wells was determined from observed relationships in a depth interval of well APS#3 that had both sonic and density log data. An overall S_v gradient of ~ 23.5 MPa/km was calculated (Figure 5).

Appropriate data, such as extended leak-off tests (XLOT) or mini-frac tests (Zoback, 2010), were unavailable from the Astor Pass wells. As such, low confidence, lower limits for the horizontal minimum stress in wells APS#2 and APS#3 are determined from formation integrity tests (FIT) at three different depths. These values are considered a lower limit estimate due to the location of the FIT test on a schematic XLOT time-pressure curve (Zoback, 2010). This method provides a S_{hmin} gradient of ~ 13.2 MPa/km (Figure 5).

The magnitude of S_{Hmax} (maximum horizontal stress) was calculated using the determined S_{hmin} (likely an underestimate) and S_v values, borehole breakout widths as observed on acoustic and resistivity borehole images (e.g. Figure 5), pore pressure (equal to hydrostatic pressure), and an assumed range of unconfined compressive strengths (UCS = 100, 150, and 300 MPa) for andesite lavas (e.g. Siratovich et al., 2014; Wyering et al., 2014) which probably have similar strength characteristics to the Astor Pass basaltic andesites. This was done using the equation:

$$S_{Hmax} = \frac{UCS + 2P_p - S_{hmin}(1 + 2 \cos 2\theta_b)}{1 - 2 \cos 2\theta_b}$$

Where P_p is pore pressure and:

$$2\theta_b = \pi - W_{bo}$$

Where W_{bo} = borehole breakout width.

For all values of andesite UCS tested, the S_{Hmax} magnitudes were larger than the S_v magnitudes at all depths (Figure 5). As shown in other geothermal fields with an andesite lava hosted reservoir, care should be taken when assuming the strength of andesite, as it is susceptible to a variety of

effects, such as microfracture density and alteration (Siratovich et al., 2014; Wyering et al., 2014). As a confidence test of our S_{hmin} magnitudes (which are assumed to be an under-estimate), we noted the effect of increasing the S_{hmin} gradient on S_{Hmax} values. This resulted in higher S_{Hmax} values, which allows us to infer with some confidence that $S_{Hmax} > S_v$ in the Astor Pass area. S_{Hmax} calculated with a UCS value of 155 was utilized for the slip and dilation tendency results incorporated into the 3D model.

Slip and dilation tendency values (calculated using 3D stress Morris et al., 1996; Ferrill et al., 1999) were calculated for each fault plane using this stress model (Figure 5). These values provide quantification of the along-fault and fault-to-fault variation in the potential for faults to be critically stressed for either slip or dilation. These values are considered as proxies for fracture permeability and fluid flow favorability along the fault zones.

2.2.3 Fault intersection density

Fracture permeability is also favored within jogs, intersections, ramps and other discontinuities along fault zones and at the tips of faults (Curewitz and Karson, 1997; Pollard and Aydin, 1988; Scholz et al., 1993). Stress concentrations in these areas generate local zones of enhanced fracturing, producing and maintaining local fracture permeability. Any strand in the cluster of high-density fault segments within one of these fault interaction areas has the potential be reactivated, accentuating the potential for fracture permeability (Curewitz and Karson, 1997; Sibson, 1996). Based on the 3D geometry of modelled fault zones, all discrete fault tips and fault intersections within the Astor Pass geothermal field were spatially registered, and the spatial density of fault tips/intersections was calculated. The density of fault intersections and fault tips, calculated as a density of intersections and fault tips per unit volume (using methods similar to Alberti, 2011) are considered as a proxy for elevated fracture permeability.

2.2.4 Reservoir lithology

The lithology of potential geothermal reservoir intervals has a significant effect on the permeability and therefore the economic potential of a geothermal system (e.g. Hinz et al., 2011). Relatively incompetent or incoherent lithologic units (e.g. shales and unconsolidated sediments) do not support dense, interconnected, fault and fracture networks or allow large-magnitude fluid flow to the same degree as more competent units (Hill, 1977; Sibson, 1996). Therefore, competent lithologies that host critically stressed fault zones and a high density of fault intersections/tips make the most attractive reservoir intervals, whereas more incompetent lithologies are unlikely to serve as adequate reservoirs. At Astor Pass data obtained from a pressure-temperature-spinner survey (Reeves et al., 2012) allowed evaluation of the flow in the various 3D modelled lithologies and therefore indicate relative lithologic favorability.

2.2.5 Temperature and fluids

Along with permeability, elevated temperature is another key element required for geothermal circulation. Within the Great Basin region average heat flow is 85-90 mW/m² and as high as 100 mW/m² (Blackwell, 1983; Blackwell et al., 1991; Lachenbruch and Sass, 1977). This regionally high heat flow is primarily associated with Cenozoic extension and resultant crustal thinning

(rather than active magmatism). This is true at Astor Pass, as well, where $^3\text{He}/^4\text{He}$ ratios indicate that heat within the geothermal system is primarily derived from circulation of geothermal fluids in crustal materials rather than young mantle-derived magmas (Cooper et al., 2012).

Locally, convective/advective heat transport associated with geothermal circulation is a more efficient means of transporting heat to the near surface than conductive transport. High temperature anomalies observed in wells or near-surface temperature measurements can therefore serve as a proxy for the presence of circulating fluids. Zones of relatively high conductivity in magnetotelluric surveys are also commonly interpreted as smectite clay alteration, which is generated by active (or perhaps extinct) fluid flow (Newman et al., 2008; Spichak and Manzella, 2009). Isothermal temperature gradients, are also commonly interpreted as caused by fluid circulation and high fracture connectivity. Care must be taken in interpreting temperature data, however, as high temperature anomalies in wells and shallow temperature data can indicate either upflow zones, which are ideal for sustained geothermal production, or outflow zones which can be shallow features spatially removed from their deeper source.

Each of the above proxies for permeability, temperature, and fluids was normalized to a 0-1 favorability scale, with zero the least favorable for geothermal fluid flow and one the most favorable. This was done because we cannot constrain how, for example, fracture permeability associated with critically stressed faults quantitatively relates to fracture permeability generated at fault tips or fault intersections. By normalizing and summing these proxies, we can semi-quantitatively examine and evaluate the locations where the permeability proxies are most highly collocated and therefore indicate high favorability for geothermal fluid flow (e.g. Figure 6).

3. Results

3.1 3D Structural and stratigraphic modeling

The 3D geologic model of the Astor Pass geothermal prospect consists of 19 fault planes and 7 lithologic units (Figure 7). The Astor Pass graben is bounded by steeply west-dipping, anastomosing normal faults on the east side and one steeply east-dipping normal fault on the west side. The stratigraphy consists of Quaternary alluvium and Pleistocene tufa resting in angular unconformity on a ~1000 m-thick section of the middle Miocene lower Pyramid sequence. In the Astor Pass area the lower Pyramid sequence consists primarily of basalt to basaltic andesite lava flows with intercalated ash flow tuffs and local rhyolite bodies. The lower Pyramid sequence nonconformably overlies Mesozoic granodioritic basement rocks (Bonham and Papke, 1969; Faulds et al., 2011a; Mayhew, 2013; Vice, 2008; Vice et al., 2007). The predominant west-dipping fault set accommodated ~20-30° eastward tilt of the Miocene strata, which is evident in the 3D model (Figure 7).

Predominantly propylitic, hydrothermal alteration occurs in well cuttings in both the Miocene and Mesozoic sections at Astor Pass. Mafic lithologies vary from propylitic alteration of mafic minerals, to minor zeolite mineralization, and to unaltered. Ash flow tuff intervals commonly have abundant secondary clays. Between 600-800 m downhole in the two deeper wells, APS#2 and APS#3, propylitic alteration is pervasive. Chalcedony and quartz veins and filled voids along with trace pyrite also occur locally in this section. The top of this interval corresponds with the

occurrence of slickenlines and gouge in well cuttings. Well cuttings in this pervasively altered 600-800 m downhole section also include several intervals of rhyolite and tuff within the predominantly basaltic andesite section. These are interpreted as local rhyolitic intrusions (Figure 7). This zone of intense alteration, gouge, and slickenlines lies at the intersection of the well bores and fault 1. Although some of the intense propylitic alteration may be associated with intrusion of the rhyolites, close association with slickenlines and gouge suggests that the alteration is more likely associated with the modern geothermal system, especially since little alteration has been noted proximal to the rhyolite intrusions in surface exposures. Below ~800 m downhole in APS#2 and #3 propylitic alteration of mafic minerals is common, yet less extensive than within the 600-800 m section. Propylitic alteration and secondary quartz and calcite occur both along with and in the absence of slickenlines and gouge in cuttings below ~800 m. Secondary quartz veins are present, though not pervasive, in the granodioritic basement to total well depth in APS#2 and #3.

A steeply west-dipping, normal fault zone (fault 1) and a steeply northeast-dipping, dextral-normal fault zone (fault 2) intersect at the surface near the Pleistocene tufa deposits (Figures 2, 3 and 7). This fault intersection was originally interpreted as the primary controlling structure of the geothermal system (Faulds et al., 2011a; Vice, 2008; Vice et al., 2007). Our results, incorporating new subsurface geophysical data, well data, and 3D geologic modeling, are consistent with this interpretation.

In addition to faults 1 and 2, a third, north-striking, steeply west-dipping normal fault (fault 3) also intersects fault 1 at the Pleistocene tufa deposits (Figures 2 and 3). Neither fault 2 nor fault 3 are penetrated by the APS wells (Figure 7). 3D modeling shows that this complex three-way fault intersection produces two discrete fault intersection zones in the subsurface, one plunging steeply southwest and one plunging moderately northwest (Figure 7; Siler et al., 2012). As evidenced by their proximity to both the Pleistocene tufa deposits and shallow temperature anomaly (Figures 2 and 3), these two intersection zones are likely important in channeling geothermal fluid upflow to the near surface.

3.2 Fracture permeability potential

Stress field calculations for the Astor Pass geothermal field indicate that the minimum horizontal stress (S_{hmin}) direction is oriented 093° . The stress magnitudes of the maximum horizontal stress (S_{Hmax}) are larger than the vertical stress (S_v ; Figure 5), indicating that Astor Pass lies in a predominantly strike-slip faulting stress regime. The proximity of Astor Pass to the Walker Lane is consistent with this interpretation. However, because Astor Pass straddles the boundary between Walker Lane-style strike-slip and Basin and Range-style extensional tectonics, it may have alternated between periods dominated by strike-slip ($S_{Hmax} > S_v$) or normal faulting ($S_v > S_{Hmax}$) during evolution of the fault system and geothermal system. Whether the S_v magnitude or the S_{Hmax} magnitude is larger has a profound effect on the attitude and orientation of fault segments that are critically stressed for *slip*, which considers shear stress on fault planes, but very little effect on which segments have a high tendency to *dilate*, which is controlled by the stress components acting normal to fault planes. Because the orientations of fault segments that are critically stressed for slip under normal vs. strike-slip faulting stress regimes is highly

variable, we employed dilation tendency, which is much less sensitive to the polarity of the stress field, as a more reliable proxy for the stress control on permeability and fluid flow throughout the development of the Astor Pass geothermal system. The dilation tendency of Astor Pass faults, in the absence of subsurface permeability data throughout the field, therefore serves as a proxy for along fault and fault-to-fault variability in the critically stressed nature of faults and indicates fault segments that are most likely to have accentuated fracture permeability and favorability for geothermal fluid flow.

Under these prevailing stress conditions, north-striking (003°) steeply dipping fault segments at Astor Pass have the highest dilation tendency (Figure 8). As a result of significant variability in both fault strike and fault dip at Astor Pass, there is also substantial along-strike variability in dilation tendency. Fault segments with strikes that are increasingly oblique to 003° and/or with dips that are increasingly more shallow have increasingly lower dilation tendency and therefore lower favorability for fracture permeability and fluid flow (Figure 8). Still, many faults and segments of faults are critically stressed for dilation, including segments of all three faults that we interpret to control the geothermal system (Figure 8).

Fault intersection/fault tip density is another structural characteristic that we employed as a proxy for fault zone permeability in our geothermal fluid flow favorability mapping. Discrete fault tips and intersections between two faults occur as sub-vertical to shallowly plunging tubular bodies of elevated permeability potential, whereas areas with a higher density of fault intersections/tips form broader volumes of elevated permeability potential (Figure 9).

Pressure-temperature-spinner data from APS#3 give an indication of the relative permeability potential of different lithologic units at Astor Pass. These data indicate that flow contribution to the APS#3 wellbore is highest from intervals between 1258-1269 m and 1284-1289 m downhole and that only minor contributions to flow occur elsewhere in the well (Reeves et al., 2012). Both of these high-flow-rate intervals lie in basaltic andesites in the middle Miocene lower Pyramid sequence. APS#3 crossed fault 1 much higher in the section, at ~600 m downhole, did not intersect discrete faults at or near these depths with high flow rates in either the Miocene section or the underlying Mesozoic granodioritic basement rocks. Thus, these high flow intervals are associated with generally high transmissivity and fracture interconnectivity in the Miocene section, rather than with discrete fault zones (Reeves et al., 2012). A 30 day flow test at 450 gpm of APS#3 produced immediate and substantial drawdown in APS#2 and APS#1, which are completed at ~1300 m depth and ~550 m, depth respectively, both in the lower Pyramid sequence. Drawdown was also seen in a shallow (~50 m) water well NGC4 (Figure 2 and 7), which is completed in unconsolidated Quaternary sediments ~1 km north of APS#3 (Reeves et al., 2012). These observations indicate high reservoir connectivity throughout the lower Pyramid sequence and overlying Quaternary sediments. Because very little contribution to wellbore inflow comes from the unfaulted granodioritic basement rocks, the basement must have very low primary permeability in this area, and any fluid upflow through the basement must occur on discrete structurally controlled permeability pathways.

3.3 Temperature and fluids

Anomalously elevated temperatures at 2 m depth are clustered around and to the south and southeast of Pleistocene tufa deposits (Figure 2, inset), suggesting that local upflow of geothermal fluid is focused in this area (Coolbaugh et al., 2006; Kratt et al., 2010). Magnetotelluric data show a zone of high conductivity near the Pleistocene tufa deposits, also indicating local and focused fluid upflow (Reeves et al., 2012). Equilibrated, downhole temperature measurements indicate isothermal $\sim 95^{\circ}\text{C}$ temperatures from ~ 100 m depth to the bottom of all three APS wells (~ 500 and ~ 1300 m). The 30 day flow test of APS#3 produced fluids with $\sim 95^{\circ}\text{C}$ maximum temperatures (Reeves et al., 2012). Aqueous geothermometry from Astor Pass suggests reservoir temperatures of $\sim 115^{\circ}\text{C}$ (quartz geothermometer with no steam loss, Cooper et al., 2012), but reservoir temperatures may be as high as 143°C (Na-K-Ca-Mg geothermometer Coolbaugh et al., 2006). At nearby Needle Rocks, geothermometers reflect reservoir temperatures as low as 116°C (chalcedony) and as high as 213°C (Na-K-Ca-Mg, Coolbaugh et al., 2006). Boiling surface and sublacustrine outflow also occurs at Needle Rocks.

3.4 The geothermal favorability map

The final geothermal favorability map, constructed through summing the normalized dilation tendency and fault intersection density volumes, reveals several locations within the Astor Pass area where critically stressed fault segments and high densities of fault intersections are collocated. These areas are favorable for high permeability (Figure 10). The surface expression of recent geothermal activity at Astor Pass, however, is relatively localized and marked by Pleistocene tufa deposits (Figures 2 and 3), evidence of sublacustrine geothermal outflow during lake high-stand, a conductivity anomaly (Reeves et al., 2012), and a shallow temperature anomaly (Kratt et al., 2010), all of which are collocated (Figure 2). We conclude that of the areas in the subsurface that are prospective for elevated permeability, only those capable of supplying fluids to the near-surface in the vicinity of the Pleistocene tufa deposits are utilized as fluid flow pathways for the modern geothermal system. The two most likely geothermal fluid flow conduits, therefore, are associated with the moderately northwest-plunging and steeply southwest-plunging fault intersections (Figures 7 and 10). Two other steeply plunging fluid flow conduits lie ~ 100 m west of the modern geothermal features. Though still favorable, these secondary conduits are associated with less dense areas of fault intersection and northwest-striking fault segments rather than with more ideally oriented north-striking fault segments (Figure 10).

4. Discussion

The geothermal gradient at Astor Pass is not sufficiently high to generate the 95°C sampled fluids *in situ*. This indicates that there is geothermal upwelling from a 95°C or hotter geothermal reservoir at depth beneath the northern end of Pyramid Lake that charges the highly interconnected stratigraphic reservoir in the Miocene lower Pyramid sequence with geothermal fluids. It is not clear whether the Needle Rocks and Astor Pass thermal anomalies are both part of the same geothermal system or are separate systems (Cooper et al., 2012). If they are part of the same system, the reservoir at depth may be as hot as 213°C , the highest temperature obtained from geothermometry from the Needle Rocks (Coolbaugh et al., 2006). If not, the deeper reservoir beneath Astor Pass may reach temperatures between 115 - 142°C , the temperatures

obtained from geothermometry of Astor Pass geothermal fluids (Coolbaugh et al., 2006; Cooper et al., 2012).

Primary lithologic permeability in the granodioritic basement is low (Reeves et al., 2012), so the fluid upflow within and through the basement, which charges the shallow reservoir, must be structurally controlled. Fluid upwelling through the granodioritic basement beneath Astor Pass is likely focused on north-striking, steeply-dipping fault segments and fractures within the two discrete structural intersection zones beneath the Pleistocene tufa deposits; the moderately northwest-plunging intersection zone and near-vertical southwest-plunging intersection zone (Figures 7, 10 and 11). Additional upflow is possible on steeply plunging, but less ideally oriented fault strands and intersections ~100 m to the east (Figures 10 and 11). Fluids upwelling along these discrete structural conduits in the granodioritic basement charge the shallow stratigraphic reservoir in the Miocene lower Pyramid sequence (Figure 11). These upwelling fluids imprint the relatively high geothermometry on the fluids produced from the lower Pyramid sequence at Astor Pass.

Temperatures in the Miocene stratigraphic reservoir are isothermal, indicating that this stratigraphic package has high connectivity and transmissivity through much of the Miocene stratigraphic section (Reeves et al., 2012). The 30 day flow test of APS#3, which caused drawdown in APS#1, APS#2 and NGC4, also supports high connectivity and transmissivity throughout the Miocene and younger section. This reservoir connectivity allows for temperature equilibration throughout the stratigraphic reservoir (Reeves et al., 2012). The 30 day flow test did not produce fluids hotter than 95°C, suggesting that the shallow equilibrated reservoir extends at least into the top of granodioritic basement, which would have been intersected by the cone of depression sampled during the flow test (Reeves et al., 2012).

Even within the highly interconnected, primarily stratigraphic reservoir in the lower Pyramid sequence, fluid flow to the near surface is concentrated along the modelled fault intersection zones, as evidenced by the localized shallow temperature anomaly, conductivity anomaly and Pleistocene tufa deposits. Fluid outflow into the unconsolidated Quaternary sediments mixes with the general north-northwest to south-southeast, down-valley gradient of cool groundwater toward Pyramid Lake. This has the effect of ‘smearing’ out the shallow temperature anomaly to the south-southeast of the Pleistocene tufa deposits (Figures 2 and 11; Reeves et al., 2012).

Both the deeper, hotter structural reservoir and the overlying shallow, cooler reservoir are viable geothermal prospects. The shallow, stratigraphic reservoir, with moderate temperature and high flow rates (~95°C and ~450 gpm; Reeves et al., 2012) is cooler than typical Great Basin geothermal fields that generate electricity with binary technologies, but these temperatures and flow rates are viable for numerous direct use and small scale electricity applications. The deeper reservoir is a higher risk prospect. The depth of the deep reservoir is unknown, but it may be deeper than ~1300 m, the depth of existing drilling. Reservoir temperatures as high as 213°C are indicated by geothermometry. However, fluids at these temperatures have not been sampled at Astor Pass, and it is unclear if these temperatures, obtained from fluids at Needle Rocks, are associated with the geothermal system at Astor Pass or with a separate system at Needle Rocks. Reservoir temperature estimates based on geothermometry can also have a variety of significant

uncertainties (see Ferguson et al., 2009 for a review of the uncertainties associated with aqueous geothermometry). Future exploration in this deeper, hotter reservoir should be focused on the modelled fault intersections at basement depths, <1300 m. Increasingly higher drilling costs and risks are associated with increasingly deeper drilling. Further, uncertainty and risk in exploring the deeper reservoir stems from the lack of constraints on the geometry of discrete structural targets in the deep reservoir, which is below the depth of existing seismic reflection imaging, lithologic control, and 3D modeling. However, it is probable the major fault intersections continue into at least the upper part of the Mesozoic basement at approximately the same orientations as established in the Miocene sections.

5. Conclusions

Astor Pass is a blind, greenfield geothermal prospect. Similar to many geothermal prospects, surface geologic data are available along with sparse temperature and subsurface data, but extensive exploratory drilling has yet to be conducted. We demonstrate that through careful (and relatively inexpensive) analysis of surface geologic data and incorporation of subsurface data, a 3D conceptual model of fluid flow within a geothermal prospect can be constructed, prior to extensive drilling investments.

Geothermal fluid flow is most likely to occur where heat, fluids, and permeability are collocated in space. The technique that we present describes the datasets that can be used as proxies for these essential elements of a geothermal system and methodologies that can be used to compare and integrate these datasets. Our methodology synthesizes this information in 3D space allowing for identification and location of favorable fluid-flow conduits.

The application of these techniques to Astor Pass has elucidated the details of a shallow, cooler stratigraphically controlled geothermal reservoir. We also interpret the existence of a deeper, hotter geothermal reservoir that charges the stratigraphic reservoir through fluid upflow along discrete structural conduits, which in this case correspond to discrete fault intersections. Testable conceptual models based on 3D modelling, like this for Astor Pass, can facilitate more cost effective decisions regarding subsequent exploration and drilling, thus mitigating risks and costs of geothermal exploration and development.

Acknowledgments

We thank the Pyramid Lake Paiute Tribe for allowing us to conduct research on their lands. We thank Dynamic Graphics Inc., Alameda, CA for providing the Earthvision 3D software. The assistance and advice of Robert McFaul at Dynamic Graphics Inc. was invaluable in development of the 3D modeling and geothermal fluid flow favorability mapping workflow, as was the assistance of Alan Morris with 3D Stress. We thank Cecile Massoit for her help in construction of the stress model. We also thank D.M. Reeves, P.F. Dobson, I. Warren, and an anonymous reviewer for their insight and thoughtful review of this manuscript. This work was supported by American Recovery and Reinvestment Act grants from the U.S. Department of Energy (award EE0002748) to James Faulds and (award EE0002842) to the Pyramid Lake Paiute Tribe. The U.S. Department of the Interior through the Assistant Secretary of Indian

Affairs, Division of Energy & Mineral Development, funded the borehole geophysics of the Astor Pass wells.

Figure Captions

Figure 1. Regional digital elevation model of the Pyramid Lake area. Dotted outline indicates the extent of the Pyramid Lake Paiute Reservation. Box indicates Astor Pass and extent of Figure 2. Inset shows Pyramid Lake area with respect to the Walker Lane and Basin and Range tectonic regimes.

Figure 2. Simplified geologic map of the Astor Pass area (after Vice, 2008). Purple highlighted faults are interpreted as controlling structures of the geothermal system (Siler et al., 2012; Vice, 2008; Vice et al., 2007; and this work). Black double-lined rectangle indicates the areal extent of the Astor Pass 3D model. Inset shows the Astor Pass temperature anomaly as defined by 2 m probes (Kratt et al., 2010).

Figure 3. Photograph, looking east, of the Astor Pass tufa. The tufa mound is ~10 meters high and ~1 km long in the long axis of the photograph (roughly in the north-south direction). The tufa has a 'Y' shape (e.g. Figure 2). The northern limbs of the 'Y' (left side of the photograph) are aligned along fault 2 (in the foreground) and fault 3 (in the background), while the southern end of the 'Y' (right side of the photograph) is aligned along fault 3.

Figure 4. An example of input data used in construction of the Astor Pass 3D model, looking southeast. One of the 16 interpreted 2D seismic reflection profiles is 'hung' in 3D space along with the lithologic interpretation along the paths of the three APS wells. Digitized points, based on several interpreted reflection profiles, for fault 1 are shown (purple) along with the modelled fault 1 surface. Red and green rings along the well path indicate the locations of zones of lost circulation during drilling (red) as well as the location of clay gouge or slickenlines in well cuttings (green), respectively.

Figure 5. Top left, rose diagrams of stress induced fractures in APS#2 (blue) and APS#3 (red). Outer edge of rose diagram represents 20% of the picked structures. Left side, acoustic image and interpreted borehole breakouts (outlined in pink) at 1140-1150 m depth in APS#2. Right side, Astor Pass stress model incorporating data from stress induced fractures in APS#2 and APS#3. The S_{hmin} magnitude, 13.2 Mpa/km, is a lower limit, calculated based on FIT tests with $\mu=1.0$ (lower value) and $\mu=0.6$ (upper value). S_v is 23.5 Mpa/km. S_{Hmax} was calculated based on breakout widths for UCS values of 100, 155 and 300 and the S_{Hmax} calculated with UCS 155 was utilized for the stress analysis.

Figure 6. Conceptual model of the 3D geothermal favorability mapping technique. Steeply dipping fault segments orthogonal to S_{Hmin} (red) are the most critically stressed for dilation and have a higher likelihood of hosting elevated fracture permeability and fluid flow. Fault segments oblique to S_{Hmin} (blue) are unlikely to host high fracture permeability and fluid flow. Where critically stressed fault segments lie in coherent lithologic units, permeability is more favorable (red). Where critically stressed fault segments lie in incoherent lithologies and/or where non-critically stressed fault segments lie in coherent lithologies, permeability is less favorable (purple). Permeability is least favorable where non-critically stressed fault segments lie in incoherent lithologies (blue). Discrete fault intersections (dashed circles) within the most favorable zones (red) and within high temperature areas (shading) are highly favorable for geothermal fluid flow.

Figure 7. The 3D geologic model of Astor Pass looking north-northeast. Lithologies based on analysis of drill cuttings (Mayhew, 2013) are shown along the well paths. Red highlighted fault traces are those interpreted by Vice et al. (2007), Vice (2008), and Siler et al. (2012), and this work as controlling structures of the geothermal system. The moderately northwest-plunging and steeply southwest-plunging fault intersections (purple) generated by these intersecting faults are interpreted as the primary fluid flow conduits. Inset shows the Astor Pass 3D geologic model sliced at APS#3. The full modeled volume is 4 km x 4 km x 1.5 km.

Figure 8. Dilation tendency of the three faults that control fluid flow at the Astor Pass geothermal system looking northwest. Warm colors correspond to fault segments most critically stressed for dilation (steeply dipping fault segments ~orthogonal to the 093° minimum horizontal stress and based on borehole derived stress data, (Figure 5)); cool colors correspond to fault segments poorly oriented for dilation. For vertical scale, the APS#2 and APS#3 well paths are ~1300 m long.

Figure 9. Fault intersection and fault tip density at Astor Pass looking northeast. Warm colors correspond to higher spatial densities of fault intersections and fault tips, and cool colors show lower spatial densities.

Figure 10. Geothermal favorability map of Astor Pass. Warm colors correspond to the most favorable areas for geothermal fluid flow, cool colors to the least favorable. The model is sliced at ~1200 m, near the top of the Mesozoic basement.

Figure 11. Conceptual model of the Astor Pass geothermal system looking northeast. Structurally controlled fluid upflow (orange) from a deeper, hotter geothermal reservoir through the basement (red) flows out into the highly transmissive and interconnected middle Miocene stratigraphic reservoir (blue). Hot upflow from the deep reservoir emanates from an unknown depth that is below the model. Outflow into unconsolidated Quaternary sediment (yellow) mixes with the southeast-flowing (left-to-right) shallow and cool ground water gradient, resulting in a ‘smearing’ out of the shallow temperature anomaly to the south-southeast of the controlling fault intersections and Pleistocene tufa deposits (brown).

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