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Geothermal Reservoirs: Products of Cooling Plutons. DE-FG07-99ID13854

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Executive Summary

The goals of this project were to develop an in depth understanding of how geothermal reservoirs form and elucidate those features that could potentially be useful in exploration and development of additional energy reserves. Collaboration with Jeff Hulen, EGI helped closely coordinate theoretical concepts and computational experiments with geologic reality in fulfillment of the tasks for this project. Initial reconnaissance computations with Tom Brikowski, University of Texas were critical in realizing the final products of this project.

The products of this work contribute basic understanding of the dynamical conditions attendant to the formation of reservoirs in general and the Geysers reservoir in particular. The most exciting of the discoveries were a combination of mineralogical, computational, and geothermometric data sets that revealed a chaotic-like behavior of the processes is critical in the formation of reservoirs near cooling plutons. This discovery provides a fundamental basis for improving resource assessment and exploration methods for geothermal energy associated with very young magmas. Some of the main results are documented in scientific publications, and DOE progress reports. An additional publication is in preparation on the overall significance of fracture propagation and microseismic activity around young magmas, most of which is presented in the following documents.

The highlights of the results are summarized below:

Resource ranking with respect to its natural renewability is proposed through a classification scheme that affords for the stage of reservoir development with respect to the energy state of the heat source. This tripartite scheme is based on the quality of the thermal field with respect to its ability to reorganized original pore space into percolating clusters of fractures. It emphasizes the appearance of a prospective region as seen by heat flow, microseismic and mineralogical indicators.

Process behavior preserved as patterns within hydrothermal minerals appears have chaotic like qualities. Recognition of this behavior style is important for making future resource assessments and development of remote detection methods. Clearly the instance of resource formation during which the most renewable resources are available is while the chaotic-like behavior prevails.

The resonance of these processes predetermines the detection, quality, and recoverability of geothermal energy in cooling pluton environments.

Methods useful in the interpretation of standard geothermometric data sets collected in prospective drilling ventures were discovered and applied to a partial data set from the Geysers. These methods correlate fluid inclusion observations directly to magma time scales. Although preliminary in scope they hold potential for application to a few decades of geothermometric data sets on hundreds of hydrothermal systems. The results would be a close correlation of these data with time of their formation in the magma-hydrothermal cycle.

Thermal history of the Geysers Felsite has been reconstructed and a spectrum of lithocap permeabilities explored. The numerical models assist in correlating radiometric ages, argon-argon events, regional structural events and near-critical behavior with the magma's time scale. There remains a significant discrepancy between our analyses and the ages proposed by radiometric methods.

Pore space reorganization into ellipsoidal form fractures is a key process in the early stages of thermal dissipation from magmas into their lithocaps. Percolation of hydrothermal fluids through clusters of these fractures provides the principal mode of thermal energy transport away from the magmas and in the concomitant development of the pre-reservoir state conditions. A theoretical model of this process has been developed on the basis of the geometry of fractures preserved in drill core from the Geysers and on the basis of numerical approximations to fracture propagation by fluid expansion.

The results clearly display the role of the hydrothermal fluid in propagating and preserving percolating fracture networks in the lithocap. Similar to other environments and processes the most active time of this process is during prograde thermal heating of the lithocap and at state conditions near the critical region of the fluid.

Theoretical microseismic activity was computed from the fracture forming processes. These results suggest that energy release caused by prograde thermal fields above young magmas may produce a signal diagnostic of this process. The distribution function of microseismic magnitude and numbers of events is linear in log-log space similar to regional seismic activity. But the functions display a multi-fractal dimensionality that correlates directly with state conditions of the fracture failure. These observations are exciting but would require additional analysis for development into a viable exploration tool.

Progress germane to the basic science of magma-hydrothermal activity was significant as well. In the course of analyzing the geologic history of the Geysers Felsite, further evidence for the intricate coupling amongst the processes that dissipate energies from cooling plutons was uncovered. This was in part possible because of the relatively shallow emplacement depths of the Felsite. State conditions along its contact with its lithocap rocks persisted in the near field of the critical point of the hydrothermal fluid. This condition is significant because the extreme non-linear quality of the fluid properties exert a dominant control over the evolutionary behavior of the hydrothermal processes.

The persistence of dynamical behavior of hydrothermal-magma processes often evolving to chaotic style as caused by H_2O -rich fluids in shallow crustal environments has been further extended into geothermal resources as a consequence of this project.

A general feature of present and past research on geothermal resources has been the exclusion of systematic accumulation of data sets that would permit quantitative analysis of the historic extent of advective metasomatic processes in and around reservoirs. Oxygen isotopic data has been acquired and has been useful for evaluating the magnitude and timing of chemical changes around the plutons. But the opportunity to use advective metasomatic changes to better quantify the consequences of hydrothermal activity on the quality of reservoirs is woefully lacking in the geothermal program. The data called for is justified on theoretical and practical basis by Villas and Norton, 1977, and Norton, 1988. Thermodynamic stability of ions, minerals and aqueous complexes are well known. If one obtains estimated, or quantitative, values of the abundance of major chemical components, calculates these values into mineralogical abundances, as per Villas and Norton, 1977, time variations in permeability can be derived. This focus is highly recommended for future work on reservoir environments.

Published Results from this project

Norton, D., 2000. Systematics of Magma-Hydrothermal Processes: Geothermal Resources Near Cooling Plutons. Geothermal Res. Council Annual Meeting.

Hulen, J.B., Norton, D.L., Moore, J.N., Beall, J.J., and Walters, M.A., 2001, Initial insights into origin, configuration, and thermal-chemical evolution of the Aitkin Steam Reservoir, Northwest Geysers steam field, Ca. GRC-Annual meeting.

Norton, D. and, Hulen, J., Jeffery, B. 2001. Preliminary numerical analysis of the magma-hydrothermal history of the Geysers Geothermal System, California. Geothermics special issue. 30, 211-234.

Norton, D.L. and Dutrow, B.L., 2001. Complex Behavior of Magma-Hydrothermal Processes: Role of Supercritical Fluid. *Geochimica et Cosmochimica Acta*, v.65, pp. 4009-4017.

At least one additional manuscript can be derived from this work and perhaps two. The topic relates to the character and style of fracture propagation and geometric quality of the associated micro-seismic signals.

Introduction

Dynamical History of Reservoirs

All magma-hydrothermal processes display similar behavior over their active thermal life-times. We focus here on events associated with a single event in which magma is emplaced into a relatively shallow crustal level then dissipates energy into its host rocks. In concert with the assumption that magma intruded at these levels is at a rate much greater than the loss of energy to surroundings, a time fiducial reference for initiation of energy perturbation imposed on the lithocap of a magma is chosen. This time = 0 condition is taken when the bulk of the magma body reaches its static level. A time when the magma-host contact no longer advances; intrusion of dikes and sill size bodies continue but in effect the intrusive phase has ceased. As a general rule the temperature at this contact is about the average of the magma solidus and local ambient temperatures at time=0. We examine the processes attendant to the crystallization and cooling of the magma relative to this time reference.

Temperature-Time spectra

Evolution of the state of a rock located in the lithocap of a magma body is simple, Figure 30; initial values of thermal energy, temperature used as a proxy, increase rapidly, stage I, then slow to maximum values, stage II, and decay asymptotically, stage III, toward steady ambient values.

Figure 1. Temperature-time spectra for region in lithocap of a magma. Displays general stages of evolution in history of magma cooling and imposition of energy pulse onto rock at ≈ 1 km above magma-host contact. Values appropriate for Geysers Felsite like body. Inset shows lapse time prior to arrival of thermal pulse, early stage I, and its functional form

These resource stages derive from the feedback relationships implicit in the dissipation of mechanical, thermal and chemical energies from the source and specifically affords for the fundamental underlying shifts in sign change of the time derivatives of conserved quantities. To wit:

$$\text{Stage I : } \frac{dT}{dt} = (+) ; \quad \text{Stage II : } \frac{dT}{dt} \approx (0) ; \quad \text{Stage III : } \frac{dT}{dt} = (-) \quad (1)$$

The folded function appearance of the temperature-time is evident in numerical representations of processes and may be documented for any particular active process. The form of these series is derived from numerical approximations to natural situations. Coupling of

the hydrothermal processes and control by fluid properties produce distinct patterns from the idealized function shown above primarily in the addition of high-frequency oscillations and increase in the duration of Stage II, Figure 2.

Figure 2. Temperature-time spectra for region in lithocap of a magma. Similar to Figure 1, except lower confining pressure values in this calculation cause hydrothermal processes to operate close to the critical point of the H_2O -rich fluid.

This functional form and classification pertains to the entire lithocap region of magma bodies in general. The arrival of the initial perturbation and consequent stages is shifted in time proportional to distance above the magma and to the extent of hydrothermal activity, Figure 3.

Figure 3. Temperature-time spectra for regions in lithocap of a magma, distance from magma-lithocap contact ranges from 0.15 to 2.25 km in steps of 0.3 km. Location of temperature maximum at each location marked with solid triangle.

The family of temperature-time functions depicts local progression of state conditions through the modes of change labeled by Stage. Near the magma-lithocap contact, 0.15 km above pluton, Stage I arrives in but a few thousand years after intrusion. The subsequent sweep of this stage upward through the lithocap is depicted by the Δ at each respective T_{max} , aka Stage II. The rate of this sweep of T_{max} is $\approx 1\text{cm/yr}$ at 0.15 km above the pluton, slows to about 0.5cm/yr at 0.45 km then accelerates to 1.1cm/yr at 2.25 km above the pluton. This rate change of T_{max} is synchronous with free convection velocity increases as the fluid density perturbation buoys toward the surface, cf Norton and Knight, 1977.

The scale of local temperature variation with time is small, a few degrees per thousand years, and not measurable. But the Stages of development should also be apparent in the form of temperature-depth values measured at any instant in time, Figure 4. Because the rate of temperature change is functionally related to its spatial gradients, measurements of temperature gradients and correction for local variations in thermal conductivity can reveal

the Stage.

Figure 4. Temperature-depth, $T = f(z)$, for regions in lithocap of a magma, at stages relative to magma-time = 0. The linear plot, Δ , depicts ambient temperature gradient prior to magma emplacement, Stage I gradients shown for condition at time ≈ 0 of magma emplacement, upper linear section with sharp temperature increase at about 3km depth to magma values at 4 km, and slightly later in time typical of strong prograde thermal field, — — —. Stage III conditions prevail for long interval of thermal decay well after magma has crystallized and cooled to near ambient conditions at depth.

Positive curvature of the thermal field, increasing gradients in temperature with depth, is consistent with Stage I conditions of increasing temperature with time, $dT/dt \propto \nabla^2 T = (+)$. Note however that at depth in the pluton the contrary condition exists and perseveres for the entire duration of the perturbation, but in the lithocap a distinct sequence of stages are detectable. In practice exploration methods tend to focus us onto rocks at Stage III $dT/dt \propto \nabla^2 T = (-)$ conditions simply because of our fixation on abnormally high thermal near-surface gradients. These persistently anomalous thermal gradient values near surface can and have misled many exploration efforts, cf Marysvale, Montana early '70's drilling failure is one good example of this.

Processes Attendant to Magmatic Events

The state of hydrothermal activity relative to magma intrusion can be thought of as a complex interaction of processes that dissipate energy from the crystallizing and cooling magma. Because the quality of geothermal resources associated with magmatic thermal anomalies strongly depends on the instant in magma-time that the anomaly is prospected and on the location within the thermal aureole where the prospecting is carried out, resource assessment must afford for these complexities.

Although this evolutionary feature has long been understood in scientific studies of hydrothermal activity, the geothermal literature and practice have yet to focus on time-space variance. A recent exception is perhaps the exploration in the Adlin region of the Geysers where a hypothetical "young intrusion" has been promoted as the cause of moderately higher temperatures than heretofore encountered in the field. As a result of work performed by this project we demonstrated, a young pluton was possible, but time-space numerical models revealed that all conditions were most compatible with a *time* = 0 coincident with the main Felsite magma.

A schematic guide to the systematics of hydrothermal activity in this context was derived

early in our studies of these processes, Norton, 1979, and evolved with a couple of decades of study into its present form, Norton and Dutrow, 2001. We view hydrothermal activity as the manifestation of interactions amongst transport processes that redistribute mechanical, thermal and chemical energies from the magma body (or from any other thermal perturbation in the crust) into the host rocks. In this construct the products of these processes represent the temporal evolution of the system and when mapped in this context yield a time-serial fossil record of the hydrothermal events.

One major scientific advance made during the current work was a correlation of fractal-like geometric patterns in tourmaline with the chaotic-like behavior of hydrothermal activity operating at near-critical states of the fluid, Norton and Dutrow, 2001. This discovery should motivate detailed studies of both the Geysers Resource and other resources in search of similar correlations and better documentation of the hint we found.

Throughout the following summary of temporal evolution of fracture, thermal, chemical and mineral alteration features, supporting information is used from the well known fact (see references cited above) that simple feedback between processes controls their evolutionary behavior and that a prominent link is through the equation of state of the hydrothermal fluid. The directed graph of these processes provides a schematic of how a potential reservoir rock evolves in the transient thermal field of a magma, Figure 5.

Figure 5. Directed graph of hydrothermal processes following magma intrusion into the upper earth's crust. The consequent energy dissipation and crystallization processes specific to the magma are coupled respectively through the energy levels. Levels (round cornered boxes) of Thermal, dH; Mechanical, dF; and Chemical, dA; energies and their rate mechanisms of change (arrows) advective, flow, heating, strain, surface area, and reaction rates compose a feedback relationship that displays dependencies of one process on another, Norton , 1979.

Hydrothermal fracture growth at Geysers

Introduction

Connectivity of pore space over large rock domains is essential for productive fluid flows during extraction of geothermal energy. The formidable problem of thermal energy recovery from regions rich in energy but of low pore connectivity remains unsolved. This communication reviews current concepts of pore space and alteration of pore space in the thermal aureoles of crystallizing and cooling magmas and their implications to detection and extraction of geothermal energy. Fluid filled pore space in all rocks is graded in form and distribution, Clemo and Smith, 1997. This hierarchy of pore distribution evolves with time as rocks are subjected to local and regional changes in state conditions. However during this evolution pore distribution maintains a remarkable similarity in the style of its partitioning amongst dominantly discontinuous and isolated pores and but a small fraction of interconnectivity in fractures. Indeed the vast literature on rock porosity suggests a hierarchy of pore space at the hand-sample scale in a rock. Norton and Knapp, 1977, found two major domains of pore occurrence: 1) $\approx 70\%$ of pore space in hand size samples is isolated in sizes equivalent to or smaller than grain diameters. In this domain a fluid filled pore simply assumes the role of a rock forming phase, and 2) the remaining $\approx 30\%$ of pore space in hand size samples extends over lengths much greater than the grain diameters as microfractures but still relatively isolated from the view of fluid flow. In samples the size of outcrops a third domain occurs along fracture/vein trends that is volumetrically insignificant. However, this pore space critical importance to fluid transport processes by flow.

Similar distributions of pores are found in virtually all rocks, suggesting that pore space modification by the imposition of hydrothermal events is essentially one of redistribution of pores into orientations and forms consistent with evolving changes in the stress field and strong modification by mineralogic alteration processes. Although brecciation, faulting, rock recrystallization and alteration often lead to enormous local but temporary increases in pore space, total porosity variations tend to be relatively small in response to magma-hydrothermal processes attendant to formation of geothermal reservoirs. The crucial feature of pore space from the view of hydrothermal alteration and geothermal energy recovery is continuity. Increase in the ability of fluids to percolate long distances along pore pathways through propagation of the pore as an elongate slit-like feature.

One of the most important qualities of fluid flow regimes along fracture controlled percolation paths is the relatively small quantity of pore volume required to render a rock quite permeable. Fractures a meter apart with but a few tens of microns of effective percolation space can render a rock permeable in the millidarcy range, but accounts for only 0.03% pore space in rocks that have at least 1%, and often in the range 2 – 5% total pore space. In this discussion we review the geometry and hierarchical nature of pores and how a portion of this pore space becomes transformed into elongate fracture-form pores. This discussion is illustrated by a collection of pore and fracture maps made of rocks in a broad range of magma-hydrothermal environments.

Pore Hierarchy

Form and distribution of fluid-filled pore space in rock is similar from micron to the kilometer scale of observation.

Figure 6a. Initial distribution of pores in granitic rock, black regions among grains, and within grains, in a rock at an instance when stress field changes from homogeneous to heterogeneous, with arrow depicting the maximum principle stress direction, the minimum stress lies in the plane of the paper and is oriented at right angles to arrows, and the intermediate stress is oriented normal to the plane of paper and to minimum and maximum stresses. Scale bar is 1mm.

Figure 6b. Same rock as in, 6a, schematic of pore redistribution, overall constant porosity, that develops during pro-grade thermal field event, eg. $dT/dt == +$. Thermal expansion of fluid within an anisotropic stress field redistributes part of fluid-filled pore space into slit like fractures oriented in accord with regional stress field. Growth of these pores is primarily in length in response to changing confining stress or internal fluid pressure levels. The process is controlled by the filling fluids expansivity.

Figure 7. The growth process preferentially extends long pores relative to short ones. The consequent harvesting of fluid from smaller pores by longer pores enhances the bimodal distribution of pores. A major portion of the original pore space remains isolated from the fracture form. Overall there is little change in this hierarchical pattern, but the consequence of generating even small connectivities is enormous. The inevitable mineral alteration process initiated by these state changes may enhance or mitigate the connectivities.

Continued increases in thermal energy levels extend these small fractures and connect with similar microfractures developing into neighboring regions. In effect domains of connected fractures swarm along with the progressive thermal front.

Figure 8. Alteration of pore space accumulates a small fraction of the previously isolated pores into larger slit-like forms and connects with nearest neighbors. This alteration extends the accumulated pores into fracture forms over longer distances thereby subdividing the rock and enabling greater extents of fluid percolation. Percolation through interconnected domains augments heat transfer rates and consequent convection further increases dT/dt , fluid expands and fracture lengths increase within the domains.

Figure 9. Examination of the connectivity of pore space along apparently continuous fractures inevitably reveals partial or complete plugging of the flow path by mineral deposition within the pore. This plan view of the lower section of the fracture in the previous figure shows less than 0.3 open space. This percolation index is well below the ≈ 0.7 value of the percolation threshold. The plugs in this fracture are pillars of quartz that prop the opposing walls of the fracture open, but render pore space connectivity into one of a tortuous tube lying in the plane of the fracture, Norton, unpub.

Figures 8-9 represent the final expanded state of the fluid and the connectivity may have been greater at in-situ conditions during the peak of progressive thermal conditions. A common feature shared by all fracture/vein topologies is that the most extensive metasomatic changes in rocks are noted within domains of abundant fracture because of consequent high permeability. Fracture domains are typically veined shut, eg. extremely low permeability today of ore grade mineral deposits formed by this process. Extremes of metasomatism closely follows episodes of fracture but seal the rock. Typical extents of sealing and residual pore distribution is displayed in Figure 10.

Figure 10. At one time a continuous relatively permeable fracture enabled extensive fluid flow followed by metasomatic plugging of the fracture aperture. The quartz vein A-A' in a rock slab was open to flow during addition of silica. Now this vein displays pore continuities over less than a few millimeters in all directions, Norton, unpub.

At other scales fracture domains similar to Figure 11 below are universal features of each and every pluton-host rock situation.

Figure 11. Self-similar pore forms extend over large domains. Fracture controlled pores with continuity over several kilometers fragment this granitic pluton. Heterogeneous crystallization processes locally altered the orientation of the fractures, both within the pluton and within the host rocks near its contact with the host, cf insets.

Synchronous development of fractures within the plutonic mass and its host rock are characteristic of magma-hydrothermal processes. The chronology of fracture events can be reconstructed from mapping methods developed explicitly for these environments. The results show an initial control of local energy changes required for fracture failure and also transient modification of the regional stress orientations by the presence of liquid like material that does not transmit heterogeneous stresses but deflects them around magma bearing regions.

Magma bodies similar in size to the one that formed the Geysers Felsite require about 100,000 years to crystallize. This time interval is on the order of the time during which prograde thermal conditions persist in the magmas lithocap. Consequently, fracture domains tend to record the orientation of deflected stresses. The record of this change is manifest by concomitant metasomatic alteration events in virtually all pluton environments studied to date. Because fracture orientations must align with principle stresses at the time of their formation, a diversity of orientations that literally "box-the-compass", Figure 12, are considered to be a consequence of local deflection of regional stress orientations imposed by

the presence of magma.

Figure 12. Maps of fracture domains with similar orientations in effect reconstruct the time-serial history of stress orientations. This sample shows changing orientations from the initial K-spar-quartz, through quartz-K-spar-biotite-chalcopyrite and finally to quartz-sericite-pyrite vein events. Each event can be tied to a particular thermal state with geothermometric data on the fluid inclusions. This information and thermal models determine the relative age of fracture events.

Pores at the Geysers

A similar evolution of pore space is evident at the Geysers. Although the metasomatic removal of calcite played an important role and may have contributed significant quantities of pore space, the following sequence of events were observed in drill core.

Figure 13. Pore space within the rocks that host the Geysers Felsite prior to emplacement of the magma were distributed much like in figure 1, and among deformed calcite-quartz veins, ct. These irregular bleds and distorted veins are relicts from a much older metamorphic event. They are generally absent in rocks affected by the Geysers Felsite suggesting that their metasomatic alteration occurred in concert with dissipation of thermal energy from the pluton.

Figure 14. Near vertical single fractures are common in drill core of the Felsite host rocks. Quartz, q, and pore, v, filled fractures appear to have been derived from the thermal-mechanical-chemical alteration of rocks similar to those in the previous figure. Because of their low percolation values they are not likely contributors to steam production.

Figure 15. A variant of the fracture/vein structure in the previous figure. This large clot of quartz and pores formed at the intersection of several fractures is located at or near an important steam entry.

Records of multiple fracture events, such as shown in Figure 12, were not noted in the hosts rocks of the Felsite. This suggests that either the sampling has missed these or that the Felsite did not strongly alter the stress orientations when in the magmatic state. However, within the Felsite carapace concentrated tourmaline veining displays superimposed events with variations in orientation.

In summary, the distribution of fluid filled pores in rock is similar to the distribution

of mineral phases except for the pores deformed into fracture elements. We should be comfortable assuming a pore distribution in the host rock of the Geysers Felsite similar to the idealized sketches of Figures 6a, 6b and 7. And the long held view that pores deformed within and anisotropic stress field can be approximated by ellipsoidal forms.

Fracture Growth and Metasomatism around the Geysers Felsite

A comprehensive view of fracture growth and accompanying metasomatic effects has been developed by Jeff Hulen, unpub, from extensive observations of core and cuttings from the Geysers. He presents a three stage sequence similar to those sketched in the previous section, but complete with detail of the major mineral alteration events that are intimately associated with fracture development.

Figure 16. Hulen's time serial reconstruction of pore space in the Geysers reservoir represents three stages: A) Pre-Felsite metamorphosed Franciscan lithic metagraywacke (mgw), and interbedded argillite (arg) disrupted by quartz (q_1)-calcite (cal) veins cut by a shear fracture (f_1), B) Hydrothermal fractures (hf) and breccias (hbx), considered to represent the early stages of alteration by the Felsite. This calc-silicate dominated alteration stage in effect removed calcite and formed epidote (ep) and actinolite (ac) in the veins and their selvages (s), then C) Metasomatic filling of fractures and interclastic breccia voids with epidote, quartz (q_2), potassium feldspar (kf), prehnite, sericite and a variety of base metal sulfides.

The angular, residual vugs (v) remaining in stage C Hulen suggests represent the fluid storage in the reservoir that is accessed by the later stages of tectonic and hydrothermal fracture (f_2). The combined isolated and percolation pores are presumably the multiple pore space continua required to describe the Geysers Reservoir as a heat like pipe system, Pruess and Narasimhan, 1982.

Assuming that stage A represents conditions at $time = 0$ yr of emplacement of the felsic magma, then one is prompted to estimate the relative times at which B and C represent in the approximate 750kyr thermal history of the Felsite. Note worthy in this sequence of events is a less complex orientation of fractures relative to those typical of pluton environments. This topology may reflect a more dominant role played by the strike-slip regional stress regime during the prograde thermal event. Additional constraints on this timing can be derived from the thermal spectra reported for fluid inclusions.

Geothermometry of Veined Fractures

Replicate fluid inclusion measurements at a fixed location within a vein are interpreted in this study to represent a statistical sampling of fluid composition and temperature variations at that location over an interval of time while the host fracture was open to fluid percolation. The data sets are regarded as a record of the evolution of hydrothermal conditions during fracture extension processes discussed above. Fluid inclusion data are correlated through their temperatures of formation with temperatures estimated by numerical approximations to the thermal history of the magma-hydrothermal processes at locations in the numerical model analogous to those for fluid inclusions in the actual rock.

There are two data sets that are relevant from the Geysers Geothermal Deposit. Fluid inclusions in Geysers Felsite's lithocap just above the contact of the felsite with the hornfels have been studied by Moore and Gunderson (1995). They use the fluid inclusion data as an indicator of the conditions that several paragenetic events occurred. Ancillary information from that study affords for analysis of conditions with time at fixed distance from the felsite-lithocap contact. The geothermometric data are cast in the temporal evolution of the cooling felsite as predicted by, Norton and Hulen (2001) and Brikowski (2001).

The two sets of fluid inclusions in quartz analyzed by Moore and Gunderson (1995) are from widely separated points, $\approx 5 \text{ km}$, in the Geysers Geothermal reservoir, but at comparable locations within the felsite-lithocap contact, $\approx 250 \text{ m}$ above the contact. The first set is from 282 meter above the felsite, Figure 17.

Fluid inclusions in Felsite Lithocap

Figure 17. Homogenization temperatures, $Th, ^\circ C$ for inclusions in quartz from 282 m above the felsite, in lithocap. Samples from L'Esperance-2 well in north Geysers region. Quartz, epidote, actinolite, and chlorite assemblage at all temperatures are observed. From Moore and Gunderson Table 1, the pressure at this depth is estimated to have been 900 bar, equivalent to a paleo-depth of 3400 m.

Although this set of inclusions was estimated to have formed at 900 bar, they are plotted at pressure of about 300 bar in the *Pressure – Enthalpy* projection of the H_2O -system, Figure 18. The shift to higher pressures will cause a non-linear increase in the temperatures of formation of the inclusions. However, for an initial proof of concept we show these at their homogenization temperatures a condition that has little affect on the following conclusions.

These inclusions occur in the lithocap where the thermal regime is double valued. They can be interpreted as having formed within a progressive thermal field, time increasing from left to right, or in a regressive thermal field time increasing from right to left. Because of their location within a veined fracture and general bi-modal style of fracture events in the system, map determination of the fracture mode could determine whether they deposited during increasing or decreasing thermal state. As a provisional analysis the sense of the

time variation is left open and the inclusion data displayed with numerical conditions for a packet of lithocap rock at a comparable location to the L'Esperance samples, Figure 17.

Figure 18. Enthalpy-Pressure projection of the equation of state for the H_2O -system and inset of fluid inclusion temperature-histogram data from, Figure 17. Horizontal axis on fluid inclusion inset is scaled to temperature contours in enthalpy-pressure plot of EOS. Histogram from, Figure 17, is located at 300 bar and temperature values corrected for depth. Isotherms, black lines, labeled on the frame top axis in $^{\circ}C$, isopleths of density, gray lines, in g/cm^3 , and the liquid-vapor dome, H_2O L - V. The trajectory of enthalpy-pressure variations for a location ≈ 250 m above the contact of the felsite with its lithocap in the numerical model displayed by subhorizontal fat gray line punctuated with black dots. The dots are spaced at 0.05 my increments, starting with the dot at lowest enthalpy at $time = 0$ my in numerical model.

Formation of the inclusions during the temperature increase during the initial 50,000 yr, requires a much larger fluid flux than during the waning thermal phase of 150,000 yr. Fluid flow rates are about 10 times faster during the first stage than during the waning stages.

Similar arguments exist for the set of data on samples from OF27A2 well. These samples occur in a similar location to the L'Esperance samples, close to the felsite, and Their homogenization temperatures have a similar distribution, Figure 17.

Figure 19. Homogenization temperatures for inclusions in quartz from 236 meter above the felsite, in lithocap. Quartz, biotite, actinolite, and sulfide are noted, at 262°C tourmaline is observed, and appears sporadically in assemblage at all higher temperatures. From Moore and Gunderson (1995, Table 1), the pressure at this depth is estimated to have been 835 bar, equivalent to a paleo-depth of 3135 m.

Inclusions observed in OF27A2 (≈ 250 m) above the contact of the felsite with its lithocap are correlated with the same theoretical thermal history computed for the felsite, Figure 20.

Figure 20. Enthalpy-Pressure projection of the equation of state for the H_2O -system, and inset of fluid inclusion temperature-histogram data from, Figure 9. Horizontal axis on fluid inclusion inset is scaled to temperature contours in enthalpy-pressure plot of EOS. Histogram from, Figure 19, is located at 300 bar and temperature values corrected for depth. Isotherms, black lines, labeled on the frame top axis in $^{\circ}C$, isopleths of density, gray lines, in g/cm^3 , and the liquid-vapor dome, H_2O L-V. The trajectory of enthalpy-pressure variations for a location ≈ 250 m above the contact of the felsite with its lithocap in the numerical model displayed by subhorizontal fat gray line punctuated with black dots. The dots are spaced at 0.05 my increments, starting with the dot at lowest enthalpy at $time = 0$ my in numerical model.

The correlation of observed values with numerical values of temperature variation with time suggests that the fluid inclusions in quartz were either formed during the initial heating episode 0 – 50 thousand years following intrusion of the felsite, or they formed around 450 – 600 thousand years. Mechanical conditions necessary to maintain fluid percolation and quartz deposition in the fractures are optimum during the initial heating episode.

History of Fluid Inclusions Formation within Felsite

Their data on samples of drillcores from well bore DV2 at a depth location -111 m below the contact of the felsite with its lithocap were interpreted as representative of the thermal history of this portion of the felsite. Temperature of homogenization data from this data set have been binned into 10°C increments and mid-points of the bins joined, Figure 21.

Figure 21. Homogenization temperatures, $Th, ^\circ\text{C}$ for inclusions in vein quartz from 111 meter within the felsite. At temperature values less than $\approx 276^\circ\text{C}$ only quartz is recorded as the vein mineral, for greater temperatures, both quartz and tourmaline coexist. Salinities range from nil to $\approx 6\text{ wt}\%$ NaCl -equivalents.

Computed pressure-enthalpy-temperature-time values at a location comparable to the DV2 samples from the numerical analog of the felsite thermal history, Norton and Hulen, 2001, and Norton, 2000a, were correlated with the data set in Figure 21. This plot provides a direct correlation between the estimated fluid inclusion temperatures and the computed values. Data From Moore and Gunderson Table 1, the initial pressure at the depth of occurrence of the fluid inclusions was estimated to have been 300 bar, equivalent to a paleo-depth of 1153 m.

Figure 22. Enthalpy-Pressure projection of the equation of state for the H_2O -system and inset of fluid inclusion temperature-histogram data from Figure 1. Horizontal axis on fluid inclusion inset is scaled to temperature contours in enthalpy-pressure plot of EOS. Histogram from Figure 1 is located at 300 bar and temperature values corrected for depth.

Isotherms, black lines, labeled on the frame top axis in $^{\circ}\text{C}$, isopleths of density, gray lines, in g/cm^3 , and the liquid-vapor dome, $\text{H}_2\text{O } L - V$, and the solidus for a wet, $\approx 3 \text{ wt}/\% \text{-H}_2\text{O}$ magma, label *Magma Solidus*. The trajectory of enthalpy-pressure variations for a location $\approx -125 \text{ m}$ below the contact of the felsite with its lithocap in the numerical model displayed by subhorizontal fat gray line punctuated with black dots. Time dots are spaced at $50,000 \text{ yr}$ increments, starting with the dot at highest enthalpy at $time = 0 \text{ my}$ in numerical model. The encircled time dot is at $350,000 \text{ yr}$ after emplacement of the felsite magma.

Conditions predicted for the felsite cooling path, Figure 22, sub-horizontal gray line with tick marks every $50,000 \text{ yr}$, at a $\approx 260 \text{ bar}$. The correlation of observed values with predicted values of temperature variation with time shows that the fluid inclusions in quartz first formed a little after 350 thousand years, encircled time dot in Figure 22, but that most of the inclusions formed around $450 - 500$ thousand years. Recall that the x-axis on the inclusion distribution plot is also temperature.

This correlation shows that most of the inclusions formed after the local temperature in the felsite had cooled to state conditions on the low enthalpy side of the fluids critical isopleth, $0.32 \text{ g}/\text{cm}^3$. Although further interpretation of the cooling path of these quartz crystals and their inclusions is necessary in the context of the propagation of fluid filled voids in the near-critical region of the fluids equation of state, the first approximation to their thermal history is clearly depicted in Figure 22. This comparison also indicates that the early conditions from $t = 0$ to $t = 300,000 \text{ yr}$ within the felsite were not recorded by the fluid inclusion data set. Formation of these inclusions within the reservoir rocks during the waning stages of the felsite's thermal history was presumably long after many estimates of when the system became a 2-phase, vapor-dominated reservoir.

Approximate age of fluid inclusions

Correlation of the fluid inclusion temperatures of formation with the numerical analog system of the thermal history of the Geysers Felsite implies an age of formation of these inclusions relative to the emplacement age of the felsite magma.

The inclusions in the felsite appear to have formed during the waning portion of the cooling felsite stage from a period of about $375,000 \text{ yr}$ to $550,000 \text{ yr}$, Figure 23.

Figure 23. Distribution function of fluid inclusion formation temperatures from DV2 well bore scaled to time at which the region represented by these inclusions was at the corresponding temperatures.

In the lithocap there are two possible situations for the fluid inclusion ages. The first

is that they formed during the early stages following the magma emplacement when temperatures in the lithocap rocks were increasing. The second is that they formed during the waning portion of the cooling felsite stage.

Figure 24. Distribution function of fluid inclusion formation temperatures from OF27A2 scaled to time at which the region represented by these inclusions was at the corresponding temperatures and the temperatures were increasing with time.

Figure 25. Distribution function of fluid inclusion formation temperatures from OF27A2 scaled to time at which the region represented by these inclusions was at the corresponding temperatures and the temperatures were decreasing with time.

The critical test of these alternative time periods for formation of the fluid inclusions in the lithocap rocks is in the extent of SiO_2 metasomatism for the region represented by the inclusions. Because of the extreme differences in elapsed time between the two possible situations, advective rates of silica introduction could provide the deciding condition. Computation of the advective change in silica requires estimations of the gains in silica, and retrieval of the advective parameters from the numerical model.

A feature of quartz stability in hydrothermal systems is an extrema in solubility that lies just below the critical temperature for H_2O -rich fluid, Figure 26.

Figure 26. $\text{Log } M(\text{SiO}_2)_{\text{aq}}$ versus temperature at $P = 500$ and 1000 bars, for reaction $\text{quartz} \rightleftharpoons \text{SiO}_2(\text{aq})$. Spectra of homogenization temperatures, $T_h, ^\circ\text{C}$ for inclusions in vein quartz typical of lithocaps in fossil hydrothermal systems correlate closely with locus of extrema in $\text{Log } M(\text{SiO}_2)_{\text{aq}}$. Propensity of lithocap quartz to form as fluids advect aqueous silica from high temperatures, coincident with extrema, to low temperatures.

All geothermometric data sets collected from all prospective and producing geothermal reservoirs must be interpreted in the contexts presented above. In the Geysers reservoir situation a paucity of data exists and consequent serious gaps in the thermal history of this resource gem are apparent.

Quasi-Static Fracture Propagation

Fractures formed during the early stages of pluton cooling are long relative to their width (Figure 1a). This geometry can be approximated by pore space that is circular in plan view and elliptical in cross-section (Figure 1b). These pores have a major radius or length, l , and a minor radius, b , (Norton and Knapp, 1977). Consider the excess, effective, internal fluid pressure in this pore to be the pressure, P_f , e.g. fluid pressure less the mean of the confining stresses.

Figure 27. Perspective of idealized fluid pressured pore space in equilibrium with heterogeneous stress field. Plan and sections, upper left, define dimensions of fracture described by major axes, l_1, l_2 , minor axis, b , and volume V . Major axes values range from $l_1 \approx l_2$ for static equilibrium in a stress field with the intermediate stress, σ_2 , and maximum stress, σ_1 , equal, to $l_1 \gg l_2$ for $\sigma_1 \gg \sigma_2$.

The quasi-static growth of natural fractures is approximated by numerical approximations

to extension of an idealized elliptical analog form as a function of thermal and mechanical states. The analog pore is isolated from its nearest neighbors during failure events; no percolation occurs when fluid pressure increases. The pore is also subject only to Mode I, simple tensional opening, (Atkinson and Meredith, 1987). First a conservation of mass statement is derived, then a conservation of momentum statement is derived. The following derivation is motivated from earlier analysis of this phenomena by Rovetta and Norton (unpub 1994) and Knapp and Knight (1977). Mark Rovetta provided the key insight to making this formulation possible.

Conservation of Mass

The volume of an internally pressurized fracture in an elastic solid is a function of elastic parameters and the fracture radius, l , and the internal pressure, by Sneddon (1946):

$$V_{pore} = \frac{16}{3} \left(\frac{(1 - \sigma^2)}{E} \right) P_f l^3 \quad (2)$$

Where E is the Young's modulus of the rock matrix and σ its Poisson ratio. A Young's modulus for intact granitic rock of 3×10^5 bars, and Poisson ratio of 0.25, equation 2 becomes:

$$V_{pore}^{t_0} = P_f l^3 1.7 \times 10^{-5} \quad (3)$$

Where pressure is in bars and the fracture volume is in units of fracture length, l , cubed. A meter long, radius, fracture with an internal fluid pressure of 10 bars greater than the confining stress has a volume at reference time, t_0 , of , 167 cm^3 .

Differential changes in the volume, dV , of the fracture, with time, dt , follows from equation 2:

$$\frac{dV_{pore}}{dt} = 3\mathcal{Y}pl^2 \left(\frac{dl}{dt} \right) + \mathcal{Y}l^3 \left(\frac{dP_f}{dt} \right) \quad (4)$$

Where the stress values and geometric constants are collected into, $\mathcal{Y}$:

$$\mathcal{Y} = \left(\frac{16}{3} \right) \left(\frac{(1 - \sigma^2)}{E} \right) \quad (5)$$

The concomitant fluid volume change in the pore for conditions where the remote lithostatic stress is independent of temperature is determined by the equation of state of H_2O . For this derivation only a single phase, liquid, is considered filling the pore. When pressures are less than critical values, $P_{total} \leq 220$ bar the presence of liquid and vapor should be afforded for at this point. The volume change of the fluid depends on the isobaric thermal expansivity α_f , and isothermal compressibility β_f of the fluid.

$$\frac{dV_{fluid}}{dt} = \alpha_f V_o \left(\frac{dT}{dt} \right)_{P_f} - \beta_f V_o \left(\frac{dP_f}{dt} \right)_T \quad (6)$$

Negligible contributions associated with the thermal expansivity or compressibility of the rock have been omitted from this derivation to simplify the analysis. This assumption has

comparably negligible consequence on the final conclusions. Now substitute the starting reference pore volume from equation 2 into equation 6, for V_o :

$$\frac{dV_{fluid}}{dt} = \alpha_f l^3 P_f \mathcal{Y} \left(\frac{dT}{dt} \right)_{P_f} - \beta_f l^3 P_f \mathcal{Y} \left(\frac{dP_f}{dt} \right)_T \quad (7)$$

Equating the rate of fracture volume change, equation 4, with the rate of fluid volume change, equation 7, defines a conservation of mass for the pore containing a single phase:

$$\frac{dV_{fluid}}{dt} = \frac{dV_{pore}}{dt} \quad (8)$$

$$\alpha_f l^3 P_f \mathcal{Y} \left(\frac{dT}{dt} \right)_{P_f} - \beta_f l^3 P_f \mathcal{Y} \left(\frac{dP_f}{dt} \right)_T = 3\mathcal{Y} p l^2 \left(\frac{dl}{dt} \right) + \mathcal{Y} l^3 \left(\frac{dP_f}{dt} \right) \quad (9)$$

Collecting terms and solving for the length change with time:

$$\frac{dl}{dt} = -\frac{l}{3P_f} (1 + \beta P_f) \left(\frac{dP_f}{dt} \right)_T + \frac{\alpha l}{3} \left(\frac{dT}{dt} \right)_{P_f} \quad (10)$$

Conservation of Energy

The energy content, W , of a single fracture by Sneddon (1946):

$$W = \frac{\mathcal{Y}}{2} l^3 P_f^2 \quad (11)$$

An incremental increase in the fracture length, releases energy, J_s , per unit area.

$$J_s = \left(\frac{1}{2\pi l} \right) \frac{dW}{dl} \quad (12)$$

Carrying out a differentiation in equation 12 with l defines J_s for the idealized fracture geometry.

$$J_s = \frac{\mathcal{Y}}{4\pi} \left(3P_f^2 l + 2P_f l^2 \frac{dP_f}{dl} \right) \quad (13)$$

Using the differential relationship:

$$\left(\frac{dP_f}{dl} \right) = \left(\frac{dP_f}{dt} \right) \left(\frac{dt}{dl} \right) \quad (14)$$

Equation 13 becomes:

$$0 = \left[3lP_f^2 + J_s \frac{4\pi}{\mathcal{Y}} \right] \left(\frac{dl}{dt} \right) + 2l^2 P_f \left(\frac{dP_f}{dt} \right) \quad (15)$$

Combined equations

At mechanical equilibrium, the fracture's minor radius, or opening displacement at its center is b , Figure 1b.

$$b = \frac{3\mathcal{Y}}{4\pi} P_f l \quad (16)$$

In a half-space of rock with the fracture plane as its boundary, the component of momentum perpendicular to the fracture plane will be proportional to b the rate of change in b :

$$\frac{db}{dt} = \frac{3\mathcal{Y}}{4\pi} \left[P_f \left(\frac{dl}{dt} \right) + l \left(\frac{dP_f}{dt} \right) \right] \quad (17)$$

During quasi-static fracture propagation b increases slowly, and the change in momentum normal to the fracture plane remains negligibly small. Setting the time derivative of b to zero provides an additional constraint on the quasi-static condition of fracture growth.

$$\frac{db}{dt} = 0 = P_f \left(\frac{dl}{dt} \right) + l \left(\frac{dP_f}{dt} \right) \quad (18)$$

Solving for the pressure derivative with time and substituting into the conservation of mass relationships from equation 10, yields a statement for the change in fracture length with time for the condition that the energy level, J_s , is greater than the threshold level, J_s^* :

$$\frac{dl}{dt} = \frac{\alpha(T, P) l}{2 - \beta(T) P_f} \left(\frac{dT}{dt} \right) \quad J_s \geq J_s^* \quad (19)$$

Because fractures grow only when F equals or exceeds F^* , the effective fracture resistance of the rock, Atkinson and Meridith, 1987 . If F becomes smaller than F^* the propagation of the fracture stops:

$$dl/dt == 0 \quad J_s < J_s^* \quad (20)$$

The change in pressure associated with fracture growth derives from equating the change in length to change in fluid pressure in equation 18 and substituting in equation 19:

$$\frac{dP_f}{dt} = \frac{-\alpha(T) p}{2 - \beta(T) p} \left(\frac{dT}{dt} \right) \quad J_s \geq J_s^* \quad (21)$$

Now setting the length derivative with time in equation 10 to zero the equation for the case where the energy level is below the threshold is obtained:

$$\frac{dP_f}{dt} = \frac{\alpha(T) P_f}{1 + \beta(T) P_f} \left(\frac{dT}{dt} \right) \quad J_s < J_s^* \quad (22)$$

Together the equations 19, 20, 21, 22 provide a system of equations that describe the quasi-static extension of an isolated fracture during heating. Fracture growth in this model is irreversible, and there is never a reduction in the radius of the fractures on cooling or pressure decrease.

$$\frac{dl}{dt} = \begin{cases} \frac{\alpha(T)l}{2-\beta(T)P_f} \left(\frac{dT}{dt} \right) & J_s \geq J_s^* \\ 0 & J_s < J_s^* \end{cases} \quad (23)$$

$$\frac{dp}{dt} = \begin{cases} \frac{-\alpha(T)P_f}{2-\beta(T)P_f} \left(\frac{dT}{dt} \right) & J_s \geq J_s^* \\ \frac{\alpha(T)P_f}{1+\beta(T)P_f} \left(\frac{dT}{dt} \right) & J_s < J_s^* \end{cases} \quad (24)$$

$$J_s = \frac{3\mathcal{Y}}{4\pi} P_f^2 l \quad (25)$$

Example Integrations

Numerical experiments in which fracture growth response to increasing temperature at constant confining pressure is represented by equations 23,24, and 25 serve to depict the significance of this process for altering percolation state. One can think of the results as equivalent to the alteration of a packet of rock in The Geysers lithocap while it was subjected to progressive temperature increase.

Although pore space in the typical pre-felsite Franciscan graywacke is distributed among intergranular pores in the matrix and deformed calcite veins remnant from pre-felsite events, we here ignore the effects of concomitant calcite removal.

In the example computations, we examine the concept of pore space alteration caused by thermally induced fluid pressure increase within isolated pore space during a prograde thermal event. A temperature increase from ambient, $T = 100^\circ\text{C}$, to maximum values realized for contact thermal events, $T = 500^\circ\text{C}$, drives the fluid pressure equations. A confining pressure of, $P_{total} = 225 \text{ bar}$, Young's modulus of granite, $Y_{mod} = 3 \times 10^5 \text{ bar}$, Poissons ratio, $\nu = 0.25$, and critical level of fracture failure energy, $J_{critical} = 100 \text{ J/m}^2$ (Meridith and Atkinson, 1985) define the initial and transient conditions for a 2-D elliptical slit-like fracture. Fluid properties are represented by the equation of state for H_2O (Johnson and Norton, 1990).

Figure 28. Relative change in fracture length, L/L_o , as a function of temperature. Note rapid rate of fracture length change as state conditions approach those for the critical region of H_2O

Figure 29. Relative change in fracture volume, V/V_o , as a function of temperature.

Fracture length and volume increase slowly between $T = 100^\circ\text{C}$ and $T = 350^\circ\text{C}$, then increase rapidly. This nonlinear change is a consequence of the fluid properties, isothermal pressure compressibility and isobaric thermal expansion. The overall effect is a doubling of fracture length with thermal increase up to $\approx 400^\circ\text{C}$ and consequent volume increase of 8 times.

The fluid pressure not only alters the form of the pore space, but also its distribution. The orientation of the elliptical pore space form is determined by the relative magnitude of confining stresses. Initially the form and distribution of pores within the Franciscan rocks are related to the pre-felsite conditions. They appear irregular in shape and distribution. During the thermal event these pores are extended. Pore space is reoriented to extend into the maximum and intermediate confining stresses. During this process isolated pores are accumulated by extending fractures and the pre-felsite porosity self-organizes into a systematic fracture cluster. This process organizes pores into fractures, interconnects fractures and develops percolation networks.

The percolation of fluid under prograde thermal conditions in the near-field of the critical region is also conducive for the deposition of quartz along the flow channels. Examination of geothermometric data sets from quartz veined fractures from The Geysers Geothermal Deposit show reasonable correlation with the proposed fracture mechanism.

Theoretical Microseismic Activity in Pluton Lithocaps

During the prograde thermal events, Stages I and II, within the lithocap of a cooling pluton porosity is redistributed into slit-like fractures whose clusters percolate hydrothermal fluids. Observation and theory suggest that this redistribution of pores is accompanied by intermittent mechanical energy releases manifest by fracture extension. Because the extension process is controlled by a threshold value of rock resistance to failure, actual propagation of a fractures follows a thermo-mechanical stresses increase to this threshold value. The energy release during failure is determined by the integral of the mechanical work

$$dE = \int \gamma P dV \quad (26)$$

where $\gamma = 1.0 \text{ J/Pa} - \text{m}^3$ converts pdV work, to thermal energy.

Computations of the energy release for an approximate prograde thermal field such as could have been realized by a block of rock about 1 meter above the Geysers Felsite were completed. The governing conservation of mass and momentum equations used are for a slit-like fracture filled with water, Norton, 2001. The results are plotted in the context of

the Gutenberg-Richter seismic relationship among energy (or magnitude) of failure events and energy release greater than, dE , over an arbitrary time increment, $\dot{N}$:

$$\log \dot{N} = -b \log(dE) + k \quad (27)$$

A relationship that implies linearity of plots of number of failures and energy release in log-log space, with slope, b , and from the definition of geometric measures the dimension, D , is twice the slope, eg $D = 2b$. Values of $0.8 < D < 2.4$, Bak and Chen, 1995 seem typical for rock failures on the continental scale. Values for the regional USGS array at Geysers are $D \approx 2$ and local values for the Calpine array are $D \approx 4$, Smith, etal, GRC 2000. The latter values are attributed to reinjection/production activity in the reservoir, pc Bill Smith, 2002.

Energy release for a fracture failure process driven by pore fluid expansion within elliptical shaped pores subjected to a progressive thermal field yields an event-magnitude distribution characterized by a double slope, Figure 30.

Figure 30. Log Energy, PdV - Log Number of Events, for fracture extension caused by pore fluid expansion in an elliptical pore. The distribution has three distinct geometric values, $D = 0.4, 1.32$ and 3.82 . The largest value is associated with failures in the vicinity of the critical point of H_2O .

The spectra of failures is subject to many unsubstantiated assumptions. However, it does suggest that failure events associated with young magmas may display geometric properties distinct from those of regional events.

The geometric values derive from the effect of fluid properties on the energy release, Figure 31. An exponential increase in energy release is realized during the heating episode as near critical conditions are reached, $350 - 400^\circ C$ followed by a similar decrease. These values account for the $D = 1.32$ dimension of Log Events vs. Log Energy. The large $D = 3.8$ value is associated with the peak failures just prior to the decrease.

Figure 31. Temperature-Log Energy spectra for region in lithocap of a magma. Displays general stages of evolution in history of magma cooling and imposition of energy pulse onto rock at ≈ 0.1 km above magma-host contact. Values appropriate for Geysers Felsite like body.

Steam Quality

Steam quality produced by isenthalpic fracture paths whose starting state conditions lie in the supercritical region and end at the liquid-vapor surface depict typical situations expected from the fracture propagation model presented above, Figure 32.

Figure 32. Steam Quality-time spectra for typical paths of two lithocap rock and one pluton rock locations in Geysers region. Calculation assumes a fracture propagation event occurs at respective state condition of rock and final state lies on liquid-vapor surface. This assumption represents maximum steam quality values.

These preliminary computations are suggested paths for future analyses using methods that afford for the transition from supercritical fluid to 2-phase reservoir conditions. The important feature is that attainment of steam qualities of 0.5 or greater require fracture release at state conditions greater than the critical conditions of the fluid.

Some Prospecting Tools

Resource Renewability

Ranking of resources on the basis of their renewability qualities follows directly from the temperature-time spectra values for a particular resource. Using the thermal gradient criteria, and recorded fracture propagation by anomalous fluid pressure events several of the major geothermal resources have been referenced to a general thermal spectra.

Figure 33. Temperature-time spectra from Figure 2. Correlation of several resources with their appropriate stages as suggested by the general hydrothermal state of them. The Larderello/Travale and Cerro Prieto fields display features indicative of Stage I-II evolution; Salton Sea includes regions that span the Stage II and III; Geysers and Long Valley are clearly on the Stage III decay portion of the series.

In geologic context the truly renewable resources require development of regions in which the prograde thermal front has not propagated to the near surface levels. Focus on detecting this type of resource has several advantages with respect to quality of exploitable reservoir

regions and potential for stimulation of percolation paths.

Conclusions

The hypothesis that pore space in the Franciscan rocks that host the Geysers Geothermal Reservoir was redistributed by the advancing thermal energy front from the Geysers Felsite is supported by both mechanical arguments that describe fracture propagation under these conditions and geothermometric observations made on vein filling quartz. Although a portion the permeability increase results from hydrothermal alteration events that removed calcite or altered it to calc-silicate minerals, mechanical alteration of pore space into near-vertical slit-like percolation paths was an essential part of the process. These fractures are now nearly filled with hydrothermal alteration phases that required flow of fluid to transport the necessary chemical components to the site. The advective component transport in turn required an interconnected percolation path over ten's of meters from which to derive these components. Extension of the fractures is most effective during the prograde thermal event. Although quartz is more soluble at intermediate temperatures, the advective transport of aqueous, SiO_2 , into the percolation network readily oversaturates the fluid with respect to quartz, Norton, 1990.

The consequences of subjecting fluid-filled pore space to thermal conditions typical of those predicted for the early-stages of heating of the Franciscan graywacke host-rocks by the felsite, cf Norton and Hulen, (2001) and Brikowski (2001), are sequentially: 1) increase pore fluid pressure, 2) pore space altered into fracture forms, 3) fracture extension and alignment in accord with the principal confining stresses at the time, 4) preferential propagation of the longest fractures, 5) exponential fracture extension as temperature increases near the supercritical region of the fluid, and 6) consequent increases in fluid percolation as pore space is altered and concentrated into long-continuous fracture openings, and deposition of vein quartz.

These episodes occurred long before any manifestations of the embryonic reservoir appeared at the earth's surface.

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