USING GEODETIC DATA TO UNDERSTAND HYDROTHERMAL FLUID FLOW

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ABSTRACT
Fluid flow in hydrothermal and magmatic systems frequently induces observable surface deformation. Such deformation can be used to better understand the factors controlling flow, for example, the role of faults as conduits or barriers to flow. In this paper we describe how surface deformation data can be used to obtain qualitative and quantitative information about subsurface flow associated with hydrothermal systems. Examples from hydrothermal/magmatic systems at Long Valley and Yellowstone are described. At Long Valley, GPS, levelling, and EDM data are used to determine the distribution of volume change in the subsurface. The source model contains two prominent features beneath the resurgent dome: one following the regional trend of north-south faults, and another parallel to a strike-slip fault that is the locus of most of the seismic activity in the calderas south moat. An inversion of Interferometric Synthetic Aperture Radar (InSAR) observations from the Yellowstone region indicated that two cross-cutting fault systems are interacting with an underlying magma body to produce rapid and complicated surface deformation.

INTRODUCTION
The flow of fluids associated with magmatic and geothermal systems is often poorly understood. Typically, fluid flow is constrained by sparsely distributed well data. The sampling provided by well data is complicated, depending on the hydraulic connection between flow in the well and flow in the surrounding formation. Thus, well data do not provide a simple, detailed picture of subsurface flow and transport. Frequently, micro-seismic studies are used to better understand subsurface flow in geothermal areas. However, the relationship between micro-seismicity and pressure and temperature changes in the subsurface is complicated and strongly influenced by mechanical heterogeneity and regional and local stress fields. Time-lapse geophysical data could potentially provide high-resolution, three-dimensional images of changes in seismic velocity and electric conductivity that may be related to saturation and pressure changes at depth. Currently, primarily due to their expense, such surveys are not conducted at geothermal fields.

Geodetic data provide another source of information related to the movement of fluids at depth. Geodetic observations are of somewhat intermediate resolution. That is, they do not provide the high-resolution of active time-lapse geophysical surveys but they typically provide more constraints than well information. In addition, geodetic data are often directly related to volume change at depth which may be interpreted in terms of fluid movement and temperature changes. Thus, geodetic observations, such as Global Positioning System (GPS), leveling, EDM, and Interferometric Synthetic Aperture Radar (InSAR) measurements, can provide an important tool for understanding the factors controlling fluid movement in a hydrothermal system. In this paper we outline an approach for using surface deformation data. The potential of this approach is illustrated by
applications at the Long Valley Caldera in California and the Yellowstone Caldera in Wyoming.

METHODOLOGY

Inference of Subsurface Volume Change

Our goal is to image processes related to hydrothermal fluid flow. Fluid mass and temperature changes within the Earth give rise to effective volume changes. Such volume changes induce deformation which results in surface displacement. Specifically, volume change within an elastic or poroelastic Earth gives rise to the displacement of its surface (Mogi 1958, Segall 1985). The relationship between a distribution of volume change $\Psi(y)$ within the Earth and the $i$-th component of surface displacement, $u_i$, is given by the integral

$$u_i(x) = \int \limits_{V} g_i (x, y) \Psi(y) dy$$

where $g_i$ is the Greens’ function or point source response. For a homogeneous elastic half-space it is given by (Vasco et al. 1988)

$$g_i(x, y) = \frac{(V+1)(x_i - y_i)}{3\pi S^3}$$

where $V$ is the Poisson’s ratio of the half-space and $S$ is the distance between the observation point and the location of the point volume change. The integral expression for $u_i$ provides a linear relationship between the $i$-th component of displacement and the volume change in the subsurface.

As described in Vasco et al. (2000) we may convert the above integral into a sum by representing the subsurface volume by a finite number of $M$ non-overlapping cells. Each cell may undergo a distinct fractional volume change, denoted by $\rho_j$, for the $j$-th cell or grid block. The contribution to the $l$-th component of displacement at $x$ due to the fractional volume change of the grid block, $\rho_j$, is an integral over the volume of the $j$-th grid block $V_j$,

$$\rho_j(x) = \psi_j \int \limits_{V_j} g_l (x, y) dy .$$

The total $l$-th component of displacement at an observation point $x_i$ is simply a sum over the complete set of $M$ grid blocks

$$\rho_l(x_i) = \sum \limits_j L_{ijl} \psi_j$$

where $L_{ijl}$ is the integral

$$L_{ijl} = \int \limits_{V_j} g_l (x_i, y) dy .$$

Given a number of displacement observations we may constrain the volume change in the subsurface. For a set of $N$ observations there corresponds a linear system of $N$ equations. Let us scale each equation by its standard error $\sigma_i$ and consider the least squares minimization of the residuals. That is, the minimization of the sum

$$\Pi_d = \sum \left[ d_i - \sum \Gamma_{ijl} \psi_j \right]^2$$

where

$$d_i = \frac{\rho_i}{\sigma_i}$$

and

$$\Gamma_{ijl} = \frac{L_{ijl}}{\sigma_i}$$

are the scaled data and matrix coefficients, respectively.

Sparse geodetic observations provide limited depth resolution and there are trade-offs between the shape of the volume change and the depth of the volume change (Dieterich and Decker 1975). Thus, the system of linear constraints on the volume change model is likely to be ill-conditioned and any solutions will be unstable with respect to perturbations in the observations and the coefficients. To address this instability, the constraints provided by the data are augmented by regularization terms that bias the model towards a smoothly varying solution (Menke 1984). The model roughness is measured by the magnitude of its spatial derivative vector

$$\Pi_r = \sum \left( D_j \psi_j \right)^2$$

where $D_j$ is a finite difference approximation to the spatial derivative of the model. That is, an operator which differentiates the volume change in space, as discussed in Menke (1984). The penalized misfit is given by

$$P = \Pi_d + W_r \Pi_r$$

where $W_r$ is the weight which controls the relative importance of fitting the observations and obtaining a smooth model.

APPLICATIONS

Long Valley Caldera

The Long Valley volcanic area (Fig. 1) has been active for the past 3 million years. A massive eruption 0.76 Ma ago resulted in the deposition of 600 cubic kilometers of Bishop Tuff and the
simultaneous subsidence of the magma chamber roof, creating the present 17-by-32 km oval depression of the Long Valley Caldera. The formation of the caldera was followed by resurgence and four major episodes of volcanic activity (Hildreth 2004).

Recent unrest began in May of 1980 with a strong earthquake swarm at the southern margin of the caldera (Bailey & Hill 1990). The seismic activity was accompanied by the uplift of the resurgent dome in the central portion of the caldera. Substantial deformation was also observed during the summer and fall of 1997. Following that the caldera has been relatively quiet, with no significant deformation since the spring of 1998 ((Fig. 1; Langbein 2003).

Several models have been developed to explain the deformation and seismicity within the caldera (Sorey et al. 2003). Two-color EDM and leveling surveys indicate that the intrusion of a magma body beneath the resurgent dome, and right lateral strike slip within the south moat of the caldera are the principle sources of deformation (Langbein et al. 1995; Langbein 2003). Radar interferometry (Thatcher & Massonet 1997; Fialko et al. 2001), GPS surveys (Marshall et al. 1997) and gravity measurements (Battaglia et al. 2003b) confirm the intrusion beneath the resurgent dome. There is evidence that intrusion is also occurring beneath the south moat and Mammoth Mt (Sorey et al. 1993; Hill et al. 2003).

Figure 1. Long Valley caldera unrest. The plot shows the number of M>3.0 earthquakes, the uplift at the resurgent dome (benchmark W911) and at Tom's Place (several km outside the caldera), and the extension of the CASA-KRAK baseline (EDM), a proxy of resurgent dome vertical deformation.

In an effort to update the vertical deformation measurements within the caldera, 44 of the existing leveling monuments in Long Valley were re-occupied in July 1999 (Fig. 2) using dual frequency GPS receivers (Battaglia et al 2003a).

The source model is constructed by first sub-dividing the volume beneath the caldera into a set of non-overlapping cells or grid-blocks, as described in Vasco et al. (1988) and Vasco et al. (2000). Each block in the grid may undergo a distinct fractional volume change. The displacement at any point on the surface is the integrated response to changes over the entire grid. For the most part, surface deformation observations cannot resolve detailed volume change variations with depth (Dieterich and Decker 1975). For this reason, we prescribe the depth boundaries of our model grid. Based upon earlier point source investigations and physical considerations, we adopt a three layer grid with depth boundaries: 5-7 km, 7-9 km, and 9-11 km. Each layer is further sub-divided into a 41 (east-west) by 41 (north-south) grid of cells. Because the observed surface deformation is linearly related to the fractional volume change of each grid block, each observation contributes a linear constraint on the volume change of the cells in the grid (Vasco et al. 1988).

Given an adequate distribution of data we may estimate the volume change distribution within the
grid. Specifically, using a regularized, linear least-squares procedure we can solve for the spatial distribution of volume change (Vasco et al. 2000). The resulting distribution of fractional volume change in the uppermost layer of the model is shown in Figure 3.

![Figure 3](image)

**Figure 3.** Volume change in the depth interval of 5 to 7 km.

Dark reds and blues in Figure 3 signify greater fractional volume change. The largest fractional volume change forms a dominantly east-west body. In addition, a north-south trending component of fractional volume change roughly parallels a system of north-northwest trending faults within the caldera. This component extends from the east-west anomaly to the northern edge of the caldera. The volume change in the deeper layers (7-9 km, 9-11 km) is much smaller in amplitude and similar in pattern to the overlying layer shown in Figure 3, most likely due to the poor depth resolution provided by the surface deformation data.

![Figure 4](image)

**Figure 4.** Fit to data. Comparison between the single source and distributed source models.

The fits to both the EDM and GPS data are shown in Figure 4. For the most part the model fits the geodetic observations within their associated errors.

**Yellowstone Caldera**

The Yellowstone volcanic system is a tectonically active region, shaped by three caldera-forming eruptions (Vasco et al. 1990). Current volcanic activity is thought to be driven by a body of hot, crystallizing magma located at least 8 km deep. Based upon Interferometric Synthetic Aperture Radar (InSAR), the ground surface over the caldera has been observed to be in an almost constant state of motion (Wicks et al. 1999). For example, between 1992 and 1995 much of the caldera was involved in approximately 2 cm of subsidence per year (Figure 5). The surface deformation provides some clues to the structures controlling fluid flow at depth. In order to better understand the factors influencing fluid movement in the subsurface we applied the
inversion methodology described above to the InSAR data shown in Figure 5.

Figure 5. Range velocity between 1992 and 1995. The red colors signify motion away from the satellite due to the subsidence of the surface over the caldera.

Following the methodology outlined above we constructed a three-dimensional grid representing the possible volume change in the subsurface which is compatible with the observed surface deformation. The grid is composed of three layers 6-8, 8-10, and 10-12 km in depth. Each layer is sub-divided into a 41 by 41 grid of non-overlapping grid blocks. Using the least squares formulation outlined above we solve for the distribution of volume change in the grid which can explain the range change shown in Figure 5. The resulting model, shown in Figure 6, contains several interesting features.

Figure 6. Fractional volume decrease produced by an inversion of the range change observations shown in Figure 5.

The primary volume decrease underlies a northeast trending fault zone, the Elephant Back Fault Zone (EBFZ), which lies along the long axis of the caldera. This fault zone is currently active and offsets recent sediments. The fault zone connects the two resurgent domes within the caldera, which are signified by the closed curves in Figure 6. In addition to volume decreases beneath the EBFZ we observe a linear trend of volume decrease extending almost due north. This feature underlies a string of volcanic vents which fed post-collapse lava flows (stars in Figure 6). The linear trend is heading in the direction of the Norris Geyser Basin which had been the site of renewed hydrothermal activity. Furthermore, more recent InSAR observations indicate significant uplift at the Norris Geyser Basin and in surrounding regions.

CONCLUSIONS

We have presented models of subsurface volume change which are compatible with observed surface deformation data. The distributed source models are very general and exploratory in nature. They offer insight into the dynamics of fluid movement at the Long Valley Caldera and at the Yellowstone Caldera.

At Long Valley the subsurface fractional volume change is suggestive in that it underlies existing north-trending faults. In addition, the east-west component of volume change parallels a hypothesized east-west fault in the south moat region. Furthermore, the east-west component of volume change connects two sets of north-south trending faults. The more easterly set extends from the central
caldera to the south, the other set cuts northward through the interior of the caldera (Figure 3).

The model helps shed light on the relation between deformation and seismicity. According to the existing interpretation (e.g., Sorey et al. 2003), dominant sources contributing to the deformation and seismicity within the caldera include (Fig. 3): (1) aseismic inflation of a source centred at approximately 6 km beneath the resurgent dome; (2) seismic inflation of a deeper one (10-20 km) beneath the South Moat Seismic Zone (SMSZ); and (3) right-lateral strike-slip motion on a series of west-northwest-striking faults in the 10-km-wide SMSZ (Langbein 2003).

At Yellowstone, the inversion of InSAR data suggest that volume change is concentrated beneath the Elephant Back Fault Zone (EBFZ) which cuts across the caldera. This fault zone connects the two resurgent domes within the Yellowstone Caldera. The fault zone has been the site of recent deformation. In addition, the pattern of subsurface volume change suggests that a north-south oriented fault or fracture zone is controlling fluid flow above the Yellowstone magma body. The deformation at Yellowstone appears to be due to the interaction of the EBFZ, the north trending fault/fracture zone, and an underlying magma body.

The association of subsurface volume change with mapped faults and earthquakes may be due to existing faults acting as conduits for fluid and gas movement beneath the caldera. That is, the faults represent high permeability zones. Alternatively, the faults may represent zones of weakness (or mechanical heterogeneity) which concentrate deformation. Or both processes may be operating in concert, and faults may concentrate fluid and gas flow as well as concentrating deformation.

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