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Construction of Equivalent Discontinuum Models for Fracture Hydrogeology

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1. INTRODUCTION

In many rocks where matrix permeability is low, fluid flows preferentially in the fractures. The prediction of flow through these rocks becomes a problem when the fractures which carry flow are not ubiquitously interconnected. In these cases, flow paths are controlled by the fracture geometry and may be erratic and highly localized. In contrast, highly fractured or porous materials often exhibit smoothly varying flow fields that are amenable to being treated as equivalent continua. The chaotic nature of flow in some fractured materials means that the well-developed and long-used techniques for modeling flow in porous media are often inapplicable to fractured rock.

This chapter describes an alternative conceptualization for fracture flow modeling called an "equivalent discontinuum" model. In an equivalent discontinuum model, flow in fractures is modeled as flow in equivalent conductors, which are regularized to lie on some form of lattice. This would be exactly the same as an equivalent continuum model except that some of the lattice elements are removed. The equivalent discontinuum model is simply a partially filled lattice that exhibits the same hydraulic behavior as that observed in situ. Where the fracture flow system is connected, the model is connected through lattice elements. Where the flow system is unconnected, the lattice elements are unconnected.

The concept of an equivalent discontinuum is extremely simple, but experience with the construction of such models is limited. An overview of a methodology for building such a model is presented here. Parts of this methodology are well developed, and other parts are on a steep incline of development. The most experience with this methodology is for the case where the hydrology is dominated by flow in fracture zones. However, one could easily modify the methodology for sites where the conductive fracture zones are not organized into zones. One should consider the work presented here as a "box of tools," which would not be applied the same way
at every site. The fundamental philosophy behind the approach is very general and is applicable to many types of geologic media, including some heterogeneous porous media.

In order to put the equivalent discontinuum model in perspective, a few previous approaches are described below followed by an outline of a methodology used to obtain an equivalent discontinuum model. Each part of the methodology is then elaborated on in subsequent sections.

1.1. The Discrete Fracture Model Approach

Recently, researchers have attempted to find models for fluid flow based on the geometry of the individual fractures (Hudson and La Pointe, 1980; Long et al., 1982; Robinson, 1984; Dershowitz, 1984; Billaux et al., 1989; among others). These models represent the fractures as conductive segments of lines or planes that are placed in space either deterministically or according to some stochastic process (as in Figure 1, for example). Fluid flow can then be modeled on the resulting network. Examples run with these models have shown that networks of fractures can only be considered as equivalent continua under very restricted conditions: when they are statistically homogeneous; when they are sufficiently well connected; and when a large enough sample is used (Long et al., 1982).

Application of these statistical models to real field sites requires that one measure the details of the actual fracture geometry and develop models that reproduce the statistics of the observed fracture network geometry. This involves determining a stochastic rule for locating fractures and determining their orientation, extent, and conductivity. Then a network can be defined and flow patterns calculated. This discrete fracture approach is a type of stochastic simulation. What makes the simulation difficult is that the fractures occur simultaneously on many scales and not all of the fractures are important for fluid flow. Some are filled, some are channelized, and some are not connected to the hydraulic network. Furthermore, the interior of the geologic medium is not visible to us and the characterization must be made solely through remote sensing and limited sampling from boreholes, outcrops, and underground excavations.
Figure 1. An example of a scheme for generating fracture networks as a Poisson distribution of line segments.
A stochastic simulation was made of the fracture network at the Fanay-Augères mine in France (Billaux et al., 1989). Among the data used were fracture trace maps and logs and single-hole packer tests. The analysis was based on assuming the fractures were disc-shaped and uniformly permeable in their plane. Data on trace lengths, orientation, and fracture frequency were used to create a model of fractures in a 100-m cube. The drift data indicated that fractures occurred in swarms, so the scheme for locating fractures in space consisted of generating locations for fracture clusters (parents) and then generating clusters of fractures (daughters) around the parents. The spatial statistics of the models were sampled in the same way that they were sampled by the field data. Then the model parameters were adjusted until the spatial statistics of the model and the field data were the same.

This effort demonstrated that the use of one- and two-dimensional data to infer three-dimensional geometry is extremely difficult. Many three-dimensional geometries can account for the same one- and two-dimensional data. Motivated by this observation, Mauldon and Long (1990) devised a linear programming algorithm for deriving any member of the infinite set of possible three-dimensional fracture statistics that will account for the same one- and two-dimensional data.

Lack of uniqueness is not by itself fatal, because many realizations could be generated to produce a range of expected predictions. However, no matter how the three-dimensional geometry was determined, there were far too many fractures to account for the lack of connectivity that was observed in cross-hole hydrologic and tracer test results. Clearly most of the fractures were not hydrologically active. The geometric approach failed to represent the behavior of a rock mass because the great majority of the fractures included in the model as conductors did not in fact conduct water.

If most of the fractures are unimportant to hydrology, then how can what is important be identified? Fanay-Augères offered one other key fact in this regard. Two drifts were mapped in this mine: one wet, one dry. For both drifts the fracture geometry analysis seemed to indicate highly connected fracture networks. However, a major fault ran through the block of rock sur-
rounding the wet portion of the drift. It seems that the hydrology of the site is controlled by major features, i.e., fracture zones. This observation is certainly not confined to Fanay-Augèrè. In an example from the Stripa Mine in Sweden, Olsson et al. (1988a) state that 94% of the hydraulic transmissivity is found in 4% of a particular block of rock. Similar evidence exists at the Underground Research Laboratory (URL) in Canada (Martin et al., 1990). Localized fracture and fault zone control is often observed at geothermal sites (e.g., Halfman et al., 1984; Bodvarsson et al., 1985; Laky et al., 1989; Beall and Box, 1989), and in tunneling and mining the sudden encounter of large fluid inflows is very common. In these cases the hydrology of the fractured rocks is controlled by a finite number of major conductors.

Creating a fracture hydrology model by counting and characterizing all the discrete fractures might be considered equivalent to modeling flow in the pores of a porous medium instead of applying continuum assumptions. Such detailed models require more data than can reasonably be obtained for practical applications. More importantly, the details are usually not important; only a finite number of features are important and should be modeled explicitly. Behavior on a relatively small scale can logically be averaged. In conclusion, the discrete fracture approach is designed to reproduce the discontinuous nature of the problem, but because the approach uses geometry to predict behavior, it has the drawback of losing the forest for the trees.

The focus of developing a fluid flow model should be to find the most important features first. In fractured rock, the big fracture zones often dominate the behavior. In this case, it is more useful to identify and characterize the major zones that are the primary conductors in the system than it is to collect data on detailed statistics of the geometry of the individual fractures. In many cases the zones are not continuous and the permeability structure within the zones is complex. So, once the zones have been defined, the next step is to find a way to model the complex behavior in the zones. In this approach, a model is built from the large scale down, rather than from the details up, and is focused on reproducing hydrologic behavior instead of detailed geometry.
1.2. Equivalent Media Studies

Once the large-scale features have been identified, how can the behavior of the smaller conductors within and around these features be efficiently modeled? Some encouragement for being able to find simpler, more appropriate models comes from recent work by Hestir and Long (1990) who have reproduced the behavior of Poisson networks of fractures using partially filled regular lattices. A class of two-dimensional networks was defined where the fractures, modeled as line segments of equal conductivity having arbitrary distributions of length and orientation, were placed on the plane with a Poisson process. The conductivity of these networks was shown to increase as a function of network connectivity in a manner that corresponds to the increase in conductivity of a lattice as the percentage of bonds increased.

The conductivity of regular lattices exhibits universal behavior that has been extensively studied through percolation and equivalent media theory (Kirkpatrick, 1973; Pike and Seeger, 1974; Kesten, 1982; Zallen, 1983; Robinson, 1984; Stauffer, 1985; Orbach, 1986; Kesten, 1987; among many others). The work of Hestir and Long implies that it is possible to find simple lattice networks that behave like the complex fracture systems observed in the field. This simplified network is therefore an "equivalent discontinuum." It is an equivalent model because it replaces the actual details of the physical system with an equivalent lattice. The equivalent model is a discontinuum because parts of the lattice may be disconnected from other parts. The equivalent discontinuum is to fractured media what the equivalent continuum is to porous media.

1.3. Construction of Equivalent Discontinuum Models

To make an equivalent discontinuum model, the features that control the first-order behavior are characterized first, second-order behavior second, etc. In this way, it is hoped, the need for saturating the field with a huge number of detailed measurements can be avoided. The dominant features are represented explicitly, and the details of the system are represented by a partially filled lattice chosen to reproduce the hydraulic behavior that was observed in the field. The steps are these:
(1) Identify the types of feature that control the hydrology.

(2) Locate these features in the field.

(3) Conceptualize the hydraulic system.

(4) Invert hydrologic test data to find an equivalent simplified lattice that has the same hydraulic behavior as that observed in the field.

This effort is inherently interdisciplinary in nature. Any one discipline, applied without respect to information gained from other types of investigations, will generate only part of the picture. If the model is based only on hydraulic data, it will be very non-unique: many flow systems could account for the same hydraulic behavior. To narrow down the possible explanations, it is necessary to have an accurate conceptual model for the flow system. Geologic tools provide information about what theoretically controls the flow, but geologic tools do not "see" into rock. Geophysics can see into the rock, but geophysical techniques do not directly measure hydrologic properties. The most power is gained by combining all these techniques into a unified approach.

There is no one example where all the components of this methodology have been applied. In an attempt to show the major constituents of the process, a variety of experiences from different field sites in fractured rock are cited. The format of the following sections is to discuss each component of the methodology in general and then provide an example to illustrate that particular component.

For an understanding of which features are important and why, applications of geology and geomechanics are discussed (2). For an understanding of what makes these features visible, a discussion of recent advances in geophysics and geomechanics that have allowed fracture zones to be imaged is given (3). The location of the important features is identified using a combination of geologic mapping and geophysical interpretation (4). Well tests are designed to test the hydrologic role of these features (5). Based on these, a hydrologic conceptual model (a "template") is described (6). The template should contain all the likely major conductors as the basis for fluid flow calculation. The template includes as much as possible of the qualitative hydrologic attributes of the zones identified through geology and geomechanics. Finally, conductors are arranged
in a manner that conditions the model to observed well test behavior. In other words, within the template, patterns of conductance are identified that can explain the observed hydraulic behavior. A new inverse technique called "simulated annealing" is used to arrange the conductances such that they explain observed distributions of head, observed fluxes, or observed tracer test results (7). Some alternatives to simulated annealing are discussed (8) which incorporate the fractal-like nature of fracture networks.

Models produced in this manner are non-unique. The fundamental reason for obtaining a non-unique model is that there is not enough information to create a unique model. In fact, every model of a hydrogeologic system is non-unique, even if it is constructed deterministically. The beauty of the methods described here is that they can be applied in a stochastic mode such that a series of models are produced, all of which equally incorporate all the observations made on the system. With this series of models, a series of predictions can be made. Given a new set of measurements, the error of these predictions can be calculated. This prediction error is the most useful measure of the worth of the model. Examples of the use of these models as predictive tools is discussed in the last section (9).
2. GEOLOGIC AND GEOMECHANICAL DESCRIPTION OF HYDROLOGICALLY IMPORTANT STRUCTURES

As stated above, the first step in building a fracture flow model is to decide which are the most important features of the fracture system. The important fracture features are those that conduct fluid; that is, the fractures that are 1) conductive and 2) connected to a network of other conductive fractures. It makes no sense to predicate an understanding of fracture hydrology on statistics describing the geometry of thousands of fractures that play little or no role in conducting fluid. It is a combination of the overall pattern of fracturing plus the conductivity of the fractures that controls fluid flow. These are factors that cannot necessarily be determined from statistics. They can be much more readily determined by looking at the network as a whole.

Sometimes it is possible to observe both the pattern of fracturing and also see which fractures are conducting fluid. For example, in the Stripa Mine in Sweden a 50-m long drift was excavated in the granitic rock mass and every fracture with a trace longer than a few centimeters was mapped. The rock appears to be ubiquitously fractured, with a concentration of fractures in the central 10 or 15 m of the drift. It is immediately obvious when walking down this drift that almost all the inflow comes from the central fracture zone, particularly from one fracture. Careful measurement has shown that essentially all the inflow comes from this zone and 80% from the single fracture (Black et al., 1991). Relative to this zone, the myriad of other fractures are unimportant. The important features at Stripa, i.e., the features with the conductive and connected fractures, are fracture zones. Furthermore, it can be surmised that the flow systems within the zones are likely to be only partly connected. If the fractures in the zones were all conductive and connected, the water would be expected to flow from many of the fractures in the zone. Even though the fracture zone is the most important feature at Stripa, the zone itself does not behave like a porous slab.

In another example from a fractured oil reservoir in sedimentary rock, fractures of two distinct orientations were logged from borehole measurements with one of the sets far more
significant than the other. Subsequent interference testing showed that the direction of maximum permeability was aligned with the less significant set of fractures and that this direction was parallel to the direction of maximum principle stress (Teufel and Farrell, 1992). Apparently, the importance of the fractures is controlled by the stress state. Fractures parallel to the maximum compressive stress will tend to be open, whereas those perpendicular to this direction tend to be closed. In this case again, observation of a large number of fractures does not necessarily indicate which are the important fractures.

Two components are involved in determining which fracture features are important for hydrology. The first is geologic observation of fracture style. The second is inference about which of these fractures are open and connected, i.e., which type of fractures conduct fluid. On first observation, fractures in rock may appear chaotic and completely disordered. However, as rock fractures have been studied in many locations and many rock types and geologic facies, it is clear that there are recurrent themes and patterns. Given the fracture pattern, inference about the relative importance of fractures can be made in a variety of ways. First, there is simple visual observation, e.g., the Stripa mine example discussed above, where it is easy to see that only fractures within a fracture zone are conducting a significant amount of fluid. Second, the geomechanical control on fracture permeability can be inferred as in the oil reservoir case above. Third, there can be strong geochemical control over fluid flow (Norton and Taylor, 1979; Bird and Norton, 1981; and Norton et al., 1984). Many fractures that might otherwise be connected are simply filled with minerals to the point that they do not conduct. One might be able to identify classes of fractures that are not filled and focus on these. This may be a profitable approach for sites like the Stripa Mine, where it is very difficult to conceive of a stress related reason for only one of many sub-parallel fractures to be the only conductive fracture. Finally, one can use hydrologic tests focused on known features of the fracture system to determine if they are conducting. Understanding gained through an investigation of fracture style and inference about the nature of fluid flow focuses the hydrologic model and provides insight for predicting the character of fractures in unexposed parts of the rock mass.
In this section, fracture style is briefly discussed from the point of view of fracture growth mechanics. Then an example drawn from a study of fracture systems at the Grimsel Rock Laboratory in Switzerland (Martel and Peterson, 1991) is summarized to show how fracture style was identified and geomechanical arguments were used to infer the nature of flow in the rock mass. Geochemical inference, although clearly quite important, is not discussed here. The use of hydrologic testing to infer the importance of fracture features is discussed in Section 5.

2.1. Fracture Patterns

The fact that fracture patterns often follow common themes reflects common mechanisms for fracture formation. In a significant number of cases, one can describe fractures at specific sites as "variations" on these themes. For example, some of the fracture zones at Grimsel closely resemble simple idealizations. Others reflect relatively complex and varied fracture-forming mechanisms, but even for these cases, recurring fracture pattern elements and growth processes can be identified. Heterogeneity in material properties and stress history control the development of fracture patterns. Understanding the mechanisms that cause fracture patterns common to these features provides a tool for predicting hydrologic behavior where fracture patterns cannot be observed directly.

Fractures, shear zones, and igneous dikes are examples of planar structures commonly found in rock (Figure 2). Fractures can be either joints (dilatant fractures) (Figure 2a) or faults (Figure 2b). Martel and Peterson (1991) point out that joints in massive rocks usually occur in nearly planar subparallel sets (Figure 3a). Joint zones (Figure 3b) can form as groups of clustered, overlapping, subparallel joints (Dyer, 1983). Another theme in joint zones that has been identified by Delaney et al. (1986) is dike-parallel jointing. In this case, joints form during dike emplacement due to extensional conditions induced by the pressurized dike material. In bedded rocks, notably volcanic rocks, polygonal jointing occurs due to cooling stresses. Aydin and DeGraff (1988) have described how these joints nucleate and propagate downward and upward and how this pattern controls the hydraulic connection in the rock. An excellent overview of joints is given in Pollard and Aydin (1988).
Figure 2. Examples of common planar geologic structures. The feature in light gray is an arbitrary maker: (a) joint, (b) fault, (c) shear zone, and (d) dike (after Martel and Peterson, 1991).
Figure 3. Examples of joint patterns: (a) joint set and (b) two joint zones. In the upper zone, joints have formed in front of the longest joint. In the lower zone, the longest joint has propagated past previously formed flanking joints (after Martel and Peterson, 1991).
Martel and Peterson (1991) point out that detailed examinations of isotropic test specimens consistently show that shear fractures form by linking arrays of dilatant fractures (e.g., Peng and Johnson, 1972) or as primary structures parallel to the anisotropy (Donath, 1961). Pre-existing weakness often forms a locus for shear, as has been observed by Muehlberger (1986), Segall and Pollard (1983), Martel et al. (1988), Segall and Simpson (1986), and Lisle (1989), among others.

The dilatant cracks formed in association with faulting are one of the most important features for fracture hydrology. These form oblique to faults in order to accommodate the deformation due to the displacement along the fault. Such fractures are often called "tail cracks" or "horse tails" and occur in predictable locations. These dilatant fractures can link originally discontinuous faults or shear fractures (Figure 4) (Segall and Pollard, 1983; Sibson, 1986; Martel et al., 1988; Martel, 1990) and can be pulled apart by the shearing along the fault. Such "pull aparts," tail cracks, and horse tails all have essentially similar genesis and serve to locally increase the void fraction of the rock. If these dilatant features do not become completely filled with minerals, they may be very important for conducting fluid, whereas the fault or shear fracture itself may be quite impermeable due to the formation of gouge.

When fault zones are reactivated under different stress regimes and different environmental conditions (Muehlberger, 1986; Sibson, 1986), the fracture patterns that develop are likely to be quite chaotic: essentially there are fractures everywhere. In these cases the current state of stress may be the most important determinant of which of these fractures are likely to be most open and thus most conductive.

A few of the important types of fracture themes have been described here, but the above discussion of fracture style is by no means comprehensive. A great deal of work has been focused on describing fracture systems from a geologic and geomechanical point of view, and it is appropriate to draw on this work when constructing a hydrologic model.
Figure 4. Growth of fault zones from a joint set: (a) opening of joints, (b) development of faults, (c) development of simple fault zones, and (d) formation of compound fault zones (after Martel and Peterson, 1991).
2.2. An Example Analysis of Major Geologic Structures from Grimsel

Martel and Peterson (1991) give an excellent example of fracture style description for the purpose of understanding hydrology at the Grimsel Underground Rock Laboratory, which is summarized here. The granitic host rock at Grimsel has been multiply deformed and is strongly foliated. The major geologic structures at Grimsel are fracture zones and metamorphosed, biotite rich lamprophyre dikes. Two types of steeply dipping fracture zones are prominent, K-zones and S-zones. Observations and measurements of fluid flow into the laboratory indicate that these structures and associated fractures account for nearly all of the fluid circulation at Grimsel (Keusen et al., 1989).

Good exposures of K-zones can be found at the surface above the laboratory. A 100-m length of this zone was mapped by Martel (Figure 5) to document how K-zones are structured. This zone strikes NW, nearly at right angles to the foliation. The zone contains two sets of steeply dipping fractures: a set of NW-striking faults are linked by a set of smaller fractures that strike ENE. The K-zone consists of subparallel segments joined by en echelon steps.

S-zones, which have formed parallel to the foliation, display a braided fracture structure. Figure 6 shows a plan view of an S-zone as drawn by Martel. The structure of the S-zones is similar to the microstructure of the foliation of the host rock.

The differences between the K- and S-zones probably result in different hydrologic behavior. Figure 7 shows that fracturing in the K-zones is most intense at the steps. Therefore, it is likely that the steps form large vertical channels, whereas the faults connected by the steps may be less permeable. The braided S-zones are more uniform but are probably anisotropic, because the braiding is more tortuous in plan view than in vertical section. Vertical permeability is probably larger than horizontal permeability in both zones.

Subvertical lamprophyre dikes can be seen at the surface and in the drift. These are highly deformed and very discontinuous. Dike-parallel joints are found, but they are not as spectacular as the horizontal Alpine tension features (Zerklufts) (Figure 8) that extend from the lamprophyres. Some of the Zerklufts are nearly 1 m tall. Dilatant splay fractures near the ends of the
Figure 5. Map of part of a K-zone and a lamprophyre dike exposed at the surface near the Grimsel Rock Laboratory (after Martel and Peterson, 1991).
Figure 6. Photograph of the edge of an S-zone; ruler for scale. Note the braided fracture structure (Martel and Peterson, 1991).
Figure 7. Comparison of K- and S-zones. The rock foliation dips steeply to the southeast (after Martel and Peterson, 1991).
Figure 8. Block diagram showing vertical mullions and mineral-filled horizontal Alpine tension fissures extended from a vertical lamprophyre (after Martel and Peterson, 1991).
Lamprophyres are probably related to faulting within the dikes. These splay features may be quite significant hydrologically.

Lamprophyres are a significant source of water at Grimsel. Fluid can probably flow along the margins of the dikes, and the splay fractures are probably vertical channels. Horizontal flow could be accommodated along the tension fissures. However, flow paths related to lamprophyres are probably mostly parallel to the dikes and may be very poorly connected.

The S-, K-zones and the lamprophyres are responsible for nearly all fluid flow in the Grimsel rock mass. These are therefore the types of features that must be located. Once these are found, something is already known about how fluid might be conducted within the features. Thus this geologic and geomechanical description sets the stage for the remaining investigation.
3. GEOPHYSICAL IMAGING OF FRACTURES

Section 2 described how to determine the type of fracture features that are important for hydrology. Once the type of feature is identified, then the next step is to use geophysical techniques to locate and define these features within the rock mass. A variety of geophysical techniques that are useful for defining fracture systems are not discussed here. These include borehole logging and reflection methods. This section describes some of the seismic and radar cross-hole tomography techniques that can be used to image fracture systems.

With the advancement of medical transmission tomography (e.g., Mersereau and Oppenheim, 1974; Huesman et al., 1977; Slaney and Kak, 1985), applications spread quickly to geophysics, spawning work in seismic ray transmission tomography. Diffraction tomography in acoustic media and electromagnetic ray tomography in electrically resistive media have followed (e.g., Bois et al., 1972; Dines and Lytle, 1979; Olsson et al., 1988b etc.). Although tomograms are not maps of hydrologic properties per se, they offer tremendous promise for fracture hydrology because they allow us to image properties of the rock that are related to the hydrologic properties.

To be able to use a tomogram in a hydrologic investigation, it is important to understand what the image represents and what the limitations of the imaging technique are. For example, acoustic techniques image the elastic properties of the medium, and electrical techniques image the electrical properties. These in turn may be related to the permeability, but the relationship is non-unique. One simple reason for this is that both mechanical and electrical properties are a function of porosity, but unconnected porosity does not result in permeability. Consequently, there is not a unique relationship between permeability and porosity, and a tomogram is not a map of permeability. To be useful for hydrology, a tomogram must be interpreted in light of the other information about the geology of the rock, the expected style of fracturing, and the hydrologic behavior.
This section presents a brief explanation of how tomography is done and a brief discussion of seismic and radar tomography, including some of the relevant physics and new concepts under current investigation. An example is drawn from Majer et al. (1990) of the application of seismic tomography to a known fracture zone.

3.1. Tomography

To produce either a seismic or a radar tomographic image, it is necessary to be able to introduce energy into the rock (through either seismic or electromagnetic sources) at a series of points along a line and to receive the energy at another series of points (receivers) along another line in the same plane (Figure 9). For example, two parallel boreholes can host the sources and receivers. For each point where energy is introduced, the amplitude and travel time of the received wave is recorded at each of the receiver points. Each source-receiver pair is illustrated with a line in Figure 9a, and each line represents a "ray path."

Either the velocity or the amplitude data can be inverted to obtain the spatial distribution of medium slowness and attenuation. The usual method of inversion is some form of iterative algebraic reconstruction algorithm. The section of earth to be imaged is divided into many pixels, which are assumed to have constant physical properties. As the waves pass through each of the pixels, the amplitudes and velocities are dependent on the pixel properties and the assumption is made that the contribution of each pixel can be deduced by back-projecting the rays as indicated in Figure 9b. It follows that a data set consisting of many rays crossing at all angles may be jointly back-projected to yield an estimate of the distribution of velocities in each pixel needed to produce the observed travel times. The attenuation properties of each pixel may be determined in a similar manner, using the amplitude information.

Algebraic reconstruction techniques (ART) developed for this problem are iterative in nature (Peterson, 1986). One equation, i.e. one ray, is analyzed at a time, and the pixel values are continuously updated. The technique works well only in media that have small contrasts. Moreover, ART algorithms give exact solutions only if the ray coverage is adequate, the ray lengths are consistent, the ray paths are determined exactly, and there are no measurement errors. This is
Figure 9. Back projection of rays shown on the left one used to obtain the contribution of each pixel on the right.
rarely the case. Ray coverage is usually 2-sided, 3-sided at best. Often the surface of the earth is used as one of the sides, and these measurements are often of much poorer quality than the borehole data (Robertson and Fisher, 1988; Daley et al., 1988a,b; Majer et al., 1988a,b,c). Such geometry results in incomplete ray coverage and some very short ray lengths. For example, at the top and bottom of Figure 9a, there are two roughly triangular regions where the ray coverage is less than elsewhere. In these regions the resolution will not be as good. In other words, the anomalies will appear smeared and possibly out of place. If the boreholes are very far apart compared with their length, these triangular regions of low ray coverage will cover most of the region being imaged. One of the main differences between medical tomography and tomography in the earth sciences is that ray coverage cannot usually be as extensive in the underground.

For seismic tomography, true ray paths are curved, and they may be determined iteratively along with the pixel values, but straight ray paths are usually assumed. Smearing is caused by a number of problems, including inadequate angular coverage and sampling, assuming straight rays (or any improper ray path), and errors in travel time and station location. For example, if all ray paths must pass through the same feature, this feature will not show up in the tomogram (e.g., an extensive fracture zone parallel to the boreholes).

It is important to recognize that the resulting tomogram is an estimate of the slowness or attenuation structure in the rock, which is not necessarily equal to a map of the fracture zones. Cross-hole tomography can be used in conjunction with reflection techniques to provide additional independent information about the fracture zones. Single-hole reflection data are used to determine reflectors in the same manner as common surface reflection surveys. It is possible to determine where the reflectors intersect the borehole and their orientation relative to the boreholes. The locus of possible reflectors is then a cone passing through this intersection. In radar reflection, the use of a directional antenna makes it possible to estimate the actual orientation of the feature (Olsson et al., 1988b). The cross-hole reflection technique is similar to the single-hole method and can, in principle, uniquely determine the orientation of the reflector. However, the analysis is more difficult.
An integrated analysis of the data is then used to identify major features in the rock to make certain each feature is consistent with the data sets (Olsson et al., 1988a,b). The features are normally assumed to be planar fracture zones. First, a feature is identified in the tomograms and the intersections between the feature and the boreholes are estimated. These intersections can usually be corroborated with a reflector from the single-hole reflection analysis. The possible orientations from the reflection data are displayed in a Wulff polar diagram. A pair of possible planes are determined by the two possible orientations that lie on the locus of possible reflectors that also lie in the plane of the tomogram. If the feature is visible in different tomographic planes, then there is a further three-dimensional check on the geometry. In a similar manner, the cross-hole reflection data are checked for consistency with the location and orientation of the zone. Combining geophysical techniques in this manner usually provides the most confidence that a perceived feature is real.

Beyond cross-corroboration of geophysical measurements, the geophysical results can be interpreted in conjunction with geologic information. In this way, expert judgement is used to eliminate apparent anomalies that are due to imaging errors, and it may be possible to eliminate anomalies that are not due to hydrologic features. An example of such an interpretation is given in Section 4.

3.2. Seismic Tomography

Seismic techniques image the mechanical properties of a rock mass. These properties depend on rock type, porosity, fluid content, and fracture distribution. For example, elastic wave velocities generally increase with increasing rock stiffness and saturation and decrease with increasing rock porosity. Signal attenuation is related to physical parameters in a complex manner.

Water has a significant effect on the propagation of seismic waves. As a rock is saturated, there is a dramatic increase in compressional wave velocity (up to 20 to 30%) during the very last stages of saturation (90 to 100%) (Mochizuki, 1982). This is in marked contrast to the effect of saturation on electrical conductivity, where there is a dramatic decrease in resistivity from zero to
about 50% saturation and not much change after that. The attenuation is also strongly affected by saturation, as shown in Ito et al. (1979).

Imaging the earth with acoustic waves (P-waves) is fairly well established. The P-waves travel faster than the secondary waves (S-waves) and are easily identified as the first waves to arrive at the receivers. The use of S-waves for imaging fracture systems is more recent and could potentially provide more information about the fracture geometry at depth (Crampin, 1978, 1981, 1984a,b, 1985; Stewart et al., 1981; Leary and Henyey, 1985; Majer et al., 1988a,b; Douma, 1988). The idea is to introduce polarized S-waves and determine how the direction of polarization affects the velocity and attenuation. If the particle motion is in the plane of the fractures, the wave will propagate more quickly than if the particle motion is across the fractures. S-wave analysis has the potential for being able to determine the dominant fracture orientations. The outstanding difficulty with this method is that it is often difficult to introduce S-waves of sufficient amplitude into the rock from a borehole at frequencies high enough (100's Hz to 1000's Hz) to image the fracture features of interest.

A recent approach to modeling seismic wave propagation is to treat fracture interfaces as a boundary condition in the seismic wave equation. At the boundary, seismic stress is continuous, but seismic particle displacements are discontinuous, i.e., the fracture is modeled as a displacement discontinuity (Schoenberg 1980, 1983). The ratio of the stress to displacement across the interface is the specific stiffness, which defines the elastic properties of the fracture. For a completely elastic system, this displacement discontinuity theory results in frequency-dependent reflection and transmission coefficients for each interface as well as a frequency-dependent group time delay (Pyrak-Nolte et al., 1990). The theory explains concurrent changes in velocity and attenuation as a function of frequency and indicates that each fracture adds to the delay of the seismic wave and decreases the amplitude by some factor. Thus, as the wave passes through four fractures, the velocity delay will be twice as much and the attenuation four times as great compared with a wave passing through two fractures. The power of this theory is that it forms a basis for the interpretation of tomographic data that, theoretically, can provide indications of fracture
spacing, size, and orientation.

3.3. Radar Tomography

Electromagnetic techniques such as radar are used to sense variations in parameters such as electrical resistivity, dielectric constant, and magnetic permeability (Telford et al., 1976). In many rocks, it can be assumed that the rock matrix properties are relatively constant and the water content dominates the electromagnetic behavior. In this case, hydrologic features such as fracture zones, increased porosity, and water content will dominate. However, in general, non-fluid-conducting porosity cannot be distinguished from conducting porosity.

Radar signals are sensitive to changes in dielectric constant and electrical conductivity (Sen et al., 1981). Radar velocity is a function of \( c/\sqrt{\mu\varepsilon} \), where \( c \) is the electromagnetic wave velocity, \( \varepsilon \) is the dielectric constant, and \( \mu \) is the magnetic permeability. The amplitude is a function of \( (\sigma/2)^2\sqrt{\mu/\varepsilon} \), where \( \sigma \) is the electrical conductivity. The electromagnetic wave velocity, \( c \), and magnetic permeability, \( \mu \), are, for all practical purposes, constant. Slowness, therefore, depends only on the dielectric constant, \( \varepsilon \), and attenuation depends on both the dielectric and the electrical conductivity, \( \varepsilon \) and \( \sigma \).

Both the electrical conductivity, \( \sigma \), and the dielectric constant, \( \varepsilon \), increase with the water content. On the basis of the functions given above, the slowness tomogram should give the most direct correlation with fluid paths in the medium, because it depends only on contrasts in \( \varepsilon \). However, rock exhibits more conductivity contrast than dielectric contrast. As a result, the attenuation tomograms can be more useful for imaging differences in rock properties.

As with most geophysical techniques, radar is most powerful when it is asked to detect changes in the system. For this reason, radar tomography has been used both before and after the injection of saline water into the rock (Olsson et al., 1991). Saline water increases the electrical conductivity of the hydrologically conductive features, and as a result the difference tomogram highlights these features dramatically. With this technique it is possible to essentially "see" water flowing through the rock.
3.4. Imaging a Fracture Zone with Seismic Transmission Tomography — An Example from the Grimsel Rock Laboratory

The Fracture Research Investigation (FRI) in the Grimsel Rock Laboratory was designed to test, calibrate, and improve technology for imaging and characterizing fracture zones with seismic tomography (Majer et al., 1990). The dominant feature in the FRI site is a mylonitic fracture zone crossing between two parallel drifts 20 m apart. The fracture zone is visible in both drifts and was considered to be a feature whose location was known *a priori* (Figure 10). Because water could be seen dripping out into the drift from the zone, the zone was considered to be hydrologically active. The object of the experiments was to gather high-quality P- and S-wave data across the fracture zone to determine the seismic visibility of this known hydrologically active fracture zone.

In addition to the seismic studies, a detailed geologic investigation provided the appropriate background and insight for interpretation of the tomograms, and a limited number of hydrologic tests were used to check the visual observation that the fracture zone was hydrologically active and to see if other features marked by seismic anomalies might also conduct fluid. Finally, attempts were made to modify the fracture zone properties by inflating the fracture, measuring the stiffness and permeability change *in situ*, and observing if a simultaneous change in the seismic response could be detected.

Two boreholes were drilled between the drifts, one passing through the fracture zone, and the other 10 m away. Consequently the fracture zone was accessible from four sides. Boreholes 87.001 and 87.002 are 86-mm holes drilled from the lab tunnel to the access tunnel to provide a means of performing cross-hole seismic work, obtaining core through the fracture zone, and carrying out hydrologic experiments. Borehole 87.003 is a 127-mm hole drilled through the fracture zone for obtaining large core for laboratory analysis and also for hydrologic testing. In addition to these holes, 76 shallow holes were drilled into the lab and access tunnel walls between boreholes 87.001 and 87.002 at 0.25-m spacing to allow the placement of the seismic sources and receivers in order to obtain ray coverage from all four sides.
Figure 10. (a) Geologic plan view of FRI site. (b) Schematic diagram of FRI zone and Grim­sel Test Facility. (c) Perspective drawing of the FRI site.
Seismic sources were placed in the holes (boreholes 87.001 and 87.002 and the shallow holes in the sides of the tunnel) and activated. The data from a three-component accelerometer package was recorded at 0.5-m spacing in boreholes 87.001 and 87.002. The receiver package was also placed in the shallow holes to give complete four-sided coverage. Nearly 60,000 ray paths (x-, y- and z-components) were collected in the FRI zone, with lengths from 0.5 m to nearly 23 m. The peak energy transmitted in the rock was 5,000 to 10,000 Hz, thus yielding a wavelength of approximately 1 to 0.5 m in the 5.0-km/sec velocity rock. Two surveys were completed, one in 1987 and one in 1988. The 1987 and 1988 experiments were essentially similar, but the use of an improved source and different clamping in 1988 resulted in a higher received frequency content and increased signal power.

After the data were collected and acceptable first arrival time-values were obtained, the travel times were inverted. A pixel size of 0.25 m was chosen because this was expected to be the size of the smallest anomaly that could be seen given the average wavelength of 0.7 m and station spacing of 0.5 m. Figure 11 shows the results of inverting the 1987 data.

The main features identified in the 1987 results are the large shear zone (Feature A) extending from the middle of borehole 87.001 to the intersection between the access tunnel and borehole 87.002 and the low-velocity zones adjacent to the tunnels, assumed to be damaged zones. Feature B (an anomaly roughly orthogonal to Feature A) and Feature C in the south of the area are also visible.

Figure 12 is an image that was produced using the 1988 data set. The differences between the 1987 and 1988 results are these:

1. The prominent shear zone observed in 1987 (Feature A) is shown as a single strong low-velocity zone about 2 m wide. The corresponding zone in the 1988 results consists of two or three very thin (< 0.5 m thick) zones that become discontinuous at about 4 or 5 m from the laboratory tunnel.

2. Feature B was partially masked by the low-velocity zones bordering the tunnel walls in the 1987 result but is very prominent in the 1988 case. This feature is the dominant
Figure 11. Tomographic image of FRI from 1987 survey.
Figure 12. Tomographic image of FRI from 1988 survey.
feature of the 1988 results.

(3) Feature C was more prominent in 1987 than in 1988.

(4) There is little evidence of the extensive 1987 damage zones near the tunnel walls in the 1988 results, and the average velocity values in the 1988 field are slightly higher.

Compared with the 1987 results, the shear zone (Feature A) in the 1988 tomogram appears to produce a weaker velocity anomaly, although the actual numerical differences are not great. The 1988 results indicate that Feature A is not a simple single planar feature, and thus the permeability along the zone may also be variable rather than being a single well-connected feature.

Possible reasons for the differences between the 1987 and 1988 tomograms with respect to the main shear zone (Feature A) can be identified through geomechanical studies aimed at relating the mechanical stiffness of a fracture to the behavior of acoustic waves passing through the fracture. Values of fracture stiffness were estimated using the in-situ deformation measurements (Majer et al., 1990). For these values of stiffness (about $10^{11}$ Pa/m), the difference in the 1987 and 1988 dominant frequencies would result in significantly more delay of the wave traveling across fractures in 1987 than in 1988. Thus the displacement discontinuity theory alone could easily account for the smaller anomalies in 1988. The displacement discontinuity theory also predicts that as frequency increases, for a constant stiffness, the velocity or delay caused by the fracture becomes less important relative to the attenuation (Pyrak-Nolte et al., 1990). Apparently the frequencies used and the value of stiffness of the fracture were such that attenuation is important. If attenuation was great enough to eliminate the first arrivals, then the 1987 tomogram was effectively a mixed velocity-attenuation tomogram.

The most unexpected result from the 1988 inversion is the dominance of the low-velocity feature (Feature B), which extends from the intersection of the access tunnel and BOFR 87.002 to the large low-velocity feature near BOFR 87.001. Again, the difference between the 1987 and 1988 data in this region is not large. The visible fractured area at the access tunnel where the anomaly begins consists of subhorizontal fractures and a tension fissure. From geologic considerations, it is most likely that this feature is associated with a lamprophyre or an especially large
tension fissure, but this has not been confirmed.

Feature C, extending, approximately 8 m along borehole 87.002 from the access tunnel, may be due to small fractures or, as laboratory work suggests, a difference in the rock type. The darker colored core from this vicinity does not show significant fracturing, but the core velocities are lower for this type of rock than for the lighter colored granite elsewhere.

The prominence of the low-velocity zones associated with the damage zones around the tunnels in the 1987 results may be due to the initial pulse of the signal being highly attenuated. Consequently, an artificially low-velocity result was obtained by picking later arrivals. The attenuation data from 1988 also indicate the presence of damage zones near the tunnels. Again, this suggests that at the frequencies of 5 to 10 kHz, the effect on the velocity was much less than that on attenuation.

The 1987 seismic results were used to guide the well test design. The hydraulic tests confirmed the hydrologic significance of the kakirite-bearing fractures in the FRI zone identified by the seismic tomography. These are discussed in Section 5. Other tests indicated the possibility that leakage out of the plane of the fracture could occur along secondary features that were seen as anomalies in the tomograms.

The results of the seismic field work at Grimsel indicate that the original premise of using P- and S-waves for mapping fracture content is valid. The main fracture zone in the FRI site was detected using P-wave tomography. The S-wave was completely attenuated by the fractures and could not be detected. However, given a strong enough S-wave source, one might find that S-wave data would be even more sensitive to fracture content than P-wave data. Except for Feature B, all the anomalous velocity zones are coincident with geologic structures or anomalies associated with the stress relief at the tunnel walls.

Both seismic and electrical cross-hole tomography may be useful techniques for non-destructive imaging of rock properties over distances as great as a few hundred meters. It is important to note that tomograms must be interpreted with other information. Perceived anomalies on tomograms do not correspond uniquely to geologic features in a rock mass. The
inversion process itself commonly produces artifacts that must be distinguished from the anomalies associated with real geologic structures. Furthermore, anomalies on tomograms can correspond to a variety of geologic effects. Independent information on the geology can be used to determine which geologic features are most likely represented in the image.

The most promising potential for determining which anomalies are important lies in combining different types of information. At FRI certain frequencies could effectively produce a combined attenuation and velocity tomogram that gave a vivid image of the known hydrologic feature. Likewise shear wave information has great potential for sorting out the structure of fractured rock. Olsson et al. (1988a,b) show an example where the velocity tomogram and the attenuation tomogram give complementary anomalies, and this turns out to be the signature of a certain type of leached zone in the granite. How to best combine geophysical information to make a relevant interpretation is a topic of ongoing research.
4. INTERPRETATION OF THE HYDROGEOLOGIC STRUCTURE OF THE FRACTURE SYSTEM FROM GEOLOGIC AND GEOPHYSICAL INFORMATION

Section 2 described how geologic and geomechanical analysis can be used to describe the fracture features that are important for hydrogeology. Application of the geologic approach to a field site includes mapping exposures of the important features. Section 3 described how geophysics can be employed to image the fracture structure inside the rock. This section shows how information from geophysical and geologic investigations can be combined to determine the three-dimensional geometry (position and orientation) of the major structures that might conduct water, i.e., a geometric model of the geologic structure. Geologic and geophysical investigations clearly can complement each other. Geologic investigations are well suited to identify, locate, and characterize exposed features, but they are limited in their ability to determine how far to project known features and to detect unexposed features. On the other hand, geophysical investigations can locate unexposed features but are limited in their ability to determine uniquely the type of geologic features they detect.

4.1. Methodology

There are two types of geologic information available for integration with the geophysical data: geologic maps made at outcrops and underground exposures and data from boreholes. Maps present a system in the form of an integrated picture rather than as a series of unrelated points. Small-diameter boreholes provide the least expensive way to sample the geology directly within an unexposed volume of rock, but the interpretation of borehole data has limitations. For example, borehole records typically do not allow the shape and dimension of fractures to be well defined, nor do they provide information about the relationship between fractures. In addition, the orientation of a fracture zone can differ significantly from the orientation of the fractures in the zone. Figure 13 shows a simple example of how the same borehole fracture record can result...
Figure 13. Two markedly different fracture zones can have the same appearance where they intersect a borehole (shown in heavy line): (a) a series of joints, and (b) a fault zone. Dotted box is for reference.
from entirely different fracture configurations. Work on the structural systematics can reveal how individual fractures are arranged in fracture zones and therefore can be extremely valuable in interpreting borehole fracture data.

A clear use of geophysical information is to help project features within a site. A key contribution of geologic information is to prevent geophysical data from being interpreted blindly. To produce a preliminary model of the site structure, geologic structures either exposed in the site vicinity or inferred from geophysical data are projected into the site. The model is revised to incorporate the results of site-specific geophysical tests. The geophysical and borehole information should be interpreted in a manner consistent with the systematics of the local structures. The model should be re-examined and refined as more site-specific information becomes available.

The result of this process is an interpretation of the geologic and geophysical data for the purpose of representing the important hydrologic features. Both the gross arrangement of the major structures and the information on the internal systematics of the major structures should be considered. The interpretation is extremely useful in framing a hydrologic testing plan as described in Section 5. The combination of the structural and well testing interpretation is the basis for a hydrologic conceptual model, as shown in Section 6.

4.2. An Example of Construction of a Fracture Zone Model from Grimsel

As an example of the integration of geologic and geophysical data, experience with data taken from at the Grimsel Laboratory in Switzerland (Figure 14) is summarized from Martel and Peterson (1991). Martel and Peterson focus on an area of the Grimsel Laboratory called the US/BK site, which has a cross section of about 150 by 150 m and is intersected by a number of the major geologic structures presented in Section 2. Site-specific data from surface and subsurface mapping and boreholes were combined with geophysical tomography to construct a geometric model of the major geologic structures at the US/BK site. Martel and Peterson then examine how well this model would have predicted the flow behavior inferred from the results of difference tomography. Radar difference tomography was performed at the US/BK site before and after injection of a brine solution (Niva and Olsson, 1988a,b). Because the brine has a readily
Figure 14. Map showing the tunnels and major boreholes in the vicinity of the US/BK site. Boreholes BOUS 85.002 and BOUS 85.003 are considered as bounds on the site, as is the laboratory tunnel. L60, L100, and L200 mark distances (in meters) along the laboratory tunnel (after Martel and Peterson, 1991).
detectable electromagnetic signature, the difference between the "before" and "after" tomographic data highlights hydrologically conducting features. In this way it is possible to see if the conducting features match any features in the geometric model.

As discussed in Section 2, three main systems of geologic structures have been identified in the vicinity of the Grimsel laboratory: K-zones, S-zones, and lamprophyre dikes (Figure 15). Geologic maps of the surface and the Grimsel laboratory (Keusen et al., 1989) show that many of the major structures extend to depth as roughly planar features.

The major features of Martel and Peterson's preliminary model (Figure 16), based on surface and subsurface geologic information and on borehole data, are (from north to south):

Feature 1: a discontinuous series of three northeast-striking S-zone segments,
Feature 2: a lamprophyre-bearing K-zone north of the BK room,
Feature 3: some northwest-striking lamprophyres,
Feature 4: a west-striking K-zone south of the BK room.

Features 1, 2, and 3 can be correlated with major structures that are mapped at the surface and shown near the northern border of Figure 14. There is no K-zone at the surface that corresponds to Feature 4. The S-zone consists of discontinuous segments (1a, 1b, and 1c) that are separated by lamprophyres. This interpretation is consistent with surface observations discussed in Martel and Peterson (1991). The S-zone segments may have formed part of a once-continuous structure that was offset by slip across the lamprophyres, but the segments may also have formed part of a structure that was originally discontinuous.

Seismic tomograms (Gelbke, 1988) and radar tomograms (Niva and Olsson, 1987, 1988a,b) were used to check and extend the geologic model of the US/BK site. Both kinds of tomograms were produced using signals transmitted between the laboratory tunnel, borehole BOUS 85.002, and borehole BOUS 85.003. The tomograms are thus "three-sided."

Martel and Peterson (1991) defined several major low-velocity anomalies on the seismic velocity tomogram using the 5050 m/sec contour. In Figure 16, the major features inferred from
Figure 15. Block diagram showing major structures in the vicinity of the Grimsel laboratory. View is to the west. Zones shown in outline are S-zones. Structures marked by solid shading are K-zones and lamprophyre dikes. Numbers along main access tunnel mark distance in meters from its north-entrance (after Martel and Peterson, 1991).
Figure 16. Projection in the plane of tomography showing the features of the preliminary structural model of the US/BK site superposed on the 5050 m/sec contour from Figure 12. The geologic features are marked by circled numbers. The seismic anomalies are marked by uncircled numbers. Seismic anomalies S1-S5 are described in the text (after Martel and Peterson, 1991).
the geologic data are superposed on these anomalies. The geologic features are indicated by circled numbers and the seismic anomalies by uncircled numbers.

Martel and Peterson (1991) conclude from the location of anomalies S1a and S1b that the tomogram supports the geologic model for Feature 1 as a series of en echelon S-zone segments separated by lamprophyres. However, there is no anomaly corresponding to Feature 1c. The seismic anomaly S2 corresponds to the lamprophyre, Feature 2, but anomaly S2 is only in the vicinity of the intersection of Features 1 and 2. Martel and Peterson conjecture that the anomaly may be due to increased fracturing near the intersection of these two features.

Martel and Peterson (1991) propose that the anomalies S3b and the northwest part of S2 are indicative of lamprophyres that are found in corresponding locations at the surface but are not seen any other way at depth. Because Feature 4 does not show up as an anomaly on the seismic tomogram, these authors conjecture that K-zones such as Feature 4 may only create geophysical anomalies where there is a step in the zone.

Martel and Peterson (1991) also compared the geologic model with the radar tomograms collected by Niva and Olsson (1988a,b). Although the radar tomograms did not show continuous, well-defined anomalies that correspond to the features of the model, they did show distinct anomalies where the geologic features intersect. Comparison of the geologic maps with the seismic and radar tomograms led Martel and Peterson to add Feature 3, the northwest-trending lamprophyres, to the model, as shown in Figure 17. They conclude that the tomograms support the geologic model over the previous proposals developed by Keusen et al. (1989) and increase confidence in the utility of tomography in projecting the major geologic features into the target site.

Martel and Peterson (1991) checked this revised model against radar difference tomograms to see how well the model identified major flow paths. The difference tomograms were obtained by injecting brine in two phases at points shown in Figure 18 and repeating the tomographic measurements after each injection (Niva and Olsson, 1988a,b). Difference tomograms (Figures 19 and 20) are prepared by inverting travel-time or amplitude differences between two tomographic
Figure 17. Projection in the plane of tomography showing the revised model of major geologic structures at the US/BK site. The strike and dip of the major features are shown by the heavy line (after Martel and Peterson, 1991).
Figure 18. Projection in the plane of tomography showing where brine was injected during the phase 2 and phase 3 tomographic measurements. Tick marks are on a 50-m grid (after Martel and Peterson, 1991).
Figure 19. Difference tomogram of radar attenuation structure between BOUS 85.002 and BOUS 85.003 from phase 1 and phase 2 measurements. The tomogram shows the increase in radar attenuation and indicates where brine has migrated during phase 2 (from Niva and Olsson, 1988a, Figure 5.12). Units are in dB/m. North is to top of page.
Figure 20. Difference tomogram of radar attenuation structure between BOUS 85.002 and BOUS 85.003 from phase 2 and phase 3 measurements. The tomogram shows the increase in radar attenuation and indicates where brine has migrated during phase 3 (from Niva and Olsson, 1988b, Figure 5.26). Units are in dB/m. North is to top of page.
surveys. These tomograms show how the region being analyzed in the plane of the tomography changed between test phases, allowing the brine paths to be traced. The brine will locally increase the radar attenuation, and it may decrease the radar velocity (Sen et al., 1981). Figure 19 was prepared using attenuation data from phases 1 and 2; it reveals the flow of brine injected during phase 2. Figure 20 was prepared using attenuation data from phases 2 and 3; it reveals the flow of brine injected during phase 3. The radar attenuation difference tomograms proved to be much more reliable in locating the brine than the slowness tomograms.

The results from the phase 1–2 and phase 2–3 radar attenuation difference tomograms are, on the whole, consistent with the predictions of the structural model. Most of the brine displayed in the phase 2 radar attenuation difference tomogram (Figure 19) appears to be contained within this S-zone segment 1b (see Figure 17). The position and shape of the brine anomaly in the phase 3 radar attenuation difference tomogram (Figure 20) indicates the phase 2–3 brine flow was concentrated along a path that extends about 10 m southeast from the injection point and then heads southwest toward borehole BOUS 85.003. This is consistent with flow being strongly controlled by both the hydrologic gradient and Feature 1c (see Figure 17). Both radar attenuation difference tomograms are consistent with the interpretation that Feature 1b (Figure 17) does not continue on strike to the south across the Feature 3a lamprophyres (Figure 17). The radar difference tomograms increase confidence in the interpretation of the geologic structure at the US/BK site. The features expected to carry flow were highlighted, and the features expected to be barriers seem to have impeded flow.

This study shows the unique contribution of the radar/saline difference tomograms. The difference tomograms highlight hydrologic features at the site that were not identified in any other way. If all of the anomalies in the attenuation difference tomograms (e.g., Features a, b, c, and d in Figure 20) accurately represent the location of significant amounts of brine, then some portion of flow at the US/BK site is following a network of fractures that were not identified as major features from the geologic or previous geophysical investigations. Without the saline/radar information, the contribution of these fractures would either be missing from the model or would
have to be included in a random way.

In conclusion, geologic and geophysical information can be integrated to give a consistent model of the major geologic structures in rock volumes with dimensions on the order of 100 m. Geologic observations establish the position, orientation, and type of structures near a site. Geophysical measurements and borehole data aid in projecting the structures within a site. Detailed geologic observations of the internal structure of the major features can provide insight into how fluid might flow along the structure, and radar difference tomography can image how tracers actually do flow. The tomograms suggest that the intersections of major geologic structures could be sites of particularly extensive fracturing and enhanced fluid flow. This study shows that an integrated geologic and geophysical investigation can contribute greatly to hydrologic site characterization.

In the case described above, there were many good geologic exposures. For many other applications, such exposures will not be available and geophysical tomograms will become the main source of information about the structure of the rock. In these cases it will be especially important to remember that anomalies in tomograms can reflect a wide range of features (different rock types, fractures, zones of hydrothermal alteration, areas of increased porosity, etc.), and not all major throughgoing geologic structures are represented as continuous anomalies on tomograms.
5. HYDROLOGIC WELL TESTING IN FRACTURED ROCK FOR CONCEPTUAL MODEL DEVELOPMENT

The previous sections have laid out some technology for developing a structural model of the major fracture features. To go from this stage to a hydrologic conceptual model for flow, it remains to be seen if these structural features do conduct fluid. There is probably no better way to see if a fracture conducts fluid than to try in some way to pump fluid through the fracture and monitor the hydraulic response in the rock. This section discusses the planning and interpretation of such hydrologic testing for the purpose of developing a hydrologic conceptual model. The construction of the hydrologic conceptual model is discussed in the next section (6).

The structural model of the fracture features provides a framework for designing the hydrologic investigation. If the structural model includes a major feature, then the well testing program can be designed to investigate the permeability of this feature. If there are a number of features, the well tests can be designed to see if they are hydraulically connected. The interpretation of the well tests should lead to a conceptual model for fluid flow in the rock; the model should describe the major hydrologic features explicitly and indicate the parts of the flow system that can be treated with an equivalent model.

Interpretation of a well test is a type of inverse method: the observations on the behavior of the system are used to estimate the system parameters. However, such an inversion always requires the assumption of some type of conceptual model. In a pumping test, for example, a flow rate can be induced at one well and the drawdowns monitored at an observation well. If the pumping well and the observation well are assumed to be connected by a pipe, an estimate of the diameter of the pipe can be derived from the well test data. If the two wells are assumed to fully penetrate the same isotropic, infinite parallel plate fracture, then the aperture or permeability of the fracture can be estimated. Likewise, if the wells are assumed to penetrate a massive continuum or a fracture connected to a boundary of some type, the parameters for these systems can be
calculated. Any quantitative interpretation of well test data must include such underlying assumptions.

In the case of well testing to support conceptual model development as described here, the interpretation is based on the supposition of a series of simple conceptual models with well-defined system geometry. The data are compared to analytical "data" that would be generated by each of these models under the same test conditions. It is the relative closeness of fit of the data to the different models that provides insight about the hydrology of the system. On the basis of this insight, a conceptual model is adopted. However, there may be a number of possible conceptual models with different combinations of geometry and flow parameters that give the same observed response at a given point. Even worse, none of the well-defined conceptual models may match the data at all. For this reason, there is always an inherent question about the uniqueness of any conceptual model. The degree of uncertainty can be reduced by the previously described geologic and geophysical studies and, if possible, by having many hydrologic observation points.

The inversion of hydraulic well test data on the basis of the adopted conceptual model can then be used to help define the parameters of the model in order to make a predictive tool. An example of well testing for conceptual model development is given here, but the example of using well testing data for inversion to obtain a predictive model is deferred until Section 7. The fundamentals of well testing are the same in either case. However, it is very important to realize that the parameters derived from a well test interpretation are always linked to the conceptual model, which is the basis of the interpretation, and one can never be sure of the conceptual model. This means that any parameters derived from well test data are only as good as the conceptual model.

5.1. Some Considerations for Hydrologic Tests in Fractured Rock

The hydrologic borehole tests that can be done in fractured rock are essentially the same as can be done in porous media: hydraulic tests and tracer tests. The theory supporting interpretation of these tests is well established for a variety of geometries of porous media under well-defined boundary conditions. For fractured rock, the standard analysis of these tests requires that
analyses be made between the fracture system and an equivalent porous system. A fracture zone, for example, might be treated as an aquifer.

Hydraulic tests are diffusion dominated. This means that the location of the perturbation and the location of the response are known, but the geometry of the flow paths between them is not (unlike seismic wave propagation, where it is possible to estimate the ray paths). As a result, each well test response is a result of some average conductance where the averaging process is unknown. Tracer tests are advection dominated, and only the properties in the direct path of the tracer affect the results of the test. However, it is not possible to interpret these properties without independent information about the flow path.

From a single-well steady-state test, it is not possible to determine anything about the arrangement of the conductors. The data from a steady-state test will only reflect the bottleneck, or least conductive part of the flow path between the perturbation point and the boundary. The test is not particularly sensitive to the arrangement of the conducting elements in space. Transient tests are a little better in that the volume that controls the average conductance changes with time. So it is possible to say more about the possible arrangement of average conductance as a function of distance from the well. Probably the most effective hydrologic testing is conducted in a cross-hole or interference manner. Cross-hole tests offer the only possibility for providing information about the disposition and interrelationship among the highly conductive features.

For transient tests, constant flow tests have some advantage over constant head tests. Analytical solutions are more readily available for constant flow tests than constant head tests. The majority of the published papers in the literature are based on constant flow rate tests. Reciprocity holds for constant flow tests over all time. That is, if you pump q 1/min from well A, the drawdown at well B will be the same as the drawdown at A when B is pumped at the rate of q 1/min. This can be an important attribute for the interpretation of well tests in any heterogeneous material. The same is not true for constant head tests. However, in very low permeability rocks, constant pressure is generally easier to achieve than a constant rate.
As a final point about well testing in fractured rock, it is worth mentioning that it can take a long time to do a well test, often months. Further, once a perturbation has been introduced by the test, it can take just as long for the system to return to equilibrium before another well controlled test can be run. Leaking packers, holes left open by accident, and pump failures are more common than not. The result is that it is very difficult to obtain high quality well test data. Experience has shown that the collection of useful data requires that sufficient time be allowed for each test and further time be allowed to repeat tests that fail. Lack of reliable data is probably the single most important difficulty in well test interpretation.

5.2. An Example of Conceptual Model Testing at FRI

Karasaki (in Majer et al., 1990, Chapter 5) gives an example of conceptual model testing based on hydraulic test interpretation for the FRI fracture zone (see Section 3). This work is summarized below. The design of the tests was guided by the results of the geologic investigation and the seismic tomography at the FRI site. Several possible conceptual models were defined and compared with the data to help determine which conceptual model fits best.

The hydraulic tests were planned on the basis of 1987 tomography results (Figure 11). Packer locations are shown in Figure 21. Each test consisted of pumping water in the interval at a constant pressure and monitoring in all the other intervals. Objectives of the test were

1. to find hydraulic connection with other zones,
2. to characterize the properties of zones that are hydrologically active, and
3. to determine the existence of zones about which there is only inconclusive evidence.

The purpose of Test 1 was to provide a hydrologic characterization of a feature that is clearly evidenced by geophysics. The packers for Interval I1.2 were placed such that they confine the main fracture zone (Feature A) as tightly as possible in order to minimize wellbore storage and isolate the hydrology of the feature. Interval I1.2 was used as the inflow interval, and pressure was monitored in all the other intervals (Wyss, 1988). Test 1 is discussed here. Tests 2 and 3 from zones I2.1 and I1.3, respectively, were inconclusive. The results of Test 1 indicate that Feature A is clearly the most significant hydrologic feature at the FRI site, as expected.
Figure 21. Packer locations used in the hydrologic tests.
Figure 22 shows the pressure transient of interference data at various observation points in Test 1. Note that the interval 13.1 responds most markedly. Data for the response at 13.1 are compared to those for the theoretical response (Figure 23). As can be seen from the figure, the pressure observed at 13.1 is significantly lower than that predicted by the analytical solution, although the shapes of the curves are almost identical. The analytical solution assumes that the fracture is infinite, isotropic, and homogeneous. Therefore, conditions must exist where one or more of the above assumptions are not appropriate. Karasaki (in Majer et al., 1990, Section 5.4) lists the plausible scenarios as:

1. **Skin**: There is a low permeability zone around the injection well, i.e., a skin that causes the effective pressure at 11.2 to be lower.

2. **Anisotropy**: The fracture is anisotropic where the maximum permeability direction is oriented vertically.

3. **Leakage**: There is leakage from the fracture to the adjacent rock, possibly through other minor fractures, so that the pressure is more dispersed.

4. **Boundary Effect**: The boundary effect of the laboratory tunnel is keeping the pressure low at 13.1.

All of the above conditions may coexist, and, as is the case with any field experiment, the possibility of an erroneous measurement should not be completely discounted. Each of the four scenarios is evaluated below by comparing the data to model results.

**Skin**: Skin effect is usually suspected when an anomalous result is obtained. However, the flow rate curve (not shown) does not match any of the skin curves, and the match for the response at 13.1 is not very good. Therefore, it seems that the conventional skin concept does not explain an observed behavior. Assuming a constant pressure drop at the borehole wall independent of the flow rate will result in a flow rate curve that is identical to that of no skin. This no-skin curve is closer to the observed curve, but it still does not explain an observed inflection in the flow rate curve.
Figure 22. Interference buildup data for test 1 at various observation points.
Figure 23. Comparison between the data and the theoretical response.
Anisotropy: Geologic observations (Section 2) indicate that the fracture zone may be anisotropic, with the highest permeability in the vertical direction. Thus the injected water may flow preferentially in the vertical direction. Preferential vertical flow could cause the observed pressure head in the horizontal direction at I3.1 to become lower than that in the isotropic case. Analytical solutions for flow to a well in an anisotropic medium were obtained through a transformation of coordinates (Kucuk and Brigham, 1979). However, an unreasonably large anisotropy ratio \((2 \times 10^9)\) was necessary to explain the pressure drop. Therefore, anisotropy is not likely to be the cause of the low pressure measurement. Of course, the fact that an unreasonable amount of anisotropy would be required to explain the well test results does not preclude the existence of anisotropy.

Leakage: In the above conceptual models, it was assumed that the flow is confined within the fracture zone. However, as can be seen from Figure 22, interference responses, although small, were observed at various intervals that are not in the plane of the fracture zone. This implies that there was a leakage from the fracture zone into the adjacent rocks, which may explain why the interference response at I3.1 was low. The solution for pressure due to a constant pressure test in a leaky aquifer is not readily available in the literature. However, if the rock into which the leakage flows is assumed to be of finite thickness and the leakage is at quasi steady-state, the solution presented by Da Prat et al. (1981) for a double porosity medium can be used. Although the flow rate match with the Da Prat solution is very good, the match with the observed pressure at I3.1 is not good at all. Probably the theoretical pressure is too high compared with the data, because the Da Prat solution assumes that the rock above and below the fracture is a finite size.

Leakage into an infinite size rock can be considered using the Laplace space solution for the normalized pressure in the fracture zone at a nondimensional distance, \(r_D\), under a constant pressure test. This solution for flow and drawdown is plotted in Figures 24 and 25, respectively, for various values of conductivity ratio between the fracture and the rock. Also plotted are the observed data. The match with the interference data at I3.1 is now much better than the match
Figure 24. Type curve match at 13.1 with a leaky fracture zone solution.
Figure 25. Type curve match at 11.2 with a leaky fracture zone solution.
with Da Prat's solution. However, the late time data of I3.1 are still not matched very well, and the observed flow rate curve is much flatter in the late time than the theoretical curves (Figure 25).

Nonetheless, the concept of leakage seems to explain the trend of the data: low interference pressure and the flattening of the flow rate curve. The weak hydrologic connection between Intervals 11.2 and 12.2 may be through this low-permeability rock matrix. It is worth noting that the 1988 seismic tomography results indicate the existence of a Feature B that extends diagonally from the access tunnel toward BOFR87.001 (Figure 12). This may be the actual conduit of the leaking water. Although a localized leakage cannot be handled with an analytical solution, it could explain the low pressure at I3.1 and the flattening of the flow rate curve.

**Boundary effects:** In the previous analyses, the boundary effects of the tunnels were neglected. However, during the injection test it was observed that water was seeping out thorough the shot-crete along the zone where the FRI fracture intersects the access drift. Because of the complexity of the geometry and the boundary conditions, a numerical model was used to examine the effects of the drifts on the measured pressure. Both models were assumed to be initially at steady-state subjected to the same hydrostatic head. Then the node that corresponds to the intersection of Interval 11.2 and the fracture was opened to simulate the field test. Results of the simulated well test were compared using the cases with and without tunnels. On the basis of this analysis, the effect of drainage into the laboratory tunnel on the interference data at I3.1 seems to have been minimal, which indicates that the permeability of the fracture around the laboratory tunnel is low. This agrees with the observation that no apparent increase of water seepage was noted in the vicinity where the FRI fracture intersects the laboratory tunnel.

The hydraulic tests have confirmed the hydrologic significance of the fracture zone, which was previously identified by seismic tomography. It appears that the majority of the flow occurred within the relatively thin fracture zone that connects Intervals 11.2 and 13.1 (Feature A). A weak but definite hydrologic connection between Intervals 11.2 and 12.2 was also observed. Feature B, identified by the seismic tomography that extends diagonally from the access tunnel to
BOFR87.001, may explain this hydrologic connection.

Although it is possible for skin, anisotropy, leakage, and boundary effects to coexist, the most plausible scenario seems to be the leakage effect outside of the fracture plane, possibly through Feature B. From this analysis, it is clear that a conceptual model for the FRI site would explicitly include the main fracture zone. Within this fracture zone, flow might be modeled using some equivalent flow system. Beyond the fracture zone itself, the conceptual model should include some conductive elements related to Feature B.
6. CONSTRUCTION OF THE HYDROLOGIC CONCEPTUAL MODEL

Once the geologic, geophysical, and hydrologic techniques investigations have taken place, the stage is set to specify the hydrologic conceptual model. As used here, the hydrologic conceptual model is taken to be:

(1) An explicit geometric description of the major hydrologic features.

(2) A format for developing an equivalent flow model in the major features or in minor features not associated with a major feature (e.g., the lattice template of an equivalent discontinuum).

To go from the conceptual model to a predictive model, it remains to deduce specific configurations of the equivalent model on the basis of the hydrologic data. In this work, the use of inverse techniques for deducing model configurations is emphasized. Consequently, this section describes the selection of a conceptual model as the basis of an inverse technique called "simulated annealing," which is discussed in Section 7.

An alternative would be to present a stochastic conceptual model and use statistical simulation to obtain predictive models (Billaux et al., 1989; Geier et al., 1990; among others). This approach was discussed in the introduction and is not pursued here, but for sites where the fracture statistics can truly represent the hydraulic behavior, statistical simulation may play a useful role in model development. Inversion and simulation are not mutually exclusive: inversions can be designed on the basis of the parameters of the statistical simulation and thereby yield a model that honors both the statistical data and the observed hydrologic behavior (Kitanidis and Voris, 1983).

6.1. Theoretical Concerns

The goal in creating the conceptual model for the inverse technique, simulated annealing, is to include hydrologic conductors in all the places that they are likely to be. Theoretically, it is
not a problem to include more conductors than are actually active, although having a large number of conductors may create practical problems with computation. However, omitting significant flow paths can create problems because simulated annealing will pick a subset of the possible conductors that can best explain the hydrologic responses observed in the rock. So our aim in creating the hydrologic conceptual model is to specify a set of conductors that represent all the important flow paths.

In a site where fracture zones are distinct and clearly dominant, the conceptual model might be defined as a set of planes or slabs representing each of the dominant fracture zones. Within each of these planes or slabs, the conceptual model specifies a lattice of conductors that form the basis of an equivalent discontinuum model. The form of the lattice might reflect flow paths indicated by the morphology of the fracture zone. For example, the K-zones at Grimsel might be assigned a lattice with dominant vertical channels reflecting the steps in the zone. If the remaining rock is expected to contain some significant but second-order flow paths, these might be represented by a background grid of conductors. If there is absolutely nothing known about a fracture network, one might use a conceptual model that simply consisted of a three-dimensional grid of conductors. As a matter of practicality, the conceptual model should be as efficient as possible in representing all the important elements of the system.

6.2. An Example from the Stripa Mine

As an example, the definition of a conceptual model is described below for the Site Characterization and Validation (SCV) block of rock in the Stripa Mine in Sweden (Long et al., 1991a). The SCV conceptual model described here was a preliminary model designed as a platform for simulated annealing and based on a characterization effort documented in Olsson et al. (1988a). The model is preliminary because it was based on the first cycle of data gathering at the SCV site. A revised and improved conceptual model was later developed for this site on the basis of a second cycle of data collection that allowed for a more consistent interpretation (Black et al., 1991; and Long et al., 1991c). The preliminary conceptual model is discussed below, and comments are given on the changes that were indicated by the second cycle of data.
Extensive geophysical data was collected on the SCV block through five boreholes (W1, W2, N2, N3, and N4), as shown in Figure 26. Radar and seismic tomography and reflections were integrated using the methodology described in Section 3.1, and four major fracture zones were identified, A, B, H, and I, and one minor zone, C (Olsson et al., 1988a). The hydrologic response was not taken into account in choosing these zones.

Single-hole packer tests, geophysical logs, and fracture frequency are also available for each of the holes. Figure 27 (after Olsson et al., 1988a) gives an example of the summarized borehole data and also shows the geophysically defined fracture zones as horizontal bands. The figure shows that fracture frequency does not correlate perfectly with fracture zone location. This is not surprising, considering that the structure of a fracture zone may be heterogeneous, for example, a stepped structure like the K-zones described in Section 2. Under the hydraulic conductivity column, those conductivities greater than $10^{-8}$ m/s are shown in solid black. It is also possible to have a hydrologically active fracture zone that does not present a permeability anomaly because the borehole penetrates the zone at a point where it is not permeable. On the other hand, it is possible to measure a high permeability where there is no major fracture zone or where the fracture zone is not connected to other fracture zones. Table 1, based on these data, gives a qualitative classification of the strength (S: strong; M: medium; and W: weak) of the geophysical and hydrologic anomalies associated with the zones. Table 1 also includes zone B', which is discussed below. The data are not completely consistent, so the process of proposing a conceptual model is one of trying to make the most consistent case possible while allowing for all reasonably probable hydraulic connections.

The goal is to have a conceptual model that can account for all the significant hydrologic responses observed at the site. The first step is to see how much of the hydrologic response can be accounted for by the geophysically defined fracture zones. If these zones are taken to have a width of about 10 m, they account for about 60% of the hydraulic transmissivity measured in the boreholes. Almost all of the remaining 40% of the transmissivity is accounted for in three locations: near 80 m in borehole W2, 152 m in N2, and from 80 to 90 m in borehole N4. There are no
Figure 26. Perspective view of the SCV block. Dotted area in the upper left represents the mined-out stopes (after J. Gale in Black et al., 1991).
Summary data sheet for borehole N2. Hydraulic conductivities greater than $10^{-8}$ m/s are shown in solid black. Hydrologic zones are marked in the right-hand column (after Olsson et al., 1988a).
identified fracture zones in these locations, but there are some nearby tomographic anomalies as well as anomalies in the resistivity, sonic, neutron, and core logs.

Table 1. Geophysical fracture zones

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<th>Zone</th>
<th>Radar</th>
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<th>Geophys. Logging</th>
<th>Core Logging</th>
<th>Hydrologic</th>
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An attempt was made to account for these hydrologic anomalies by altering the orientations of the major zones by an amount within the limits of resolution of the geophysics. After determining that no improvement was possible, zone B' was added to explain previously unaccounted for hydrologic anomalies in N4 and N2. Figure 28 shows a perspective plot as viewed along a plane parallel to zones B and C. In this figure, each zone appears as a line of dots. From this perspective, one can see that the hydrologic anomalies in N4 and N2 lie on a plane roughly halfway between zone B and zone C. To account for these anomalies, B' was chosen to be a plane between B and C, approximately parallel to zones A, B, and C. This choice of orientation was consistent with the style of the other fracture zones. The addition of B' increases to about 78% the percentage of transmissivity accounted for.

Zone B' fits well with some of the geophysical results. It is located where one initial radar interpretation put zone B ("RB" on Figure 29, from Olsson et al., 1988a). Later, the integration of radar and seismic results put zone B farther north, probably because the same anomaly is not apparent in any of the attenuation tomograms or the section N2-N4 slowness tomograms.

No hydrologic anomaly can be seen where B' intersects N3 (between N4 and N2). However, this can easily be accounted for if no permeable channel from B' intersects N3. It is more important to ensure that channels are possible where hydrology has been observed. Extra channels can always be made inactive in the annealing process, described in Section 7.
Figure 28. A perspective view of the SCV block looking up and to the northeast showing zones B and C and the two hydrologic anomalies situated between these zones in a plane parallel to them.
Figure 29. Residual radar slowness tomogram for the borehole section N3-N4 made with a center frequency of 22 MHz (from Olsson et al., 1988a).
One remaining part of W2 between zones H and B accounts for 21.7% of the transmissivity (not shown). If this transmissivity is allocated partly to H and partly to B, then 98.7% of the observed transmissivity is accounted for with a zone model. It is easy to imagine that the high conductivity found between zones H and B in W2 is due to a few conductive features that are related to H and B and possibly related to the intersection of H and B.

The resulting hydrologic zone model is shown in Figure 30 in a perspective view from the northwest. The planes of the zones are defined by discs. The next step is to assign a template lattice to each zone. The zones are not expected to be uniformly permeable, and the lattices will be configured by the inverse method described in the next section. The choice of grid is made with the support of geologic investigations of the fracture zones, as explained below.

The fracture zones have been studied by Martel (1991) and appear to be ancient faults. Not all the zones are exposed, but those that are exposed are relatively narrow and show evidence of repeated episodes of fracturing and faulting. Fractures of numerous orientations occur. Consequently, the zones are modeled as planes, and the planes are discretized using a square grid of conductors to form a generally isotropic network for the possible paths for fluid flow.

It is important to recognize that other configurations of conductors could have been proposed to account for the zone B' hydrologic anomaly and the anomaly in W2. A conceptual model is not unique; continuing characterization efforts will either build confidence in the model or change it. The second cycle of data collection that took place at the SCV included drilling more boreholes; more borehole and cross-hole geophysics; and interference testing. Evidence was collected that supported some of the zones, negated others, and pointed toward new ones (Black et al., 1991). In the new model, zones H and B were confirmed as the major features of the site, but zone B was moved slightly such that the zone B' anomaly in N2 was incorporated into the new zone B. Interference testing showed that the zone B' anomaly in N4 was disconnected from the SCV flow system and probably represents the end of a fracture zone that leads outside of the SCV block. This example shows the power of interference testing over single-hole packer tests. The single-hole tests of the first data cycle were limited to finding high permeability without any
Figure 30. The hydrologic zone model shown in perspective from the northwest looking down. Zones A, B, B', C, Ha, Hb, and I are shown. Gridding on the planes represents the hydraulic conductors of the template used for annealing.
information about how it was connected to the system. Finally, zone C was abandoned and several new minor zones identified.

More data usually results in a change in the conceptual model. However, what was probably one of the most significant results of the SCV work was that the second round of data collection did not change the conceptual model very much. In other words, it is possible for the process of characterization to converge on a conceptual model. In fact, iteration between data gathering and analysis is the only way to gain confidence in a conceptual model.
7. SIMULATED ANNEALING

An inversion technique called "simulated annealing" can be used to construct equivalent discontinuum models for fracture flow (Mauldon et al., 1991). Simulated annealing is an algorithm that is used to find lattice configurations that are functionally equivalent to the observed system: i.e., a model that simulates the observed behavior. The fracture network model is "annealed" by iteratively modifying the base model, or template, until one is found that behaves sufficiently like the observed system. This technique differs from inversion models developed in the past, such as the conjugate gradient method, or maximum likelihood method (Carrera and Neuman, 1986), because annealing is designed to completely turn off the conductivity of a portion of the region in order to determine how the conductive features are connected.

To search for patterns of conductors that behave like the observed field system, different configurations of the template are examined sequentially. In each configuration, some of the elements are "turned off," i.e., they become nonconducting. For each configuration, a well test that was also conducted in the field is simulated. The "energy" of the configuration is then defined as a function of the squared difference between the observed and the simulated responses. The problem of finding the appropriate model now becomes one of finding configurations that have low values of the energy function.

The algorithm starts from some arbitrarily selected configuration of conductors, C, and computes the energy. Then an alternative configuration, C', is selected by randomly choosing one of the elements and turning it off or on. The energy for this configuration is computed. If the new energy is lower than the energy for the old configuration, then the alternative matches the observed data better and the algorithm will decide to move to the alternative configuration. This is analogous to a downhill step. Random "uphill steps" to alternatives with higher energy can also be taken. An uphill step will be chosen with a probability that decreases with the amount of increase in energy incurred by the step. The probability of accepting the unfavorable step also
depends on a weighting parameter called the "temperature," in analogy to metal annealing. The temperature, T, is decreased as the number of iterations increases to make it more and more unlikely that an unfavorable change will be accepted (Metropolis et al., 1953).

Maulden et al. (1991) have proposed energy functions, Q, for fracture models, based on Kilpatrick et al. (1983) and Tarantola (1987):

\[ Q = \sum_j (o_j - s_j)^2 \]

where

- \( o_j \) = a vector of observed responses, and
- \( s_j \) = a vector of simulated responses.

The observed measurements could be hydrologic, geological, or geophysical. For example:

\[ Q = \sum_t \sum_j (h_{oj}(t) - h_{sj}(t))^2 \]

where \( h_{oj}(t) \) is the observed steady-state head response at well \( j \) and time \( t \) and \( h_{sj}(t) \) is the simulated steady-state head response at well \( j \) and time \( t \).

At each iteration \( k \), given \( C \), \( G_C \), the neighborhood of \( C \), and \( T \), the temperature, a matrix of transition probabilities can be defined. The probability of moving from configuration \( C \) to \( C' \), given the current configuration \( C \) is equal to the probability of selecting \( C' \) to compare with \( C \), multiplied by the probability that the system would make the transition to a given \( C' \). That is,

\[
P(C \rightarrow C' | C) =
\begin{cases}
0 & \text{if } C' \neq G_C \\
P(C' | C) & \text{if } C' \in G_C, C' \neq C \\
\frac{Q(C') - Q(C)}{T} & \text{if } C' \in G_C, C' \neq C \\
P(C' | C) \cdot e^{-\frac{Q(C') - Q(C)}{T}} & \text{if } C' \in G_C, C' \neq C
\end{cases}
\]

The temperature schedule is used to lower the temperature as annealing progresses. Hajek (1988) shows that a temperature schedule that is inversely proportional to the log of the iteration number will converge in probability to a set of minimum energy states. Maulden et al. (1991) hold that Hajek's temperature schedule is overly constraining; following the suggestion of Press.
et al. (1986), they decrease the temperature whenever 50 changes have been accepted at the current temperature. Each interval of the schedule with constant temperature is called a step. At the end of each iteration, $k$, the temperature, $T_k$, is decreased using a geometric series,

$$T_{k+1} = T_k t^k$$

where

$$0 < t < 1.$$  

Mauldon et al. (1991) also suggest that the initial temperature be chosen such that it is of the same order of magnitude as the energy difference between the first two configurations. This is done in an attempt to scale the energy difference between successive configurations between zero and one. This choice of temperature schedule does find low energy configurations. Other choices of temperature schedule are possible and are currently a topic of research.

7.1. Hydrologic Data for Inversion

One of the significant problems associated with applying these techniques is to choose a data set as the basis of the inversion. In principle, any physical phenomena of interest that can be numerically modeled and also monitored in the field can be used in the inverse method. In practice, it can be quite difficult to pick a good data set for analysis.

The simplest approach has been to use the steady-state head distribution resulting from a pumping test. The energy function is constructed as a function of the differences between modeled and measured heads or drawdowns. Drawdowns induced by such a test are relatively simple to measure, and steady flow is easy and quick to model, allowing many iterations of the model to be practical. However, for steady flow, the pattern of drawdowns does not change when conductance of the medium is scaled up or down. So using a single steady-flow test will only give a pattern of conductance contrasts that matches the head distribution. The value of these conductances can then be scaled up or down until the observed (or applied) flow conditions are matched. This means, not surprisingly, that the models obtained largely by matching drawdowns should be more sensitive predictors of drawdown than they are of flow.
Greater sensitivity to absolute conductance can be gained by combining a series of steady-flow tests. In this case, each of the separate tests is modeled at each iteration, and the factor incrementing the element conductances is chosen to best fit all the flow data. If constant head boundary conditions are used at the pumping well, the energy function can be constructed as an appropriately scaled combination of squared head differences and squared flow differences. If constant flow is applied at the pumping well, the energy function can include the head at the pumping well treated as any other observed head. Using multiple steady tests may actually be the best for the inversion, because there is no dependence on storage coefficient and the time required for steady flow calculations is very small. However, in the field each steady test is very time consuming and consequently few are usually available.

Alternatively, one can use the transient interference data from a constant flow test. For example, the flow rate used in the field can be specified in the model in order to predict the transient drawdown response. At each iteration, the model predicts curves of drawdown versus time that can be shifted in both the x- and y-directions in log-log space until a best match is obtained to the real curves. This process is similar to matching data to a Theis curve, but in this case the shift corresponds to scaling the conductances and storage coefficients for the lattice elements in the model. With several observation points, it becomes necessary to find the best shift on average. Although this process is conceptually simple, the vagaries of numerical calculation combined with the vagaries of real data can make curve matching extremely difficult to do automatically for thousands of iterations. The energy at each iteration is the sum of the squared differences in log of head for each observation point at selected times. The advantage of using this type of data in inversion is that the transients reflect the distribution of heterogeneities in space. A steady test is more likely to reflect the biggest bottleneck, irrespective of where it is. The disadvantage of transient data is that it is necessary to make an assumption about the relationship between storage and conductance; in other words, there is more information, but another parameter to specify.

A slightly different procedure must be used if the transient test is a constant head test. In this case both the transient drawdown data at the observation wells and the transient flow rate at
the pumping well should be used. Thus the energy will be a mix of log of squared flow differences and squared log of head differences, and it is necessary to decide how to weight these. Another nuance is that the y-shift is not correct for the head data but is correct for the flow data. This is because head is pegged by the constant head boundary condition. Consequently, constant head tests are somewhat more sensitive to the initial estimate of the element conductances and practically more difficult to use in inversion than constant flow.

If several different tests are available, these can be combined. In principle, any combination of steady, constant flow transients or constant head transients can be combined. The main drawback for combining a large number of transients is the possibility of using an enormous amount of computer time. The number of calculations needed is the number of tests times the number of time steps times the number of iterations, and it is not difficult to conceive of a problem that could take about a month to invert.

For multiple constant flow transients, the procedure is relatively straightforward. At each iteration, each test is modeled and the curves are all shifted in both x- and y-directions to get the best fit. Theoretically, a steady-flow case is a subset of a constant flow test, and the steady drawdowns predicted by the model can be matched to the data by a shift in the y-direction. The x-shift is irrelevant for steady-state conditions. However, once a constant head test is included in the inversion data, the y-shift cannot be used to match the drawdown data from the constant head test. One approach is not to use the y-shift on any of the drawdown curves. In this case, it is necessary to have a good a priori estimate of channel conductance and possibly to revise this estimate periodically during the inversion. Again, flow data should be included in the energy function for the constant head case.

7.2. A Synthetic Example

In order to see how the simulated annealing algorithm works, Mauldon et al. (1991) developed a series of synthetic cases. In these cases, the "real" system is completely known, so that the results of annealing can be evaluated absolutely to the steady-state data. Synthetic cases are generated using the fracture network generator FMG (Long et al., 1982, Long, 1983). FMG
produces random realizations of a population of one-dimensional conductors in a two-dimensional square region. An interference test is modeled on this network by creating a constant flux internal boundary at a centrally located node. The program TRINET (Karasaki, 1986) is then used to calculate the head response at a series of observations nodes. These responses become the “real” data that annealing tries to match.

A dimensionless network with two fracture sets was generated (Figure 31), and a template for annealing was developed using a grid with orientations close to those of the two fracture sets in the synthetic case (Figure 32). The annealing algorithm found a minimum energy solution that appears by eye to match the flow geometry well (Figure 33). Figure 34 shows the energy versus the iteration number for the annealing run.

Mauldon et al., 1991 also show that varying the percent of conducting elements in the initial configuration does not change the percentage of conducting elements in the final configuration. Another study shows that building a template with conducting elements oriented similarly to the real system appears to give better results. Peterson (pers. communication) found that annealing can find major “holes” when no well is located in the hole if a sufficient number of interference tests are available. A single interference test may not be enough. These studies are ongoing and can also be used to indicate which schemes for choosing configurations to test are best.

7.3. Steady-State Annealing: Application of Annealing to the MI Site

Annealing has now been applied at several fracture sites with encouraging results. Summarized here is one such application by Mauldon et al. (1991) at the MI site in the Grimsel Mine using steady-state data. Annealing was applied to cross-hole tests that were conducted in the plane of a fracture zone (shown as shaded planes in Figure 35) that intersects a drift. Called the MI fracture zone, this zone is an S-zone, described in Section 2, and is similar to the FRI fracture zone, described in Section 3.3. This means that the geologic investigations indicated that vertical permeability should be the most significant.

From the drift, eleven boreholes have been drilled into the plane of the fracture zone.
Figure 31. The synthetic case used to generate well test data for use in annealing. Dots represent points where "well" data were generated, the central one being the pumping well. Scale can be considered dimensionless.
Figure 32. An example template developed for annealing the synthetic well test data.
Figure 33. A configuration resulting from annealing the synthetic well test data.
Figure 34. The dimensionless energy versus iteration curve for the synthetic annealing case.
Figure 35. The layout of the MI experiment showing the plane of fracture zone that intersects the laboratory tunnel and the eleven wells drilled into the zone from the drift.
Several constant pressure tests and a constant injection test were conducted in the migration fracture while the pressure was measured at a number of observation points. An unmeasured amount of water flows from the zone into the drift, which is not sealed. A steady-state case for annealing was constructed by using the head distribution resulting from flow into the drift. The heads for this case were obtained from the end of a long recovery period after one of the injection tests (Figure 36).

Figure 37 shows the template that was developed for this site. A dense mesh is used in the vicinity of the wells, where it is expected that annealing will be able to resolve the pattern of conductance. A coarser mesh is used outside of this region in order to allow the numerical simulation to be insulated from the boundaries. The template includes vertical conductors in accordance with the geologic investigations. The hexagonal pattern is similar to the braided pattern observed for S-zones.

Figure 38 shows one of the six different results of annealing. In Figure 38, one can see regions where the inversion is predicting lack of connection. Also, annealing has found a lack of connection between the well at the extreme upper left and the outer boundary. Annealing is disconnecting this well because the measured head in the well was close to zero and the outer boundary is held at 100 m. As the drift boundary is zero head, steady-state annealing encourages what may be an anomalous connection to the drift rather than the more likely case that this well does not intersect the active fracture network. Transient annealing may be able to identify true lack of connection given sufficient different cross-hole data.

The steady-state results show that the data can easily be matched. Using the multiple solutions obtained for this case, Mauldon et al. (1991) were able to perform a cross-validation study, described in Section 9. The difference between the observed heads and those found numerically in each solution is very small. The annealing algorithm seems to "smear" the nearby measured flow response over regions with no data available. However, unlike Kriging, the nearby measurements are not linearly interpolated over these regions. The algorithm finds a random flow geometry that works, and this will vary in each solution.
Figure 36. Head records in the eleven wells during pumping of well 9 and subsequent recovery.
Figure 37. The MI template.
Figure 38. The first of five solutions with similar geometries, case 1.
7.4. Transient Annealing: An Example from the Stripa Mine

Transient co-annealing was applied to the H-zone at the Stripa mine in Sweden, which was described in Section 6. A series of seven wells (C1, C2, C3, C4, C5, W1, W2) penetrate this zone, and an interference test, called the C1-2 test, was conducted in these holes. Another experiment, called the "simulated drift experiment" (SDE), measured the flow rate into six additional holes (the D-holes), which also penetrate the H-zone and are drilled within a 2-m-diameter cylinder. The entire data set is described in Olsson et al. (1988a) and Black et al. (1991). An inversion of this data using Simulated Annealing is summarized from Long et al. (1991c).

An inversion based on the C1-2 cross-hole test is used to predict the flow rate in the SDE. First, a model is annealed to the C1-2 interference data. Then the C1-2 node in the model is closed and the equilibrium heads imposed by the boundary conditions are obtained. Next the head at the D-holes is set to create the estimated drawdown imposed by the SDE (220 m) and to calculate the flow into the D-holes.

A two-dimensional template was chosen to model the H-zone for this example on the basis of three major considerations. The template is configured to get as much detail as possible in the vicinity of the D-holes; to have a large enough mesh to prevent the transients from reaching the boundary too soon; and to keep the number of elements and bandwidth as small as possible to keep the annealing time as small as possible. The variable density mesh is in keeping with all of these considerations. There are 5 nested grid regions, each having twice the grid spacing of its inner neighbor. In addition, 200-m-long elements connect each of the nodes lying on the outer edge of the outer grid region to the applied boundaries. This allows us to have a 1.5-m spacing grid in the vicinity of the D-holes, applies the boundaries approximately 400 m from the pumping wells, and keeps the total number of elements down to 4687.

The element conductance and storativity are scaled such that the whole region has the same average transmissivity and average storage. Each of the well intervals is included as a node in the mesh. The behavior of the full mesh was checked by running the C1-2 test and examining the drawdown curves. These were smooth, showing no evidence of transition from one grid region to
The boundary conditions in the model were chosen to get a reasonable match to the estimated equilibrium head values. Outer boundary conditions were picked that were within reason while allowing drainage into the Z-shaft, which intersects the H-zone, below the D-holes, about 150 m away. A node fixed at -75 m head is added at the location of the Z-shaft to provide this sink.

Figure 39 shows the two-dimensional template annealed to the Cl-2 test, and Figure 40 shows the match between the observed well responses and the model responses. Figure 41 shows the energy versus iteration curve. Annealing was able to match all of the data quite well. The boundary conditions were then changed to represent the conditions at the end of the SDE experiment. This model predicted that the flow rate into the D-holes should be 0.77 l/min, which matched exactly the measured flow rate into the D-holes. Despite this excellent result, to use this method correctly, the prediction should really be based on several inversions, as shown in Section 9.

7.5. Conclusions

Although transient response of a system is more sensitive to the distribution of permeability than the steady-state response, Davey et al. (1990) have found difficulty in matching transient well test data in the case where there is a large variation in conductivity as well as interconnectivity. This indicates that annealing with a discrete range of conductances rather than a binary distribution may be necessary in some cases.

These inverse problems have many acceptable solutions with different flow geometries. If the same generalized geometry is found in a certain region for many solutions, it is more likely to be real. However, in some cases expert opinion indicates that an aspect of the flow geometry is an artifact of the process. An example of such an artifact may be the steady-state solution of the MI fracture that disconnected a well from the boundary.

In general, the inversion of well test data yields a non-unique solution for the fundamental
2-D Mesh co-annealed to C1-2 and SDE
(Dead-end elements dotted)

Figure 39. The two-dimensional model annealed to C1-2.
Figure 40. Comparison of C1-2 well test response data to model results.
Energy Vs. Iterations of annealing
C1-2 pump test [2-D Case]

Minimum energy 0.393296 at iteration 2867
Maximum iteration 2867

Figure 41. The energy vs iteration curve and temperature schedule for the two-dimensional model annealed to the C1-2 test.
reason that there is rarely, if ever, enough data to specify a unique solution. The advantage of the inversion technique described here is that it can be applied to find several solutions, all of which are equally likely. Furthermore, a single well test can be used to predict a second well test, and the first two well tests can be used to predict the third. In this way it is possible to see if the predictions improve, which implies an improvement in the uniqueness of the solution. The best way to assess the utility of the model is by estimating the error associated with the model predictions, as discussed in Section 9.
8. APPROACHES THAT INCORPORATE Scaling

Annealing is not the only possible way to find networks of fractures that honor the hydrologic data. Another approach that holds promise is related to the fact that fracture networks may exhibit scaling behavior. It may be possible to find objects that exhibit self-similar properties and also behave like the well test observations. This approach has tremendous appeal for the simple reason that the self-similar properties provide a logical path for scaling up our understanding to larger regions.

It is not hard to believe that some fracture networks might be described by fractals. In fact, the name "fractal" is derived from the word "fracture." Barton et al. (1987) have described fractal properties for fracture outcrops. Hestir and Long (1990) have shown that fluid flow in a fracture network is equivalent to the problem of percolation on a lattice, in which case the network can be characterized by clusters of conductors that form at scales that exhibit self-similar geometry (Orbach, 1986). Fractal representation of fracture flow geometry may provide another way to find equivalent discontinuum models.

Flow to wells in such geometry differs from that in Euclidian geometry. Barker (1988) has provided a technique for determining the fractional flow dimension of a network through a well test. He solved the generalized equation of flow to a well by letting the flow dimension be a variable. Thereby, the flow dimension is allowed to be fractional, say, a dimension of 1.6 or 1.8 (as opposed to integral; i.e.; two- or three-dimensional space). Polek (1990) has shown that this "flow" dimension is closely related to the geometric fractal dimension. A fractal description of a fracture network may be a good basis for inversion, because interference data contain the signature of the fractal dimension.

Fractals are a subset of a larger class of objects, called "attractors," that can be generated with an "iterated function systems" (IFS). Some new ideas for inverse techniques based on IFS are described below.
8.1 Iterated Function Systems

Iterated function systems have become a standard tool for modeling self-similar geometrical structures (Barnsley, 1988); they were originally developed for use in computer graphics in order to find efficient means for storing the information describing each pixel of a complex picture. In this application, one identifies an iterative process that will create the picture rather than storing the information for each pixel. The iterative process is defined by an IFS, which has a relatively small number of parameters. The use of an IFS essentially exchanges the use of computer storage for the use of computer time. In this application, an IFS is used to create an equivalent model of a complex fracture system. Instead of trying to describe this system "pixel by pixel," an IFS is found that can describe the complex geometry of the system with a small number of parameters.

An IFS creates a picture starting with an initial set of points and a set of iterative functions. At each iteration, each function in the system operates on the set of points and, according to the parameters in the function, translates, reflects, rotates, contracts, or distorts the set of points. After many iterations, the points in the picture coalesce toward an "attractor," which may be a fractal object. The shape of this attractor changes gradually when the parameters of the IFS gradually change.

To create an IFS, one first specifies a function $f$ that maps sets to sets:

$$f(A_0) = A_1$$

where $A_0$ and $A_1$ are (compact) subsets of two- (or three-) dimensional space. A set $A_\infty$ can then be defined by

$$A_{n+1} = f(A_n) \quad n = 0, 1, \ldots$$

$$A_\infty = \lim_{n \to \infty} A_n .$$

Given certain restrictions on the set function $f$, one can show (Barnsley, 1988) that $A_\infty$ exists, is independent of the starting set $A_0$, and generally has a fractional Hausdorff dimension. Hence $f$ determines a fractal, $A_\infty$. If there is a function $f$ that is easily parameterized, then the fractal $A_\infty$ is parameterized as well. In this way a small number of parameters can be used to characterize a complex geometry.
A wide variety of iterated function systems can be defined, but they fall into two main categories: deterministic and probabilistic. A deterministic IFS has uniquely determined parameters and thus creates a unique attractor $A_\infty$. A random IFS chooses some or all of its parameters randomly from probability distributions, so that multiple realizations of $A_\infty$ differ. The iterated function systems used in the hydrologic inversions given here are deterministic and of the form given above; those used in the fracture growth scheme discussed in Section 8.4 are random.

One important example of a deterministic IFS used extensively by Barnsley (1988) is

$$f(A) = g_1(A) \cup g_2(A) \cup \ldots \cup g_k(A).$$

Here the $g_i$'s are so called affine transforms:

$$g_i(A) = \bigcup_{x \in A} g_i(x)$$

$$g_i(x) = B_i x + b_i$$

where $B_i$ is a matrix and $b_i$ a vector. The parameters characterizing $f$ are the entries in the $B_i$'s and $b_i$'s. The matrix, $B_i$, serves to rotate, reflect, distort, and contract and the vector, $b_i$, translates.

An example IFS using $k = 3$ affine transformations that contract and translate, resulting in a fractal called a Sierpinski's gasket, is shown in Figure 42. The IFS is specified by

$$B_1 = B_2 = B_3 = \begin{bmatrix} 0.5 & 0.0 \\ 0.0 & 0.5 \end{bmatrix},$$

$$b_1 = (0.0, 0.0), \quad b_2 = (0.5, 0.0), \quad b_3 = (0.0, 0.5).$$

Figure 43 shows the attractors generated by a sequence of functions $f_1, f_2, \ldots, f_6$, where $f_1$ is the Sierpinski's gasket, and for $j = 2, 6$ every parameter of $f_j$ differs from the corresponding parameter of $f_{j-1}$ by a small increment. The continuous change in parameters is manifested as a continuous change in the attractors, which is a useful but not necessary condition for an IFS-based inversion procedure to work.

### 8.2 Inversion Based on Iterated Function Systems

To use an IFS as a basis for hydrologic inversions, the points of the attractor are mapped into hydrologic parameters (conductivity, storativity, etc.) on a lattice, and use an inversion algo-
Figure 42. Generation of a Sierpinski’s gasket using the affine transformations given in the lower right figure.
Figure 43. A series of attractors generated by functions whose parameters differ by small increments.
algorithm to find parameters of the IFS. An attractor is superimposed on the lattice, and the conductance and storativity of the lattice elements that are close to each point on the attractor are incremented as shown in Figure 44. The conductance of a lattice element can be incremented as many times as there are points on the nearby attractor. In this way the small number of parameters of an IFS define a broad conductance and storativity distribution in thousands of elements.

The inversion algorithm searches for IFS parameters that define a heterogeneous system that behaves like the observed well tests. First, a model of the flow system is constructed using a lattice of elements modified by an arbitrary IFS. Then, the parameters of the IFS are optimized until the model produces a good match to the well test data. As in simulated annealing, the match is quantified by the energy, $E$, which represents, in a single number, the total amount of mismatch between the observed and modeled drawdowns, and is a convenient way of quantifying the closeness of fit of the model to the data during the course of an inversion.

The optimization can be done in a variety of ways. Several routines available in standard numerical libraries have been used (Long et al., 1991b), including downhill simplex and direction set methods (Press et al., 1986). One optimization technique that seems to work well is simulated annealing. In this case, new values of the IFS parameters are chosen randomly and accepted or rejected according to the annealing algorithm, as described above.

Some parts of the inversion algorithm are arbitrary. For example, we choose the number of affine transforms, $k$, that make up the IFS. The number of points, $M$, for the IFS to use in creating the attractor is chosen on the basis of experience. The larger $M$ is, the greater the contrast in permeability can be. One could use a high value of $M$ to model highly conductive features in a relatively impermeable matrix or a lower value of $M$ to model conductive features in a slightly impermeable matrix. Further, we also arbitrarily choose how to relate the increment in conductance and storativity represented by each point of the attractor. Moreover, the way in which the points of the attractor increment conductivity and storativity is arbitrary. The effect of these arbitrary choices can be examined by trying different choices.
Figure 44. "Step" mapping between points on the attractor and increments in hydrologic properties of the lattice.
One of the attractive features of this approach is that it may be possible to choose subclasses of iterated function systems which tend to produce features observed in a geologic investigation. For example, a class of IFS might be found that always produces a specific type of brittle shear zone, for example, a K- or S-zone, described in Section 2. In these cases the search for hydrologic behavior could be confined to the subclass of IFS that is used to represent the geology. Along the same lines, once the form of the IFS has been identified that best explains all the data, the model will have fractal-like properties that may help to extrapolate behavior to scales that cannot be tested in situ.

8.3 Example Inversion of the Stripa H-zone Data

A simulated annealing inversion of the H-zone data at Stripa was presented in Section 7.3. Presented here is a preliminary IFS inversion based on the C1-2 cross-hole test (Long et al., 1991b). The IFS inversion uses the same template and the same boundary conditions as the simulated annealing inversion. The IFS inversion is also used to predict the flow rate in the SDE in the same way as for the simulated annealing inversion.

The inversion was started using three affine transformations in which each point on the attractor incremented both the conductance and the storativity of the nearest element. Figure 45 shows the attractor obtained at the 799th iteration, where the energy had dropped to about $E = 13$ from an initial value of about $E = 45$; a better solution could be found by continuing the process.

We again make a prediction of the flow rate for the SDE by applying a constant head boundary at the D-holes such that the same estimated drawdown at the D-holes (220 m) that was imposed during the SDE is imposed in the model. The actual flow rate to the D-holes from the H-zone during the SDE was estimated to be about 0.7 l/min. Our calculation gives 0.4 l/min which is low, but reasonably close.

One interesting attribute of this inversion is that the attractor first resided entirely in the upper left-hand corner of the mesh and consistently migrated to the lower right with each iteration. Thus the conductance in the vicinity of the C- and D-holes is consistently increasing, which
Figure 45. The attractor found at iteration 799 for the central 200 x 200 m section of the mesh. The six D-holes are in the immediate vicinity of d2.
means that the predicted SDE flow rate is increasing with each iteration. An attractor that produces too low a conductance near the D-holes is consistent with the fact that imposing a 220-m head drop at the D-holes would require setting the head boundary at the D-holes to a value that is unreasonably low (−137 m compared to the measured value of −8 m). Clearly, it is possible that further iterations will continue to move the attractor down and significantly improve the solution. However, another possibility is that there are too many points in the attractor, which results in too high a contrast. This Stripa example is preliminary, but it does demonstrate that the concept can be applied.

8.4 Iterated Functions to Describe Fracture Patterns

One exciting possibility for IFS inversion is that it may be possible to condition the hydrologic inversion on geologic information. If an IFS can be found that reproduces the geometry of a geologic system, then inversion searches could be restricted to this class of functions. Probably the best way to find such classes of iterated functions is to base the functions on an understanding of how the system in question develops. For example, the functions could reflect the growth mechanics of the joints. A preliminary example of such a description of joint growth is summarized from Hestir et al. (1990).

Hestir et al. (1990) generated sequences of fracture patterns that have self-similar properties as well as the complex geometries observed in the field with a first-order growth scheme. To build a fracture pattern with this scheme, a beginning set \( A_0 \) is defined to be a given existing set of fractures in a mostly unfractured rock. The function \( f \) applied to a set \( A \) of fractures is defined to be a rule that grows new fractures from each of the existing fractures in \( A \). This is done by looking at each fracture in \( A \) in its own local coordinate system (Figure 46a, the solid line) and growing one fracture from it using rules defined in that local coordinate system. The method is called first order because new fractures are grown without accounting for interaction between existing fractures.

The new fracture growth is chosen at random from a finite number of possibilities (the dashed lines in Figure 46a), each with a given probability of occurring. The growth possibilities
(with probability $p_1 p_2 \ldots$) shown in Figure 46a result in the fracture pattern given in Figure 46b. The rules governing fracture growth can be based on fracture mechanics. For example, the probabilities for growth are scaled to the size of the fracture, and the growth positions are located approximately where stress concentrations would be in the absence of interactions. Further, growth orientations can be picked on the basis of stress trajectories. This method has also been used to produce realistic looking sequences of two-dimensional joint patterns (Long et al., 1991b).
Figure 46. (a) A fracture (solid line) shown in its own coordinate system has the possibility of growing according to the dashed lines at each iteration. (b) Application of this growth scheme results in this fracture pattern.
9. QUANTIFICATION OF UNCERTAINTY IN MODEL PREDICTIONS

Two important types of uncertainties affect model predictions: 1) uncertainty in the data, which can be due to measurement error or simply to lack of measurement, and 2) uncertainty in the conceptual model, i.e., uncertainty in the underlying assumptions about geometry or physical processes. Both types of uncertainty are inherent in predictive models in the earth sciences.

There is never enough data and the conceptual model can always be questioned. Consequently, it is not possible to construct a unique model of fracture flow. Given this situation, it is probably better to view the data as determining a series of possible models from which a series of predictions can be made. For an equivalent discontinuum model, there are several ways to choose different configurations of conductors to compare. The simplest is to use a series of configurations defined near the end of an annealing process. These configurations are easily available, but they will probably be very similar to each other. Another way to find different configurations is a Monte Carlo approach where annealing is performed several times, each time following a different random path leading to a different final configuration.

With a series of models, it is possible to quantify the error associated with using the models to make a prediction. In this approach, predictions are made with the model and checked against subsequent measurement. This means that the model is judged by its ability to predict the system response accurately. "Prediction error" so defined is a lump measure of error caused by incomplete data and model assumptions. The calculation of prediction error is made by using the model to make a series of different predictions. For each prediction, prediction error is obtained by comparing the calculated result with the measured result. The root mean square of these errors is called the prediction error.

Calculation of prediction error is straightforward. The more numerous and diverse types of predictions that can be included in the estimate of prediction error, the more confidence is gen-
erated in the model. If the model works well to predict flow under one set of boundary conditions, it may not predict well for a different set of boundary conditions. However, if it predicts well for two sets, then it is more likely to predict well for three, even better for four, etc. This is the essence of the iterative approach: cycles of measurement and prediction are repeated until the addition of new data does little for the ability to make predictions.

A limitation of this approach is that there are many cases where extensive in-situ testing is not possible. Thus good statistical samples of prediction error are usually not available. Further, the use of the models must often be extended to classes of physical conditions, phenomena, or time scales that have not been tested in the field. A model that works for one case is not necessarily a valid model for radically different boundary conditions or different phenomena. The only way to quantify uncertainty under these conditions may be to find bounding cases.

9.1. Prediction Error

The best possible way to evaluate prediction error is to make a prediction for a known quantity that has not been used to build the model. Unfortunately, one usually needs all the available data to build a good model. One way around this problem is to set aside one data point, construct a model using the rest of the data, predict the value left out, and calculate a prediction error. If this is done for each data point in turn, a distribution of prediction errors will be generated that can be used to estimate the prediction error for a model using all the data. This process is called cross-validation. Cross-validation may be extended to calculate multiple solutions for each data point set aside. For example, in the MI study discussed in Section 7, steady-state head values are available at 8 wells. If one well is left out at a time for each annealing, a range of prediction errors for pressure measurements can be calculated. This range of errors can be used to estimate the prediction error for the full model.

When multiple solutions are available, cross-validation can be used to choose a function of these solutions that is a good predictor. Suppose there are a number of annealing solutions used to predict each measured data point value in a cross-validation study. Cross-validation can determine whether the mean or median of these predictions gives a lower prediction error.
9.2. Example Cross-Validation from the Migration Site

Mauldon et al. (1991) carried out a cross-validation study with the MI data (Section 7) to choose a good predictor for head and to estimate the prediction error. At the MI site, eight wells and one drift intersect the permeable fracture zone. The steady-state observed pressure response, $H_{\text{obs}}^i$, was calculated at each well, $i$, under conditions of steady flow to the drift. Mauldon et al. (1991) calculated the prediction error associated with using the annealing model to predict the steady-state head response at a ninth nearby well by doing the following:

1. Leave the steady-state head value for well $i$ out of the energy function.
2. Find five different annealing configurations, cases 1 through case 5.
3. For each end configuration, calculate a predicted steady-state value for well $i$. These predicted head values are $H_1^i, \ldots, H_5^i$.
4. Calculate the mean squared prediction error for well $i$:

$$\text{PE}^2(i) = \frac{1}{5} \sum_{j=1}^{5} |H_j^i - H_{\text{obs}}^i|$$

The estimated prediction error is

$$\hat{\text{PE}} = \left[ \frac{1}{8} \sum_{i=1}^{8} \text{PE}(i) \right]$$

$\hat{\text{PE}}$ is then an estimate of the error involved in using one annealing model to predict the head response of any other well in the vicinity.

$\hat{\text{PE}}$ was computed for the mean or median predicted head response of the five solutions for each well left out. The median proved to be a slightly better predictor of steady-state pressure at a given point on the grid. In addition, $\hat{\text{PE}}$ for predictions made using each solution independently was larger than $\hat{\text{PE}}$ for predictions made by generating five solutions and using the median value as the prediction. The estimated prediction error found using a single solution was 4.3 m, and the estimated prediction error for using the median of five solutions was 3.3 m (Table 2). The predic-
tion error for well 11 (the well in the upper left corner of Figure 38) was very large and tends to have a big effect on the prediction error (see Table 2). If well 11 is ignored, the estimated prediction error using the median is 2.3 m. The estimated density of prediction error for the median of five solutions also shows that the median is expected to give a lower prediction error. Therefore, better predictions can be made if they are based on multiple annealing solutions instead of a single solution.

Table 2. The observed steady-state head values at each well and the predicted head values found using the median value for five annealing solutions. In each case the steady-state head at the indicated well was left out of the energy function.

<table>
<thead>
<tr>
<th>Well Left Out</th>
<th>Obs. Head</th>
<th>Median (h) (m)</th>
<th>PE (i) (m)</th>
<th>PE (i,h) (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>4</td>
<td>9.97</td>
<td>6.64</td>
<td>3.8</td>
<td>3.3</td>
</tr>
<tr>
<td>5</td>
<td>10.95</td>
<td>5.95</td>
<td>5.0</td>
<td>5.0</td>
</tr>
<tr>
<td>6</td>
<td>10.22</td>
<td>7.72</td>
<td>4.0</td>
<td>2.5</td>
</tr>
<tr>
<td>7</td>
<td>0.64</td>
<td>0.988</td>
<td>3.2</td>
<td>0.3</td>
</tr>
<tr>
<td>8</td>
<td>3.37</td>
<td>0.96</td>
<td>2.5</td>
<td>2.4</td>
</tr>
<tr>
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<td>8.07</td>
<td>9.99</td>
<td>2.1</td>
<td>1.9</td>
</tr>
<tr>
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<td>2.8</td>
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</tr>
<tr>
<td>11</td>
<td>1.04</td>
<td>11.37</td>
<td>11.1</td>
<td>10.3</td>
</tr>
</tbody>
</table>
10. CONCLUSIONS

This chapter represents a compilation of a significant amount of research, which took place over several years and involved many people. Even so, it is by no means a complete reference. A line of reasoning has been presented that could be applied at a fractured site to deduce a model for the hydrologic behavior of the system. The main points of this approach are (1) that interdisciplin ary interaction is a critical part of maximizing understanding and reducing uncertainty and (2) that an equivalent discontinuum model for fracture flow based on behavior and constructed from the top down may be an appropriate approach for modeling some sites. This chapter has attempted to describe how the elements of different efforts can be linked and related to a final product.

Clearly, this is not a completely solved problem, and there is much more to be done. Probably the most important progress will be made simply by trying to create predictive models for an increasing number of sites. Only in this way can what works be distinguished from what does not: hydrogeology is in many ways a heuristic science. There is a lot of experience in porous materials and as such there are theories that are useful for these cases. To know that the models developed for fractured rock are valid, it is simply necessary to use these models to make predictions and see how well they do.
11. REFERENCES


