**The plumbing system of the Pagosa thermal Springs, Colorado:
Application of geologically-constrained geophysical inversion and
data fusion**

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36 **Abstract.** Fault and fracture networks usually provide the plumbing for movement of
37 hydrothermal fluids in geothermal fields. The Big Springs of Pagosa Springs in Colorado is
38 known as the deepest geothermal hot springs in the world. However, little is known about the
39 plumbing system of this hot spring, especially regarding the position of the reservoir (if any)
40 or the position of the major tectonic faults controlling the flow of the thermal water in this
41 area. The Mancos shale, a Cretaceous shale, dominates many of the surface expressions
42 around the springs and impede an easy recognition of the fault network. We use three
43 geophysical methods (DC resistivity, self-potential, and seismic) to image the faults in this
44 area, most of which are not recognized in the geologic fault map of the region. Results from
45 these surveys indicate that the hot Springs (the Big Spring and a warm spring located 1.8 km
46 further south) are located at the intersection of the Victoire Fault, a major normal crustal fault,
47 and two north-northeast trending faults (Fault A and B). Self-potential and DC resistivity
48 tomographies can be combined and a set of joint attributes defined to determine the
49 localization of the flow of hot water associated with the Eight Miles Mesa Fault, a second
50 major tectonic feature responsible for the occurrence of warm springs further West and South
51 from the Big Springs of Pagosa Springs.

52

53 **Keywords:** Electrical resistivity tomography, inversion, geothermal field, hot springs.

54

55 **1. Introduction**

56 There is currently a renewed interest in using geophysical methods to track the
 57 plumbing system of hydrothermal fields (Aizawa et al., 2009; Mariita, 2009; Byrdina et al.,
 58 2013; Soengkono et al., 2013; Komori et al., 2013a; Muñoz, 2014). However, there is not a
 59 single geophysical method that can perform this task alone (Mariita, 2009). The seismic
 60 method can be used to provide structural information about the subsurface but it is generally
 61 not working properly in heavily fractured media (because of scattering of the seismic waves)
 62 and is not a direct indicator of geothermal features (Kent et al., 2013). Resistivity alone,
 63 obtained either using galvanometric and induction-based techniques, is not necessarily an
 64 indicator of geothermal reservoirs or flow pathways. Indeed, fractured reservoirs with hot
 65 fluids can have the same resistivity signature than clay-rich rocks, especially in presence of
 66 smectite and illite (Komori et al., 2010 and 2013b; Revil et al., 2010). Potential field methods
 67 (gravity, magnetic and self-potential measurements) can be used to some extent to
 68 characterize the subsurface of geothermal fields through density, magnetization, and fluid
 69 flow tomography (Revil et al., 2008; Aizawa et al., 2009; Okuma et al., 2009; Linde et al.,
 70 2014). However, because of the strong non-uniqueness in the inverse solution, these methods
 71 are generally not considered as stand-alone methods (e.g., Onizawa et al., 2009). In the last
 72 decade, there have been a lot of works to merge geophysical methods together through joint
 73 inversion methods (e.g., through the cross-gradient approach, see Moorkamp et al., 2010). We
 74 are interested to use below two recently developed techniques, image-guided inversion and
 75 interpolation (Ma et al., 2012; Zhou et al., 2014) and data fusion with scaled attributes (Revil
 76 et al., 2013), to combine geological and geophysical methods to track the plumbing system
 77 associated with the deepest hot spring in the world located in Pagosa Springs, Colorado, USA.
 78 Image-guided inversion provides a way to introduce structural elements such as facies
 79 boundaries in the inversion of the geophysical data. Image interpolation is a technique used to

80 refocus a blurred image. Data fusion will be used to define a geothermal index characterizing
81 the occurrence of the hydrothermal plumbing system.

82 Located approximately 325 km southeast of Denver, Colorado, and 30 km north of the
83 New Mexico border, Pagosa Springs houses a promising geothermal resource in southwestern
84 Colorado. In the heart of downtown Pagosa Springs, on the southern edge of the San Juan
85 River, lies the second largest hot springs in Colorado after Glenwood Springs (Warren, 1986).
86 This spring, known as the Great Pagosa Hot Spring or Big Spring, is located at an elevation of
87 2158 m. Surface temperatures of the Big Spring have been measured to be around 58°C with
88 flow rates up to 1 cubic meter per minute (Barrett and Pearl, 1976; Warren, 1986). The
89 maximum depth of the conduit of the Big Spring is unknown, but according to a test
90 conducted in August 2011, it is greater than 305 m. This makes the Big Springs as the
91 deepest geothermal hot spring in the world (see
92 [http://www.guinnessworldrecords.com/news/2011/9/colorado-claims-deepest-geothermal-hot-](http://www.guinnessworldrecords.com/news/2011/9/colorado-claims-deepest-geothermal-hot-spring-record)
93 [spring-record](http://www.guinnessworldrecords.com/news/2011/9/colorado-claims-deepest-geothermal-hot-spring-record)).

94 Historically, the springs were used by the Ute Native Americans as healing waters and
95 later as bath houses and spas. The name “Pag Osah” means literally “healing waters” in Ute
96 (Warren, 1986). Since 1982, the town of Pagosa Springs had a small district heating system
97 installed that provides heat to some of its buildings. According to Vail (1896), The Big
98 Spring has discharged a large quantity of hot water over the surface as recorded by the
99 travertine mounds located around the spring. The flow of water through the spring may have
100 dropped however because of the thermal wells drilled nearby in the area (approximately 27
101 thermal wells reported by Galloway, 1980).

102 In order to understand better the plumbing system of the Big Spring and the origin of
103 the thermal water, we conducted three geophysical surveys throughout the Pagosa Springs
104 area in May 2012, October 2012, and May 2013. These geophysical surveys complement

geological and geophysical studies made earlier in this area back in the 1970s and 1980s by the Colorado School of Mines (Warren, 1986) and a seismic reflection survey performed in 1977 by GFI, Inc. Conflicting mapping of the faults in Pagosa Springs comes from the Mancos shale, a Cretaceous shale that covers the area of interest and complicates the tectonic interpretation of surface features.

In the present study, geophysical techniques were used to gain additional insights about the position of the faults in this area and the plumbing system of the hot springs. We used direct current (DC) electrical resistivity and self-potential methods, which are known to be effective tools to localize faults (Storz et al., 2000; Gélis et al., 2010; Richards et al., 2010) and to decipher their impact on groundwater flow (Revil and Pezard, 1998; Ball et al., 2010; Aizawa et al., 2009; Vogt et al., 2014). We also reinterpreted a vintage seismic profile from Warren (1986). We will use recently developed techniques of data fusion with attributes (Revil et al., 2013) and image-guided inversion using geological models, i.e., geologically-constrained geophysical inversion (Zhou et al., 2014).

2. Background

We first provide the geological background in the Pagosa Springs area in order to identify the general fault networks in this region, their origin, and the possible flow paths between potential heat sources and the hot springs located in Pagosa Springs. The second part of this section summarizes some of the works done in Pagosa Springs that can be used to map the position of the tectonic faults.

2.1. Geological Background and Fault Networks

We will see later that the origin of the thermal water of Pagosa Springs cannot be understood without a broad understanding of the geology and tectonics of the region. Pagosa

Springs lies on the Archuleta anticlinorium (Figure 1), a complex 120 km long anticline characterized by a network of mostly northwest-southeast trending folds and normal faults with some additional northeast-trending transverse faults (Wood et al., 1948; Ryder, 1977). The anticlinorium serves as a border between the San Juan Basin to the southwest and the San Juan Mountains to the northeast (Figures 2 and 3). Beneath the San Juan Mountains rests another sedimentary basin called the San Juan Sag (Figure 3). The anticlinorium extends from the San Juan Uplift northwest of Pagosa Springs to the southeast into northern New Mexico where it terminates at the Nacimiento Uplift, Chama Basin, and Brazos Uplift. These structural features resulted from the Laramide Orogeny that formed the modern day Rocky Mountains.

The uplift of the Colorado Plateau initiated the Laramide Orogeny caused a total rise of about 3 kilometers of the whole area by the end of the Miocene. However, localized uplifts varied in magnitude both spatially and temporally, creating highlands and lowlands, which deformed the pre-existing sedimentary packages. According to Brister and Chapin (1994), the anticlinorium resulted from loading of sediments in the San Juan Basin and the San Juan Sag (shown in Figure 2) and compression from the northeast during the Eocene in the latter part of the Laramide. However, it is unclear if other regional stresses also contributed to the formation of the anticlinorium. In a previous mountain building sequence, called the ancestral Rocky Mountain Orogeny, the location of the anticlinorium lay on the southwestern boundary of the Uncompahgre uplift. This northwestern trend of the anticlinorium is likely influenced by faults in the Precambrian basement resulting from pre-Laramide tectonics. Brister and Chapin (1994) illustrated this basement faulting in their cross section AA' shown in Figure 2 (see position in Figure 1). A regional profile is also shown in Figure 3.

The fault and fractures systems present on the Archuleta Anticlinorium emerged from a combination of competing regional stresses. Since the basins and uplifts exist on opposite

sides of each other, the anticlinorium experienced divergent stresses caused by the stretching of sedimentary packages across the uplifting basement material. This resulted in a series of mostly normal faults in the area that are nearly vertical at the ground surface. The faults in the northwest-southeast orientation along the anticlinorium likely resulted from the Eocene northeasterly compression and basin loads. The Victoire and Eight Mile Mesa faults that will be discussed extensively in this work are two of these normal crustal faults.

Another fracture system strikes in the north to northeasterly direction. According to Galloway (1980), these faults are most likely connected with the Miocene dike intrusions, indicated by the displacement of these dikes. According to Baltz (1967), the dikes are likely a result of stresses unrelated to the Laramide tectonics, which Brister and Chapin (1994) suggest ended prior to the Oligocene. Additionally, there is another dominant fracture system trending in the northeasterly direction. These predominant regional fracture systems were mapped by Kelly and Clinton (1960) and are overlain on the Laramide structure map in Figure 1.

These two regional fracture systems seem to meet at Pagosa Springs forming a Y-shape as shown in Figure 4. This feature can be inferred from the regional geomorphology through the relationship of topography with the underlying geology and faults or the presence of quaternary deposits from ancient river beds. In the southern portion of the study area, the local hills trend in a north-south orientation. However, in the northwest and northeast sections of the study area, the hills trend in the northwest-southeast and northeast-southwest directions, respectively. Streams and rivers also appear to follow these trends, most notably the San Juan River which makes a drastic bend at the Big Spring (Figure 4). This intersection of faults controls locally the position of the Big Spring and the discharge of thermal water in this area. However, the available geology maps for the Pagosa area do not show the presence of faults at the main spring location because of the lack of surficial evidences to support their existence.

The most prominent fault in the area is the Eight Mile Mesa Fault, approximately 2 miles southwest of the Pagosa Springs and striking in the northwest-southeast direction. This fault is shown in the geologic map in Figure 4 and the geologic cross section in Figure 5 (whose location is partially marked in Figure 4 with BB'). In addition to the Profile BB', we established three additional geological cross-sections based on the outcrops and depending on the faults and the displacements of the outcropping geological layers. The position of these three profiles EE', FF', and GG' is shown in Figure 4 and the corresponding geological cross-sections are shown in Figure 6. These profiles also demonstrate the existence of the Victoire Fault, another major tectonic feature of the hot spring area. We will bring geophysical evidences supporting the existence of these two faults in the next sections.

2.2. Lithology

Erosion stripped away the early Laramide formations on the Archuleta Anticlinorium and deposited late Upper Cretaceous and Paleocene sediments such as the Kirtland Shale and Animas Formation, respectively, into the San Juan Basin and San Juan Sag (Brister and Chapin, 1994 and Figure 3). Note that none of these formations appears in our area of investigation, either on the surface or at depth. This erosion exposed an older Cretaceous sedimentary package including the Mancos Shale and Dakota formations in the Pagosa Springs area, which Galloway (1980) identifies as the two near surface aquifers in the region.

The Dakota formation's lower and upper members are medium to coarse-grained and conglomeratic and fine to medium grained sandstones, respectively. The Mancos Shale (typical thickness of 680 ± 40 m) is a dark-grey to black fissile shale interbedded with thin discontinuous sandstones and limestones. A layer of carbonaceous, dark shale separates these sandstone members (Wood et al., 1948; Galloway, 1980). The Mancos shale is divided into upper and lower members based on the amount of calcareous materials. The Jurassic

sedimentary packages below the Dakota sandstone and Mancos shale formations are the Morrison, Wanakah, and Entrada formations. The sedimentary formations of the Pagosa area are relatively uniform in thickness and generally dip gently (5-10°) to the northeast (Figures 5 and 6a). Below the Jurassic formations lies the Precambrian basement rock, which is at a depth of 417.9 meters at the well P-1 drilled in Pagosa Springs.

2.3. Hydrogeological Background

The O-2 and P-1 wells were drilled during the summer of 1978 and provide key information regarding the flow of water in the area of Pagosa Springs (Galloway, 1980). Severe caving occurred below 177 m in Well O-2 (which reaches a total depth of only 240 meters). Well P-1, drilled approximately 28 meters northeast of O-2, was successfully completed to a depth 452 meters (Galloway, 1980). Figure 6b shows a detailed stratigraphic column of the P-1 well. At approximately 78 meters in the Dakota sandstone, a small displacement fault was encountered with water at 60°C and flow rates between 5.7 to 7.6 m³ per minute. Temperature profiles acquired at both O-2 and P-1 are shown in Figure 7. The Dakota formation contains the water with the highest temperatures. The temperature decreases with depth and becomes isothermal in the lower parts of the Morrison down into the Precambrian basement. This is significant because it shows that the Dakota formation provides, at least locally, fluid flow paths for geothermal waters. According to Galloway (1980), the Dakota sandstone seems to be fed locally by a vertical feature (possibly the Victoire fault or a fault associated with the Victoire Fault as discussed below). Other wells in the Dakota sandstone formation around Pagosa Springs have encountered only cold groundwater. This shows that the Dakota formation is not the geothermal reservoir in this area.

The movement of hydrothermal fluids in the Pagosa Springs geothermal field appears

primarily fault-controlled rather than flowing laterally for any extent through the Dakota formation. Galloway (1980) concluded this due to the confined lateral extent of warm to hot water within the fringe of downtown Pagosa Springs. Wells drilled and completed in the Dakota outside of this extent in the Pagosa area had only cold water. Additionally, geochemical analyses of the hydrothermal fluids suggests that the water flows through Precambrian crystalline rock based on the presence of relatively high concentrations of boron, fluoride, and lithium (Galloway, 1980). The high total dissolved solids and an isotopic analysis of oxygen-18 and deuterium (based on the global meteoric water line defined by Craig, 1961) indicate a meteoric origin for the hydrothermal fluids (Table 1). Geochemical data of the Big Spring from two Colorado Geological Survey studies (Barret and Pearl, 1976; and Cappa and Hemborg, 1995) is shown in Table 1, which reasonably agrees with geochemical data collected from both the Big Spring and wells O-2 and P-1 presented in Galloway (1980).

In the Pagosa geothermal field area, there are no magmatic intrusions that are young enough to supply heat to the system. This implies that the likely source of heat to this geothermal system comes from deep circulation through cutting crustal faults and an elevated geothermal gradient. Plouff and Pakiser (1972) mentioned the existence of a large negative gravity anomaly, with an approximate areal extent of 100 by 150 kilometers, over the San Juan Mountains and possibly associated with a large concealed batholith genetically related to the San Juan caldera cluster. Lipman (1975) describes the history of the batholith as an inward progressive rise, differentiation, and crystallization based on evolution of the San Juan caldera clusters. Drenth et al. (2012) revisited the gravity data with improved processing and modeling techniques and corroborate the presence of a large batholith, although slightly smaller than the one proposed by Plouff and Pakiser (1972). A large batholith underneath the San Juan Mountains could account for the elevated geothermal gradient observed in

southwestern Colorado.

2.4. Previous Geophysical Studies

Prior to the recent investigation, the Colorado Geological Survey teamed with the Colorado School of Mines in 1977 and 1978 to study the district geothermal system at Pagosa Springs. This was part of an ongoing effort to evaluate geothermal potential throughout Colorado. The surveys included soil mercury tests, seismic reflection, and electrical resistivity. Also included in this study is a drilling program that yielded information regarding heat flow, chemical composition of thermal waters, mineralogy and petrography, stratigraphy, and hydrogeology of the study area (Galloway, 1980; Warren, 1986). Two successful seismic reflection surveys were performed in 1977 by GFI, Inc, Colorado (Warren, 1986). The longest profile is approximately eleven kilometers in a north-south orientation. It extends from just south of the Mill Creek valley north along the east side of the San Juan River, through downtown Pagosa Springs, and continuing north to 4 Mile Road as shown in Figure 8. The short profile, seismic line 3, ran in an east-west orientation and covered at least one mile across the Mill Creek valley. Vibroseis point 25 was used to tie the P-1 well with the seismic data and determine the reflectors for the Precambrian basement and Dakota sandstone. Warren (1986) interpreted a number of high angle faults including two at the opening of the Mill Creek valley, labeled in Figure 8 and discussed further later. Warren (1986) also interpreted a high angle fault that runs in the east-west direction at vibroseis point 25. However Galloway (1980) raised doubts about the existence of this fault because the seismic line is offset by a quarter mile at that location. From the results of the seismic surveys, Warren (1986) claims that high angle basement cored faults dominated the structure in the Pagosa Springs region.

3. Geophysical Investigations

3.1. Motivation

Electrical resistivity is an active galvanometric method providing tomograms of the electrical resistivity distribution of the subsurface. It has been used to image active and inactive volcanoes (e.g., Revil et al., 2010, Barde Cabusson et al., 2013; Komori et al., 2013a and references therein) and geothermal resources (e.g., Zhody et al., 1973; Soengkono et al., 2013). The presence of hydrothermal fluids tends to change the electrical properties of the subsurface either through alteration of rock and the water content of the rocks. Indeed, Resistivity is sensitive to porosity, water saturation, pore fluid salinity, the cation exchange capacity of the rock (related in turn to the clay mineralogy), and temperature (e.g., Revil et al., 1998; Revil et al., 2013). Therefore, contrasts in electrical properties are directly related to contrasts in these pore fluid and textural properties. Since these contrasts are often present in geothermal systems due to faulting and the circulation of hydrothermal fluids, ERT is a valuable tool in delineating geological structures and thermal boundaries (e.g., Revil et al., 2008; Bibby et al., 2009; and Hunt et al., 2009). However, electrical resistivity data does not provide information on fluid flow.

The self-potential method is the only geophysical method directly related to fluid flow in the subsurface. The self-potential method measures naturally occurring electrical field mostly associated with the flow of the groundwater (Corwin and Hoover, 1979) including at a depth of several kilometers (Jardani et al., 2008). Various geophysical studies have shown the usefulness of combining self-potential and electrical resistivity data to determine fundamental structures and fluid flow paths in geothermal fields (see Zhody et al., 1973; Hase et al., 2005; Revil et al., 2008; Aizawa et al., 2009; Vogt et al., 2014).

3.2. Field Investigations

The Colorado School of Mines conducted geophysical investigations of Pagosa Springs in May 2012, October 2012, and May 2013, which resulted in twenty-two DC electrical resistivity profiles and 1317 self-potential measurements. The survey locations were determined by orientation to mapped faults, proximity to hot springs and the warm spring, geological observations, previous geophysical surveys, and most importantly permit availability. The lack of profiles in the northern section of the study area was due to permitting issues. Locations of profiles with respect to key geographical features are shown in Figure 9. For every profile shown in Figure 9, electrical resistivity data was collected.

The electrical resistivity profiles were obtained using the ABEM Terrameter SAS 4000, cables with 20 meter takeout spacing, and stainless steel electrodes. For all resistivity surveys the Wenner- α array geometry was deployed due to its higher signal-to-noise ratio. The Wenner- α array is a classical resistivity array with equal distance between the electrodes A, B, M, and N (A and B denote the current electrodes and M and N the voltage electrodes) and the voltage electrodes are located in-between the current electrodes (e.g., Loke, 2011).

The self-potential measurements were performed using Geonosis Pb/PbCl₂ non-polarizing electrodes and a high impedance multimeter with an internal impedance of 100 M Ω . For the larger surveys, self-potential measurements were taken at the same locations as resistivity stations, using 20 meter station spacing. Reference electrodes were buried close (<10 meters) to the first station of a profile. If the reference location was moved due to cable length, then the reference electrode was re-buried close (<10 meters) to the last measurement made using the previous reference. Whenever the reference electrode was moved, 2-3 stations were re-measured to allow for re-referencing of the self-potential data and avoid accumulating errors along the profiles.

3.3. Methodology for Merging Geophysical and Geological information

The methodology we used to merge the geophysical and geological information is shown in Figure 10. We start with a classical Gauss Newton inversion of the geophysical data (see Appendix A) and an initial guess for the geological cross-section (based on outcrops, well data, and the general geology of the area or eventually a migrated seismic section, see Zhou et al., 2014). The geological map is eventually revised a first time to incorporate the geophysical information. Then, the revised geological map can be used again to invert the geophysical data using image-guided inversion. In geologically-constrained geophysical inversion, each geological facies of the geological map is grey-coded (from 0 to 1) and a structural tensor is determined to describe the textural character of the image based on the grey scale gradients. For resistivity tomography, the grey scales can be obtained by scaling the resistivity according to $0 \leq \mathfrak{I}_i = \log_{10}(\rho_i) / \log_{10}(\rho_m) \leq 1$ where the resistivity $\rho_i \in \{\rho_0; \dots; \rho_i; \dots; \rho_N\}$ represents the means of the prior model for the conductivity of each facies, ρ_m denotes the resistivity of the facies with the highest resistivity $\rho_m = \text{Max}\{\rho_0; \dots; \rho_i; \dots; \rho_N\}$. $\mathfrak{I}_i = 0$ is coded in white whereas $\mathfrak{I}_i = 1$ is coded in black and intermediate values are coded in grey scales.

The structural tensor is used to guide the inversion of the resistivity data by modifying the model covariance matrix in the regularizer (see appendix A). After convergence (still using a Gauss-Newton algorithm), the result of the tomogram is submitted to a final step, image-guided interpolation. This approach is based on solving an inverse pseudo-diffusion problem to refocus the tomogram using the structural information contained in the geological image. This technique is usually used in image processing to refocus blurred images. We use an inverse approach to refocus the tomogram as explained in details in Zhou et al. (2014). The final tomogram is finally reinterpreted to provide a final geological cross-section of the subsurface. This approach is used for tomogram PGSA-0. Therefore, the final tomogram in

Figure 11 represents a compromise between an initial guess for the geological cross-section and the geophysical data.

For the self-potential inverse problem, we are looking to invert the volumetric source current density that can explain the observed self-potential data observed at a set of stations taken along the resistivity profile and given a resistivity model. The inversion of the self-potential data is discussed in Appendix A (see Jardani et al., 2008, and Rittgers et al., 2013, for a complete description of the method followed in this paper).

3.4. Use of Combined Attributes

Revil et al. (2013) recently suggested the use of attributes (scaled from 0 to 1) to merge the information between self-potential and resistivity tomography and to image specific targets (in their case, a coal seam fire). The rationale for this choice is that both self-potential and resistivity data are not unambiguous descriptors of geothermal flow. In our case, the recorded self-potential data can be a combination of streaming potential data associated with the flow of the groundwater and low-resistivity areas can be corresponding either to our target (low resistivity because of the hot water, high water content, or high salinity) or to areas rich in clays (e.g., the Mancos shale). In this latter case, however, shales have low permeability values and would not be the source of any current density associated with the flow of the groundwater. Therefore, we need to define a combination of attributes that uniquely defines the geothermal target under consideration. For instance if we consider N types of scaled direct attributes $\mathfrak{S}_i$ defining a target, the combined attribute of the target is given by,

$$\mathfrak{S} = \prod_{i=1}^N \mathfrak{S}_i . \quad (1)$$

We can combine them to generate a Normalized Geothermal Index (NGI), $\mathfrak{S}$, comprised between 0 and 1 with 1 indicating a high probability of finding the geothermal target. We will

use this approach to analyze the data of the PGSA-02 profile. The thermal part of the target needs however to be ascertain through in situ sampling of the pore water using at least one well along the flow path. In the present case, the discharge of thermal water from the Eight Mile Mesa Fault further south is an indication that the groundwater circulation associated with this fault corresponds to thermal water.

4. Results

4.2. Application to Profile PGSA-02

PGSA-02 is a 6.4 kilometer long resistivity profile performed approximately 2 kilometers south of the city of Pagosa Springs (see position in Figure 9). This profile includes 2843 apparent resistivity measurements derived from the Wenner- α array for a total of 318 electrodes with a 20 meters spacing along the profile. The regular grid used for the inversion corresponds to 4147 model parameters on a rectilinear mesh. The resistivity tomogram PGSA-02 obtained using a Gauss-Newton inversion with isotropic smoothness and the L2 norm together with a preliminary geological cross-section are shown in Figure 11a and 12b, respectively. Figure 12a shows the self-potential data along this profile. The image-guided inversion after image-guided interpolation is shown in Figure 12b. The image-guided inversion (before interpolation) has a data misfit equal to 12.4% (defined in Appendix A).

The resulting resistivity tomogram shows very clearly the presence of the tectonic faults including the Eight-Mile Mesa Fault, the San Juan Fault, and the Victoire Fault as well as the Stinking Springs anticline and the different geologic formations (Figure 13). One of the units in this sedimentary package, the Cretaceous-aged Dakota formation, is a package of highly silicified sandstones interbedded with thin shale beds that has locally a high density of fractures, sometimes filled by silica as discussed above. This formation appears highly resistive ($>500\text{-}1000\text{ Ohm m}$) in contrast to surrounding units. The resistivity contrast

between this unit and the overlaying conductive Mancos Shale provides the means to identify individual units as well to identify new faults through sharp lateral contrasts in resistivity. The Mancos shale is typically highly conductive (in the range 10-50 Ohm m). The Morrison formation, a Jurassic-aged package of sandstones interbedded with frequent shale layers, underlies the Dakota Sandstone. This formation appears in the ERT profile as more conductive in its upper section (which is very clayey) and with a resistivity similar to the Dakota sandstone in its lower section.

The self-potential data needs to be detrended before inversion to focus on the negative self-potential anomalies associated with the lateral transfer of fluids along the faults. The detrended self-potential profile (using a second-order polynomial) is shown in Figure 14a along with the predicted data from the inversion. The inversion of the self-potential data is shown in Figure 14b (L2 norm, first iteration). We see that most of the source current is localized on the hanging wall of the Eight Mile Mesa Fault. There is no self-potential source of electrical current associated with the Victoire fault, where it crosses the CC' profile. The normalized geothermal index is shown in Figure 15 confirming that only the Eight Mile Mesa Fault seems to carry thermal water at the position of the profile. One implication is that most of the hot water circulating along the Victoire fault further north has already discharged to the Big Springs (with a small contribution to the warm spring). Regarding the Eight Mile Mesa Fault, it seems that the flow is focused on the hanging wall of this normal fault. As discussed by Withjack et al. (1995), the hanging wall of normal fault is often fractured. This network of fracture can provide a highly permeable pathway for the flow of the thermal water.

4.2. Other Profiles

We present three additional resistivity profiles (Figures 16, 17, and 18) to support the existence of the Victoire fault and its extension to the northwest of Pagosa Springs. Figure 19

corresponds to a reinterpretation of the geological profile that is also compatible with the geophysical data. The resistivity tomograms indicate clearly the existence of a normal fault (that we named the Victoire Fault) close to the Big Spring, which extend both south and north from the hot springs. This fault was not mapped by Galloway (1980) but was reported by Warren (1986). Like the Eight Mile Mesa Fault, it is likely that the hanging wall of this fault is heavily fractured allowing the hot fluids to move upwards towards the surface, especially at the junction of the Victoire Fault and the transverse Faults A and B.

5. Discussion

According to Warren (1986), the oxygen 18-deuterium ratio in the thermal waters indicates a meteoric origin. The source for the Big Springs may correspond to the San Juan Uplift and its postulated underlying batholith, located 60-100 km in the Northwest of Pagosa Springs. This is consistent with a higher shallow heat flow observed in the northwest of Pagosa Springs along McCabe Creek (see Warren, 1986, his Figure 17). We believe that the hot water flows along a network of mostly northwest-southeast trending folds and faults associated with the Archuleta anticlinorium (Figure 1). This network of faults, serving as conduits, includes the Eight Mile Mesa and Victoire faults. In contrast, the northeast trending transverse fault (Faults A and B in Figure 4) are characterized by a lower heat flux anomaly (Warren, 1986). In addition, ground water hydrogeology along the southeast flank of the San Juan uplift (Northwest of Pagosa Springs, see Figure 1) indicates that the groundwater is flowing southeast parallel to the direction of the Archuleta anticlinorium (Ryder, 1977).

In the case of the Victoire fault, upwelling thermal water may be coming from flow along the fault in the basement since the reported depth of the Big Springs is lower than the Dakota sandstone formation (Figure 5b). We suggest that the fractured Dakota sandstone serves as a local aquifer for the hot water coming from the Victoire and Eight-Mile Mesa

454 faults. Local sealing above by the Mancos shale and below by the Upper Morrison formation,
455 both clay-rich, maintains the flow in the Dakota formation, which acts as a confined aquifer
456 until it locally outcrops.

457 The Big Springs and the warm springs are located along the Victoire fault, at the
458 intersection between this fault and two northeast-southwest trending faults named Faults A
459 and B, respectively (see Figures 4 and 9). Fault A lies along the San Juan River East of the
460 Pagosa Springs while Fault B is along Mill Creek. The intersection of these faults likely
461 creates a stress regime favoring the ascent of the hot water transferred along the Victoire fault
462 through a set of local fracture conduits. We do not know if Faults A and B are transmissive of
463 hot waters but from the lack of other thermal manifestations, this is unlikely. Our observations
464 on Profile CC' indicates that most of the hot water circulating along the Victoire fault is
465 upwelling at the Big Springs. In the case of the Eight-Mile Mesa fault, groundwater flow
466 seems to stay confined on the hanging side of the fault as shown in Figures 14 and 15.

468 **6. Conclusions**

469 In order to better constrain the flow of geothermal water, we have developed geology-
470 constrained geophysical inversion and the use of attributes to improve the detection of
471 geothermal targets. Our work provides evidences for the Victoire fault, a north-northwest
472 trending normal fault just to the east of the Big Spring of Pagosa Springs. We also suggest
473 that normal faulting in the Mill Creek Valley has down-dropped the area to the south. The
474 position of the Big Spring is located at the intersection of the Victoire fault and another fault
475 named Fault A with a northeast to West direction. The geothermal heat source can only be
476 located 60-100 km to the northwest to the Big Springs and corresponds likely to the batholith
477 shown in Figure 1. The localization of the low resistivity anomalies around the springs and
478 the high temperatures reached in wells around Pagosa Springs suggest a fault controlled

geothermal reservoir with the Victoire fault serving as conduit and feeding local fractures in the Dakota formation.

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Appendix A. Inversion of the Geophysical Data

In this appendix, we describe the basic equations used for the inversion of the DC resistivity and self-potential data. For DC resistivity, we minimize the following objective function P^β to find an optimal resistivity model $\mathbf{X}^*$

$$\mathbf{X}^* = \arg \min P^\beta(\mathbf{X}), \quad (\text{A1})$$

$$P^\beta(\mathbf{X}) = (\mathbf{G}(\mathbf{X}) - \mathbf{d})^T \mathbf{C}_d (\mathbf{G}(\mathbf{X}) - \mathbf{d}) + \beta \mathbf{X}^T \mathbf{C}_m \mathbf{X}, \quad (\text{A2})$$

where P^β is the sum of a data misfit function and a model objective function (a regularizer). In Eq. (A2), $\mathbf{X}$ denotes the resistivity model (log-resistivity distribution to enforce positivity of the resistivity of each cell), $\mathbf{G}$ denotes the non-linear forward operator allowing to go from the resistivity model to the apparent resistivity data (we use a finite difference approach), ($\mathbf{G}(\mathbf{X})$ represents the predicted data given a resistivity model $\mathbf{X}$), $\mathbf{d}$ denotes the vector of measured apparent resistivities (measured data vector), $\mathbf{C}_d$ denotes the inverse of the data covariance matrix (taken diagonal), β is the regularization parameter, and $\mathbf{C}_m$ is the smoothing matrix (inverse of the model covariance matrix). The minimization of the objective function,

$$\partial P^\beta(\mathbf{X}) / \partial \mathbf{X} = 0, \quad (\text{A3})$$

using the Gauss Newton algorithm leads to the following iterative solution,

$$\mathbf{X}_{i+1} = \mathbf{X}_i + d\mathbf{X} = \mathbf{X}_i + (\mathbf{J}^T \mathbf{J} + \beta \mathbf{C}_m)^{-1} \mathbf{J}^T \mathbf{C}_d [\mathbf{G}(\mathbf{X}_i) - \mathbf{d}]. \quad (\text{A4})$$

where i denotes the iteration number, $\mathbf{J}$ denotes the Jacobian matrix. During this process, β is adjusted using the L-curve approach to provide a solution that maintains the data fit and also honors the model constraints. Inversion is performed until a convergence criterion is chosen on the data misfit term until the norm $\|\mathbf{X}_{i+1} - \mathbf{X}_i\|_2^2$ reaches a threshold.

The relative Data Root Mean Square (DRMS) data misfit error is calculated as

$$\text{DRMS} = \sqrt{\frac{\sum_{k=1}^N \frac{(\mathbf{G}(\mathbf{X}) - \mathbf{d})_k^2}{d_k^2}}{N}}, \quad (\text{A5})$$

where N denotes the number of data.

Ground water flow drags the excess of charge contained in the diffuse layer coating the surface of the grains of a porous material. A generalized Ohm's law can be written for the total current density $\mathbf{J}$ (A m^{-2}) as $\mathbf{J} = \sigma \mathbf{E} + \mathbf{J}_s$, where σ denotes the low-frequency electrical conductivity of the porous material (in S m^{-1}), $\mathbf{E} = -\nabla \varphi$ (φ denotes the electrical potential expressed in V), and $\mathbf{J}_s$ is a source current density of electrokinetic nature (i.e., associated with the drag of the excess of charge in the diffuse layer as mentioned above). The constitutive equation can be combined with a conservation of charge equation written as $\nabla \cdot \mathbf{J} = 0$ to get the field equation for the electrical potential φ ,

$$\nabla \cdot (\sigma \nabla \varphi) = \nabla \cdot \mathbf{J}_s, \quad (\text{A6})$$

Equation (A6) can be written in a matrix form as $\mathbf{d} = \mathbf{K}\mathbf{X}$, where $\mathbf{X}$ is a mode vector containing the M source of the volumetric current density terms $\mathfrak{S} = \nabla \cdot \mathbf{J}_s$, $\mathbf{d}$ denote the vector of electric potentials φ observed at N self-potential stations taken along the resistivity profile (see for instance Figure 12), and $\mathbf{K}$ is the $N \times M$ kernel matrix. For the self-potential inverse problem, the objective function to minimize is expressed as

$$\mathbf{X}^* = \arg \min P^\beta(\mathbf{X}), \quad (\text{A7})$$

$$P^\beta(\mathbf{X}) = (\mathbf{K}\mathbf{X} - \mathbf{d})^T \mathbf{C}_d (\mathbf{K}\mathbf{X} - \mathbf{d}) + \beta \mathbf{X}^T \mathbf{C}_m \mathbf{X}, \quad (\text{A8})$$

where $\mathbf{C}_d$ correspond to the data ($N \times N$) weighting matrix, $\mathbf{C}_m$ to the ($M \times M$) model smoothing matrix (we use the Laplacian operator to ensure smoothness), and β is the regularization parameter determined from the L-curve approach. The least square solution corresponding to $\partial P^\beta / \partial \mathbf{X} = 0$ is given by

$$\mathbf{X}^* = [\mathbf{K}^T (\mathbf{C}_d^T \mathbf{C}_d) \mathbf{K} + \beta (\mathbf{C}_m^T \mathbf{C}_m)]^{-1} (\mathbf{K}^T (\mathbf{C}_d^T \mathbf{C}_d) \mathbf{d}). \quad (\text{A9})$$

We use the same data misfit as defined for the resistivity problem (see equation A5).

Inversion is performed until a convergence criterion is chosen on the data misfit term until the

543 norm $\|\mathbf{X}_{i+1} - \mathbf{X}_i\|_2^2$ reaches a threshold. As discussed by Jardani et al. (2008), the model
544 vector needs to be scaled to avoid favoring solutions close to the self-potential stations.
545

Tables

Table 1. Mean of select geochemical properties of hydrothermal fluids in the Big Springs.

Data from Barrett and Pearl (1976) and Cappa and Hemborg (1995). This water has a meteoric origin.

Property	Value (1)	Value (2)	Value (3)
pH	6.6 ± 0.2	6.5	6.5
Total Dissolved Solids (mg/L)	3189 ± 96.5	3400	3430
Conductivity (S/m)	0.459 ± 0.072	0.4340	0.39
Temperature (°C)	56 ± 1.6	54	60
Discharge (gpm)	248 ± 16	260	-
Boron (B - mg/L)	2 ± 0.22	2.3	-
Calcium (Ca - mg/L)	213 ± 44	230	390
Chloride (Cl - mg/L)	182 ± 4	180	100
Fluoride (F - mg/L)	4.7 ± 0.3	4.8	1.4
Lithium (Li - mg/L)	2.9 ± 0.2	2.9	2.4
Magnesium (Mg - mg/L)	19.4 ± 8	24	35
Silica (SiO ₂ - mg/L)	59 ± 4	59	20.7
Sodium (Na - mg/L)	760 ± 39	730	640
Sulfate (SO ₄ - mg/L)	1432.4 ± 75.6	1300	1720
Bicarbonate (HCO ₃ ⁻ - mg/L)	848 ± 3	856	540

(1) Data from Barrett and Pearl (1976) and Cappa and Hemborg (1995). Big Springs.

(2) Data from Galloway (1980). Big Springs.

(3) Data from P-1 (Galloway, 1980). Dakota formation.

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Figures

Figure 1. Map of the Laramide structures of the region surrounding Pagosa Springs (modified from Brister and Chapin, 1994). Overlain onto the structure map are dominant fracture systems mapped by Kelly and Clinton (1960) in light grey. Profile A to A' indicates the location of the cross section shown in Figure 2 crossing the Archuleta anticlinorium. The San Juan Mountains correspond to the area denoted as "San Juan Sag" on the map. The San Juan uplift corresponds to the potential heat source for the geothermal system feeding the Pagosa Springs. Converging arrows are used to define the synclines while the diverging arrows are used to describe the anticlines.

Figure 2. This cross-section AA' (modified from Brister and Chapin, 1994) cuts the Archuleta anticlinorium approximately perpendicular to its folding axis shown in Figure 1. At about the center of the cross section is the Precambrian basement fault believed to influence the location of the anticlinorium. Some of the main tectonic faults discussed in this work are parallel to the Archuleta anticlinorium. The San Juan Mountains correspond to the area denotes as "San Juan Sag" on the map.

Figure 4. Map of the geology in the study area adapted from Galloway (1980). Interpreted faults from this research are shown as a thick red dashed or dotted lines. The line marked on one end as B' shows the extent of the Galloway (1980) cross section shown in Figure 5. The Eight Mile Mesa fault and the Victoire fault continue further northwest in the direction of the Archuleta anticlinorium. Note that the Big Spring and the warm springs are both located at the intersection between the Victoire fault and lateral normal faults (Faults A and B, respectively). The dike is shown close to the warm springs (see also feature 3 in Figure 13).

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744 **Figure 5.** Profile BB'. **a.** The geological cross-section adapted from Galloway (1980).
 745 Formations are general uniform in thickness, unless eroded at the surface, and dip at all
 746 locations northeast of the anticlinal axis. **b.** Modified cross-section. The faults in the cross
 747 section are from mapped faults in the Pagosa Springs area. We have added the vertical conduit
 748 of the Big Springs (see the red line beneath the Big Springs) indicating that the Big Springs
 749 have a root located below the Dakota formation. **c.** The picture shows the Big Springs in
 750 1874. The mound around the pool is a travertine, which is covering an area of $\sim 0.07 \text{ km}^2$
 751 (Picture credit: T.H. O'Sullivan, The hot springs for which Pagosa Springs, Colorado, is
 752 named, modified from ARC524217).
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756 **Figure 6.** Geological cross-sections around the Eight Mike Mesa and Victoire Faults. **a.**
757 Profiles EE', FF', and GG' (see position in Figure 4). These profiles were determined using
758 geological information including outcrops, pending faults, and the thickness of the geologic
759 formations. **b.** Stratigraphic column of the P-1 well just to the northwest of the Big Spring
760 with rounded value for the depth of the basement. The location of the well is shown in Figure
761 4.
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Figure 7. Temperature profile for wells P-1 and O-2 located north from the Big Springs. The location of the P-1 well is shown in Figures 4 and 9 but not O-2 (780 feet, 240 m) which is approximately 28 meters southwest of P-1. The temperature profile indicates subhorizontal ground water flow through the Dakota Sandstone and Mancos Shale formations. **a.** Thermal Profiles. **b.** Picture of the fractured Dakota sandstone showing silica precipitation (in white) along the cracks in the upper member of the Dakota formation.

Figure 8. This image shows the long north-south seismic profile acquired in 1977 by GFI. According to Warren (1986), the data were processed using standard demultiplexing, sorting, stacking, deconvolution, and filtering operations. The dashed line indicates the top of the Dakota Sandstone, the solid line shows the top of the Precambrian Basement (in yellow), and the vertical dashed lines mark placements of faults as interpreted by Warren (1986). The warm and Big Springs are associated with the two faults A and B characterized by the strongest displacement. The Dakota sandstone outcrop south of Fault B. In the insert, the square has a side of 1.60 km.

Figure 9. Position of the resistivity profiles together with the topography of the area (thick lines color-coded according to the time at which the profiles were acquired). The profile CC' corresponds to PGSA-02, which is a 6.4 km resistivity and self-potential profile. Note that Reservoir Hill and McCabe Creek are associated with terrace deposits. The profiles are made of filled circles (see box) using one filled circle per electrode.

Figure 10. Geophysical inversion technique used in the present study **a.** Use of image guided inversion and interpolation to merge geological and geophysical information together using a Gauss-Newton technique. The interpolation is an image-guided interpolation used to refocus the image as described in Zhou et al. (2014). **b.** Data fusion using a combination of attributes defined from a set of tomograms (for instance self-potential and resistivity tomography). The quantity $\mathfrak{I}$ defines a scaled attribute comprised between 0 and 1.

Figure 11. The electrical resistivity tomogram PGSA-02 (profile CC' in Figure 9). **a.** Classical Gauss-Newton inversion of the apparent resistivity data with isotropic smoothing for the regularizer. **b.** Preliminary geological cross-section used for the image-guided inversion of the resistivity data. This cross-section shows the Stinking Springs anticline (trending N40°W) and some of the major faults known in this area. This cross-section is used to provide an anisotropic smoothing in the inversion of the apparent resistivity data.

Figure 12. Electrical resistivity tomogram PGSA-02 (CC') and self-potential profile along Profile CC' (acquisition performed in May 2013, total length of ~6.34 km). **a.** Self-potential profile. All self-potential data were referenced to a station located five meters south of the first station of the profile. The positive peak at 2100 m (affecting two stations) is likely a very local source (measured twice) and is not accounted for in the inversion of the self-potential data. A1 and A2 represent two negative self-potential anomalies with respect to the background. **b.** Resistivity tomogram using the methodology developed by Zhou et al. (2014) for image-guided inversion and interpolation.

Figure 13. Interpreted resistivity profile together with the pictures of several outcropping formations. The resistivity tomogram shows the position of the different faults including the Eight-Mile Mesa Fault and the Victoire Fault. Picture (1): The upper member of the Dakota Formation, viewed from the North. Picture (2): The lower member of the Dakota Formation outcropping close to the San Juan River. Picture (3): A north-trending andesitic dike cutting the Mancos Shale (view is due North West, 1.6 km south from the Big Springs).

Figure 14. Tomography of the self-potential sources. **a.** Detrended self-potential profile (in red) and fit of the data (in blue). **b.** Volumetric source current density (in mA m^{-3}). According to this model, the main self-potential current source along this profile is localized in the hanging wall of the eight Mile Mesa fault.

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Figure 15. Distribution of the normalized geothermal index. This index reflects the possibility of warm water flowing normal to the profile in the direction of the head gradient. The flow of hot water seems localized only on the hanging wall of the Eight Mile Mesa fault. There is no flow associated with the Victoire Fault, which may imply that the hot water flowing in the Victoire fault has discharged at the Big Springs. The flow direction is inferred from the piezometric data reported in Galloway (1980).

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858 **Figure 16.** Profile PAGO-02. **a.** Result of the classical Gauss-Newton inversion with isotropic
 859 smoothing. **b.** Image guided inversion and interpolation and its interpretation. The tomogram
 860 may show a fractured upper member of the Dakota Formation (Kd) with antithetic faults
 861 located in the hanging wall of the Victoire normal fault. Note that the Mancos Shale (Km) is
 862 outcropping on Reservoir Hill but can be locally masked by the terrace deposits shown in
 863 Figure 4. Note that the thick red line beneath the Big Springs denotes the "conduit" for the hot
 864 springs.
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Figure 17. Profile PAGO-06. **a.** Result of the classical Gauss-Newton inversion with isotropic smoothing. **b.** Image guided inversion and interpolation and its interpretation.

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Figure 18. Profile PAGO-08. **a.** Result of the classical Gauss-Newton inversion with isotropic smoothing. **b.** Image-guided inversion and interpolation and its interpretation.

Figure 19. Geological profiles compatible with the outcrops and the resistivity tomograms along the profiles Pago 02 and Pago 06. The big Spring seems to be fed by the Victoire Fault in the crystalline basement.