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METASOMATISM AND PERMEABILITY

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ABSTRACT. Permeability, a fundamental rock property of hydrothermal systems, controls metasomatic alteration processes. It not only controls the distribution and abundance of the chemical and mineral alteration but is in turn altered by the metasomatism. However, permeability cannot be directly measured on the fossil remains of a geologic system but must be estimated by indirect methods. One of the more promising methods is based on the mineralogy and abundance of metasomatic alteration in a rock and can be derived from transport theory.

Metasomatic alteration is caused by the flow of fluid from one chemical environment into another along percolation networks active during the metasomatism. The driving mechanism for this process can be related to a shift in chemical affinity caused by advective fluid flow. The magnitude of fluid flow in regions around magma-related thermal anomalies is directly proportional to the driving force and the permeability of the rock through which it flows. The field of force that drives these fluid flows can be estimated accurately from data on the geometry of the causative thermal anomaly and from equation-of-state for the fluid. Numerical analyses reveal that the buoyancy force generated by purely thermal effects is similar for all pluton environments, and that only the permeability of the percolation network varies significantly from system to system. As a consequence of this global similarity a theoretical relation is developed that can be used potentially for estimating the permeability for a broad spectrum of fossil hydrothermal systems.

The estimation of permeability from the metasomatic changes induced by the flow of fluids through the percolation network in turn provides insight into other mechanisms involved in metasomatism. However, general applications of the theory still pose many challenges to modern field mapping techniques and to the techniques of integration of field data into numerical analyses.

SYMBOLS

$\mathbf{g}$	— gravity vector, [cm s^{-2}]
$\hat{g}$	— magnitude of gravity vector, [cm s^{-2}]
k	— permeability, [cm^2]
G	— arbitrary system variable, [units]
$\mathcal{G}$	— concentration of system variable, [units cm^{-3}]
$\hat{I}$	— total number of minerals
$p = 1, 2, \ldots, \hat{P}$	— product mineral index
P	— fluid pressure, [bars]
p	— perturbation fluid pressure, [bars]
T	— temperature, [°C]
$r = 1, 2, \ldots, \hat{R}$	— reactant mineral index

$i = 1, 2, \ldots, \hat{P}, \hat{P} + 1, \ldots, \hat{I}$ — total minerals index
$\mathbf{H}$ — fluid force vector, [dynes cm^{-3}]
$\mathbf{v}_f$ — Darcy velocity of the fluid phase, [cm s^{-1}]
v_l — Darcy velocity of fluid along path l, [cm s^{-1}]
$\mathbf{i}, \mathbf{j}, \mathbf{k}$ — unit vectors
v_x, v_y, v_z — cartesian components of the velocity vector, [cm s^{-1}]
δ — finite difference operator
ϕ_f — volume fraction of the fluid phase
ϕ_r — volume fraction of reactant mineral
ϕ_p — volume fraction of product mineral
μ — dynamic fluid viscosity, [g cm s^{-2}]
ρ_f — fluid density, [g cm^{-3}]
σ — proportionality constant
∇ — gradient operator: $\mathbf{i}\,\partial/\partial x + \mathbf{j}\,\partial/\partial y + \mathbf{k}\,\partial/\partial z$, [$cm^{-1}$]
$\nabla\times$ — curl operator, describes circulation of vector fields

INTRODUCTION

Metasomatic alteration is a manifestation of changes in state conditions caused by an influx of chemical components and thermal energy carried by hydrothermal fluids as they flow from one chemical environment into another. In magmatic environments this process persists for long times and imposes systematic chemical and mineral alteration patterns onto large volumes of rock. Advective transport of mass and energy is the primary cause of metasomatism. Advection shifts minerals and pore fluids out of local equilibrium and instigates irreversible reactions in the downstream environments. During the relaxation period back toward overall equilibrium a sequence of minerals stable at local state conditions forms along the reaction path. The abundance of this new assemblage is determined by the persistence of the global nonequilibrium condition initiated by fluid flow; the composition of the assemblage is determined by both local state conditions and the affinity of the inflowing fluid (Norton, 1988). As a consequence of the advection of affinity a chronological sequence of partial equilibrium assemblages is formed; mineralogical and chemical variations measure fluctuations about the equilibrium state and the abundance or extent of alteration measures the rock permeability.

Fluid flows in the near-field regions of magmas occur predominantly through percolation networks controlled by interconnected sets of fractures. Observations of the alteration associated with fractures in fossil remains of these systems reveal that the alteration minerals filled the fractures and decreased permeability (Titley and others, 1986). These studies also disclose a chronological sequence of fracture and

fracture filling events that suggests a force field for fluid flow persisted continuously through an episodic sequence of deformations that rendered the rocks permeable. The advective fluid flows were locally mitigated by periodic sealing of the percolation networks that remain in a tightly sealed condition today. However, the extensive alteration associated with these networks provides vivid evidence that relatively large permeabilities prevailed during the thermal event. Generally the zones of most extensive alteration appear also to have been zones of the greatest permeability during the active life of the system.

A close linkage between advective fluid flow and alteration is also apparent from numerical analysis of the processes that redistribute thermal, mechanical, and chemical energy. A set of differential equations can be derived that represent symbolically the dynamics of hydrothermal systems. These equations clearly demonstrate the interdependence of processes and imply that a functional relationship is likely to exist between permeability of a percolation network and the extent of irreversible mass transfer caused by fluid flow (Norton, 1984).

Permeability values determined from outcrop samples and drillhole tests (Brace, 1980) are gross underestimates because they do not represent the permeability during the active stages of hydrothermal systems. The values during the active alteration can only be determined through properties preserved in the rock that are functions of the permeability during the metasomatic event. The most reliable approach for estimating rock permeability values that relate directly to the process of metasomatism appears to derive from relations between the extent of alteration and the fluid flow field.

This communication discusses the basis for estimating permeability of temporally and spatially complex fracture networks. Although the theoretical relations are largely drawn from information on hydrothermal systems related to magmas and are most applicable to systems in which the transport processes are closely coupled, they can most likely be extended to regional metasomatic systems as well. The theory is developed by first summarizing the advective transport process, then relating the fluid velocity field to alteration, and finally deriving an expression for permeability in terms of the fluid flow and velocity fields. The theoretical development is followed by a brief discussion of the considerations that are necessary in the application of the equations to outcrops.

TRANSPORT OF COMPONENTS

The quantitative relationships between fluid velocity and the change in composition of a lithologic unit are summarized by considering the gain and loss of a hypothetical component, G, with respect to an arbitrary region of the system. The advective transport of this component along a fluid pathline by flow and its subsequent redistribution among local minerals and fluid is depicted by a conservation of mass equation. Formulation of this equation is based on the content of G in

the region and modes of transport through the region as expressed by Euler's Transport Theorem (Slattery, 1972).

Consider a permeable region that contains a total of $\hat{I}$ mineral phases and a fluid phase that flows through a percolation network in response to buoyancy forces caused primarily by thermal perturbations. The minerals in these systems are typically in a state of partial equilibrium, a condition that constrains reconstruction of the alteration process with standard state thermodynamic and kinetic data. Therefore the nomenclature used by Helgeson and others (1970) is used here to discriminate between those minerals in equilibrium with the aqueous phase (product minerals, $p = 1, 2, \ldots, \hat{P}$) and those not in equilibrium with the aqueous phase (reactant minerals, $r = 1, 2, \ldots, \hat{R}$). The total number of minerals in a local region of the system is $i = 1, 2, \ldots, \hat{P}, \hat{P} + 1, \ldots, \hat{I}$, where $\hat{P} + \hat{R} = \hat{I}$.

The amount of each phase in this hypothetical system is represented by its volume fraction, ϕ. Let ϕ_f represent the volume fraction fluid phase in the flow channels, ϕ_p the volume fraction of product minerals, and ϕ_r the volume fraction of reactant minerals. The fluid phase flows at Darcy velocity $\mathbf{v}_f$, but the rock matrix is assumed to be stationary, $\mathbf{v}_{rock} = 0$. Because the advective transport of components by the fluid occurs over much greater distances than diffusive transport through permeable rocks, the diffusive rate of change is omitted from the governing equation, and only the rate of change in G caused by the fluid flow and reaction is considered. This simplification also leads to a more compact form of the transport equations, without introducing significant errors (Norton and Taylor, 1979). Furthermore the localized transport caused by diffusion can be readily distinguished in outcrops from the more system-wide advective transport.

Dispersion, an important mechanism for component transport in many hydrologic experiments (Freeze and Cherry, 1979), is of unknown importance as a contributor to metasomatic alteration of systems in which the thermal, mechanical, and chemical processes are closely coupled. Should this mechanism prove to be significant, then appropriate modifications of the advection terms in the following equations will be necessary.

The conditions and limitations described above lead to a conservation expression that describes the differential changes in the G per cm^3 of the system. The total rates of change in $\mathcal{G}$ for a representative portion of the system with respect to time, t, is:

$$\overbrace{\frac{\partial(\phi\mathcal{G})_f}{\partial t}}^{\text{fluid}} + \overbrace{\sum_p \frac{\partial(\phi\mathcal{G})_p}{\partial t}}^{\text{products}} + \overbrace{\sum_r \frac{\partial(\phi\mathcal{G})_r}{\partial t}}^{\text{reactants}} + \overbrace{\mathbf{v} \cdot \nabla\mathcal{G}_f}^{\text{advection}} = 0 \quad (1)$$

where conservation of fluid mass, $\nabla \cdot \mathbf{v}_f = 0$, and fluid density variations are considered to be independent of time, $\partial\rho_f/\partial t = 0$, and space $\nabla\rho_f = 0$. Eq (1) expresses the metasomatic change in $\mathcal{G}$ among the fluid, reactant

and product phases, and the fluid advection of $\mathcal{G}$. Although the details of changes in multicomponent, multiphase systems caused by reactions require explicit description in terms of the activity/composition relationships that depict mass transfer between minerals and an aqueous phase containing ions and complexes, it is unnecessary to burden the derivation that follows with the complexity of those relations. The expansion of the first three terms in eq (1) with respect to standard state thermodynamic and kinetic variables is explicitly treated in the literature by Helgeson and others (1970), Helgeson (1970, 1979), and Lichtner and Helgeson (1984). The advection term is discussed by Norton (1988).

The conservation expression 1 is now used to determine the relationship between change in $\mathcal{G}$ within a lithologic unit and the fluid flow through the unit that caused that change.

VELOCITY: AN INVERSE FUNCTION OF ALTERATION

Advective transport of chemical components dominates over all other transport mechanisms around magmatic thermal anomalies because of coinciding steep gradients in fluid composition, temperature, and pressure. Large gradients in fluid composition, $\nabla\mathcal{G}_f$, are common because of contrasting lithologic compositions and because the buoyancy field that drives fluid flow is extremely nonlinear at temperature and pressure conditions over which the reaction affinities tend to be large. Each of these conditions is particularly pronounced during the early stages of thermal activity. Rapid temperature increases, also in the early cooling period, induce large strains that generate extensive fluid-flow networks. However, fluid flow dissipates both temperature and compositional gradients. Consequently, advection is greatest early in the event because of the coincidence of large gradients, $\nabla\mathcal{G}_f$, with large fluid velocities, $\mathbf{v}_f$. As the perturbation that initiated the activity decays, so does the advective rate of change in $\mathcal{G}$.

Alteration features diagnostic of the local state conditions are produced throughout the thermal-fluid flow event, but two thermal regimes tend to dominate. Extensive chemical and mineralogical alteration is produced in early periods when temperatures are high to moderate (1000°–700°C), fluid pressures are large, and stress trajectories variable. However, the greatest mineralogical and chemical alteration occurs when advective flows coincide with temperature conditions in the near-field of the critical point in the fluid system, around 450° to 350°C. Fractures continually propagate with the advancing thermal front, form percolation networks over both temperature ranges, and maintain open percolation networks with large permeabilities.

Systematic relations implied throughout the foregoing discussion and summarized in eq (1) motivate the following proposition:

The magnitude and direction of the volume-averaged fluid velocity can be estimated in those environments where transport and consequent alteration were dominated by advection.

Support for this proposition follows directly from the advective transport eq (1), which represents the process by which a local region, perturbed from equilibrium conditions, evolves back toward equilibrium. In situations where advection greatly dominates over all other forms of transport, eq (1) is strongly hyperbolic (Mitchell, 1969; Roache, 1972) and requires special integration methods to obtain a solution. However for the purposes of this analysis the functional relations among alteration effects and velocity are more easily conveyed to the reader if the differential equation is expressed in first-order finite difference terms. The only constraint imposed on the finite difference approximation as it affects this discussion is that the transportive properties of the advective rate of change term are preserved through the use of upstream finite differences. The gradient in $\mathcal{G}$ is written such that only fluid from the upstream, source region has an effect:

$$\begin{aligned}\mathbf{v}_f \cdot \nabla \mathcal{G}_f = v_x &\frac{(\mathcal{G}_x - \mathcal{G}_{(x-\Delta x)})}{\Delta x} \\ + v_y &\frac{(\mathcal{G}_y - \mathcal{G}_{(y-\Delta y)})}{\Delta y} \\ + v_z &\frac{(\mathcal{G}_z - \mathcal{G}_{(z-\Delta z)})}{\Delta z}\end{aligned} \tag{2}$$

where v_x, v_y, and v_z are the components of the Darcy velocity vector, and the backward difference expression implicitly involves the finite coordinate increments Δx, Δy, Δz in the upstream direction of the local velocity vector. For example, if the x-component of velocity, v_x, is negative, the difference expression becomes

$$v_x \nabla_x \mathcal{G}_f = v_x \frac{(\mathcal{G}_x - \mathcal{G}_{(x+\Delta x)})}{\Delta x} \tag{3}$$

Transient changes in local composition of the fluid, product, and mineral phases are approximated by forward differences in time:

$$\frac{\partial(\phi \mathcal{G})_f}{\partial t} = \phi_f \frac{(\mathcal{G}_f(t + \Delta t) - \mathcal{G}_f(t))}{\Delta t} \tag{4}$$

$$\sum_p \frac{\partial(\phi \mathcal{G})_p}{\partial t} = \sum_p \phi_p \frac{(\mathcal{G}_p(t + \Delta t) - \mathcal{G}_p(t))}{\Delta t} \tag{5}$$

$$\sum_r \frac{\partial(\phi \mathcal{G})_r}{\partial t} = \sum_r \phi_r \frac{(\mathcal{G}_r(t + \Delta t) - \mathcal{G}_r(t))}{\Delta t} \tag{6}$$

Eqs (4) through (6) are then combined with eq (3) and solved for the fluid velocity along the pathline 1:

$$v_l = -\frac{\Delta l}{\Delta t}\left\{\frac{\Sigma\phi_r(\mathcal{G}_r^{t_1} - \mathcal{G}_r^{t_0}) + \Sigma\phi_p(\mathcal{G}_p^{t_1} - \mathcal{G}_p^{t_0}) + \phi_f(\mathcal{G}_f^{t_1} - \mathcal{G}_f^{t_0})}{(\mathcal{G}_f^{x_1} - \mathcal{G}_f^{x_0})}\right\} \tag{7}$$

Eq (7) defines the velocity components in terms of measured compositional gradients averaged over the increments of time, Δt, and distance, Δl, along the flow pathline. This equation demonstrates that components of the fluid velocity can be expressed in terms of compositional data derived from spatially and chronologically ordered alteration events present in outcrops (Titley and others, 1986). Because metasomatic effects in outcrops are the products of processes evolved in a nonlinear manner for a long time, the sampling procedure may require special techniques. Prior to discussing these mapping techniques, the relationship between permeability and velocity is examined.

PERMEABILITY: AN INVERSE FUNCTION OF VELOCITY

Permeability is a measure of the geometric properties of percolation networks that control the magnitude of fluid flow. Determining permeability is a routine procedure in many engineering situations. Well-bore and laboratory techniques are standardized over much of the petroleum and water industries. However permeability of ancient geologic systems is not as readily measured because of: (1) problems of scale (Garven, 1986) and (2) because the current permeability of these systems is not a measure of their permeability when they were active. The most permeable zones in an active hydrothermal system become thoroughly sealed by the advective metasomatic process. Subsequent to the hydrothermal event, these same zones have permeabilities that are well below the detection limits of all known field and laboratory measurement techniques. The only indication of the percolation networks that existed in the system is the presence of alteration assemblages in systematic vein sets. This observation leads to the following proposition relating permeability to the velocity associated with the advective rate of change in the concentration of components:

Because the extent of alteration is a function of fluid velocity and velocity is a function of permeability, permeability values associated with the metasomatic alteration are an inverse function of fluid velocity.

An obvious caveat with this observation is that the force field for flow must remain constant, a situation easily satisfied because of the transport conditions imposed by the critical phenomena in H_2O-rich fluids (Norton, 1984). In the previous section, equations were derived for estimating the fluid velocity along a flow path using measurements of the metasomatic alteration along the same path. A comparable inverse function for estimating permeability is next derived. Because permeability is contained in the proportionality constant that relates the velocity and force fields, we first examine the driving force that causes fluid flow.

Force Field for Flow

Fluid flow around thermal anomalies is driven primarily by buoyancy forces caused by lateral gradients in the fluid density. The magnitude and direction of these forces can be closely approximated for

most geologic environments from (1) equation-of-state data and (2) estimates of the geometry and magnitude of the thermal or solute perturbation that produced the density anomaly. The flows that result from buoyancy forces also dissipate the anomalous fluid gradients.

Buoyancy forces are caused by unstable density distributions of fluid. In geologic situations where metasomatic activity is most vigorous, lateral variations in fluid density are the most important cause. The force per unit volume that acts on a fluid particle is

$$\mathbf{H} = \mathbf{g}\rho - \nabla P \tag{8}$$

where the density, ρ, for phases in a one-component system of constant volume is a function of the state conditions, $\rho = f(P, T)$. Although this function is highly nonlinear for H_2O-rich fluids, the consequence of variable density can be examined by considering a small, linear perturbation in density such that

$$\rho = \rho_o + \delta\rho \tag{9}$$

where ρ_o is the reference density for which the force and velocity fields are zero (Boussinesq, 1903). The pressure in eq (8) can therefore be expressed as a function of the pressure, p, associated with the density perturbation and a reference pressure, $\rho_o\hat{\mathbf{g}}z$:

$$P = p + \rho_o\hat{\mathbf{g}}z \tag{10}$$

Substituting eqs (9) and (10) into eq (8) gives the force field in terms of a perturbation density:

$$\mathbf{H} = \mathbf{g}(\rho_o + \delta\rho) - \nabla(p + \rho_o\hat{\mathbf{g}}z) \tag{11}$$

Applying the ∇ operator to the pressure and density term and canceling the $\rho_o\hat{\mathbf{g}}$ and $\rho_o\mathbf{g}$ terms, because $\rho_o\mathbf{g} = \rho_o\hat{\mathbf{g}}$, leads to an expression for the force field in terms of the perturbation pressure and density:

$$\mathbf{H} = \mathbf{g}\delta\rho - \nabla p \tag{12}$$

Velocity Field

The fluid velocity averaged over the flow field is proportional to the force field in eq (12):

$$\mathbf{v} = \sigma\mathbf{H} \tag{13}$$

where the proportionality constant σ can either be derived from Darcy's Law or shown from the Navier-Stokes equation by similarity arguments (Hubbert, 1940) to be the ratio of permeability to dynamic fluid viscosity, k/μ. Substituting this ratio for σ in eq (13) gives an expression for the conservation of fluid momentum.

$$\mathbf{v} = -\frac{k}{\mu}(\nabla p - \delta\rho\mathbf{g}) \tag{14}$$

This velocity field can be shown to be a rotational field for density driven flows by operating with the curl, $\nabla\times$, on eq (14):

$$\nabla \times \left[\mathbf{v} = -\frac{k}{\mu}(\nabla p - \delta\rho\mathbf{g})\right] \tag{15}$$

The curl operator applied to the velocity field,

$$\nabla \times \mathbf{v} = \begin{pmatrix} \mathbf{i} & \mathbf{j} & \mathbf{k} \\ \partial_x & \partial_y & \partial_z \\ v_x & v_y & v_z \end{pmatrix} \tag{16}$$

produces the cross-partial terms:

$$\nabla \times \mathbf{v} = \mathbf{i}\left(\frac{\partial v_z}{\partial y} - \frac{\partial v_y}{\partial z}\right) + \mathbf{j}\left(\frac{\partial v_x}{\partial z} - \frac{\partial v_z}{\partial x}\right) + \mathbf{k}\left(\frac{\partial v_y}{\partial x} - \frac{\partial v_x}{\partial y}\right) \tag{17}$$

Reducing this equation to two-dimensional, semi-infinite half-space in the x-z coordinate plane makes it applicable to the typical cross sections used in modeling geologic situations:

$$\nabla \times \mathbf{v} = \mathbf{j}\left(\frac{\partial v_x}{\partial z} - \frac{\partial v_z}{\partial x}\right) \tag{18}$$

Applying the curl operator to the force field in eq (14) eliminates ∇p because the curl of the gradient in a scalar function is zero. Consequently, only the density perturbation and z-component of gravity remain:

$$\nabla \times [\nabla p - \mathbf{g}\delta\rho] = \begin{pmatrix} \mathbf{i} & \mathbf{j} & \mathbf{k} \\ \partial_x & \partial_y & \partial_z \\ 0 & 0 & \hat{g}\delta\rho \end{pmatrix} \tag{19}$$

Expanding this matrix gives

$$\nabla \times [\mathbf{g}\delta\rho] = \mathbf{j}\hat{g}\frac{\partial(\delta\rho)}{\partial x} \tag{20}$$

Combining eqs (14), (18), and (20) and dropping the unit vectors gives

$$\left(\frac{\partial v_x}{\partial z} - \frac{\partial v_z}{\partial x}\right) = \hat{g}\left(\frac{k}{\mu}\right)\frac{\partial(\delta\rho_f)}{\partial x} \tag{21}$$

The dependence of the cross-partial derivatives of velocity on the fluid-density perturbations and on the ratio of permeability to viscosity provides the basis for developing a functional relationship between

advective metasomatic alteration and local rock permeability. Converting eq (21) to finite-difference form and solving for permeability gives

$$k = -\frac{\mu}{\hat{g}\Delta(\delta\rho)}\left\{\frac{\Delta x}{\Delta z}(v_x - v_{x_0}) - (v_z - v_{z_0})\right\} \tag{22}$$

where $\Delta(\delta\rho)$ is measured in the horizontal plane of the system. Therefore, permeability can be equated to the fluid velocity, viscosity, and density perturbations. The velocity is known from eq (7), the fluid parameters from equation-of-state, and the density perturbation from independent computations of the circulation force caused by the thermal anomaly. Therefore, if the perturbation values are calculated for a system and combined with values for the metasomatic changes and eq (22), approximations to permeability can be made.

For conditions commonly encountered in the crust, the equation-of-state for the H_2O-System is precisely represented by any one of several algorithms (Helgeson and Kirkham, 1974; Levelt Sengers and others 1983; Haar, Gallagher, and Kell, 1984; Johnson and Norton, Univ. Arizona unpub. software). To the extent that the perturbation can be attributed to thermal effects on pure H_2O, these algorithms provide a reasonable approximation to the behavior of natural fluids and to the density and viscosity required by eq (22).

MEASURING THE PARAMETERS

The quantitative relationships among metasomatic effects, fluid velocity, and permeability present a challenge to field-mapping methods. As implied by eq (1), gains and losses of components must be mapped with respect to both space and time if the map data are to be analyzed with respect to the systems history. The basis for such mapping must include both the chronologic description of metasomatic events and correlation of a portion of the total mineral and chemical alteration with each event. Although such mapping techniques have been applied in a very restricted sense (Norton, Taylor, and Bird, 1984), the following discussion touches on some of the implications the transport equations have in determining parameters of now dormant systems.

Many references could be made to the literature in which alteration effects are used to compute water–rock ratios. Although they are prime sources of useful data for computing velocities and permeability, they are omitted because of the confusing interpretations that presume that no transport occurs by advection in systems where fluid flow is demonstrably important. Instead, the reader's attention is drawn to the summary of these data by Gregory and Criss (1986) and Criss and Taylor (1986). In essence many of the same data sets used to estimate water–rock ratios are useful to estimating the permeabilities of the systems.

A brief review of parameters that must be measured or estimated follows. Information pertaining to the definition of the fluid buoyancy field, the chronology and extent of alteration, and mapping techniques

is drawn from papers by Titley and others (1986), Villas and Norton (1977), Norton and Taylor (1979), and Norton, Taylor, and Bird (1984). These papers report on studies that have applied the theoretical relationships described above.

ESTIMATING THE FORCE FIELD FOR FLUID FLOW

The following list correlates the terms in eqs (7) and (22) with field observations and computations that are useful in assigning values to the respective forces that cause the fluid flow.

$\delta\rho_f$ The maximum density perturbation initially coincides with the contacts between a thermal anomaly and its host rocks. Because energy is dispersed from the anomaly into its surroundings, it shifts laterally away from the contact and upward into the lithocap region (Norton and Knight, 1977).

Description of the geometric form of the anomaly provides information for estimating the temporal and spatial variations in the force field. The geometric form of the pluton contact can often be estimated from a combination of outcrop tracing and magnetic and gravity data (Petraske, Hodge, and Shaw, 1978; Pollard and Johnson, 1973). Computation of the deformations to determine the stable configuration of the walls and lithocap for a given set of regional stress and magma pressure is useful in providing an independent estimate of the form of the chamber (Norton, Taylor, and Bird, 1984).

First order estimates of the initial fluid-density perturbation can be obtained from simple conductive heat-flow computations using the analytical equations (Jaeger, 1968; Lovering, 1935) appropriate for a simplified geometry of the system. A time–series of temperature values derived from these computations, together with estimates of confining pressure, can be used with an equation-of-state for the fluid to compute density gradients in the horizontal plane. Depth locations used for such calculations should coincide with the estimated depth of the observed metasomatic assemblages in the field. Although these values will not be correct in detail because of the effect of fluid flow on the temperatures, they provide a reasonable first approximation.

μ The dynamic fluid viscosity can be calculated for each of the profiles in the density determined above.

The items listed above define the buoyancy and resistive forces of fluid flow. The buoyancy force is an inevitable consequence of steep-sided thermal anomalies and pervades the entire region. The magnitude of flow then depends only on the viscose resistance and the properties of the percolation network.

MEASURING METASOMATIC GAINS AND LOSSES

The next stage requires description of the fossil percolation network and its mineralogy in order to establish the chronology of fracture events, orientations of the percolation paths, and the extent of metasomatism associated with each event. A reference state for calculating the metasomatic changes is first established; if this reference is not one of constant volume then the equations presented here must be transformed into the reference condition selected.

All veins and metasomatic fronts within a region are categorized into geochemical map units consistent with the flow-channel chronology established above. Early in the mapping program, provisional values of permeability can be derived using the techniques already described. These provisional values help define geochemical map units and delimit areas of greatest interest. In the reconnaissance stage, geometric properties and relative chronology of the fossil fluid-flow channels, as well as visual estimates of compositional changes, should be gathered. A combination of thermometric data from fluid inclusions and mineral assemblages compiled from samples collected from veins is one source for this information. Each vein sampled should be assigned a relative age with respect to other veins, and the metasomatic alteration within the contiguous matrix blocks is correlated with a specific vein set (Villas and Norton, 1977; Titley and others, 1986).

The orientations of discontinuous, slit-like features filled with alteration minerals are indicative of the system's chronology. Perhaps the most poorly defined portion of the methodology proposed lies in the mapping of fracture networks. The observer must note veins and tightly sealed pores whose ability to transmit fluids—although nil under current conditions—was enormous in the geologic past. Although visual inspections may imply extremely low percolation qualities, veins are the fossilized remains of percolation paths and must be mapped. Establishing that at least limited advective metasomatism occurred allows one to conclude that permeability values was greater than around 10^{-14}cm^2, a value estimated from numerical models, laboratory measurements of metasomatic effects, and field observations (Meyer, 1950; Norton and Taylor, 1979). To estimate permeability, the following observations must be made:

Δv_x and Δv_z The values of these parameters are derived from eqs (7) through (22). Initial estimates are best made in a region of the system where a boundary can be identified between permeable and impermeable units. Such boundaries allow the velocity difference to be referenced to a $v_{x_0} = 0$ or $v_{z_0} = 0$ condition and simplifies eq. (22).

Temporal variations in permeability are a direct consequence of geometric changes in the pore network caused by stress. Because pore shapes and dimensions

are sensitive to slight variations in the balance between fluid pressure on their walls and confining stress, the manifestation of the deformation process should be recorded. Single, isolated pores sequentially evolve from dispersed microcracks into megascopic fracture sets whose intersects produce a percolation network. Although the mechanisms that produce this evolution are not well understood, the geometric form of these networks is fundamental to deciphering these mechanisms.

Effective lengths of alteration zones measured parallel to the major flow directions are a measure of the velocity of specific reaction and composition fronts through the system. These lengths are best referenced to the sourceregion for a particular component. In essence, a chemical condition is extended from an upstream lithologic unit into the downstream unit, a distance proportional to the fluid velocity and the duration of the event. Perhaps one of the most useful pieces of supporting data is the determination of the total time duration of a particular process.

THE TIME SCALE

Thermal perturbations that cause metasomatism last 10^5 to 10^6 yrs, and during this time, a region of 10^2 to 10^3 km^3 is affected. Numerical analyses of the processes within such a region requires quantitative estimates of the chemical and mineral gains and losses throughout the system. Studies must correlate the scale of changes that occur within map units with those of the computational analysis. Independent of these scale constraints microscopic textures remain fundamental in establishing paragenetic sequences of events independent of the scale of measurements made of gains and losses of components or minerals.

The net temporal change in $\mathcal{G}$ is estimated from concentration in outcrop before and after the alteration event. A distinction among changes associated with the reactant versus product phases helps to refine estimates but is of minor importance on the first estimates of permeability. The net change in concentration taken with respect to the lithologic unit can be substituted for the difference terms in the numerator of eq (3).

The remaining term, $\Delta x/\Delta t$, is the velocity of an isograd of $\mathcal{G}$, $\mathbf{v}\mathcal{G}$. Knowing the duration of the thermal event is helpful at this step, but even a reasonable approximation based on a conductive cooling model will suffice. The time required for the thermal anomaly to return to $1/e$ of the initial perturbation (Lovering, 1935; Jaeger, 1968) is used for the value of Δt. The position of a value of $\mathcal{G}$ that is intermediate between the values used to calculate the spatial change above is marked. Then, the position of this same value of $\mathcal{G}$ prior to the thermal event is marked, and

the distance between these points is Δx. Values for the volume fractions, porosity, and density can be estimated from standard measurements of the mineral modes and gravimetric methods (Norton and Knapp, 1977).

Three difference expressions for the concentration of $\mathcal{G}$ appear in eq (7) (activity–composition relations are also required but are omitted here for reasons already indicated). Two of the expressions are with respect to time variation of $\mathcal{G}$ in the mineral assemblage; one for product minerals and one for the reactant minerals. In the denominator, a difference term occurs that describes the spatial variation of $\mathcal{G}$ in the fluid. As a first-order approximation to the magnitude of the latter term, consider the concentration of $\mathcal{G}$ in two distinct outcrops, x_1 and x_0, at time, t. Assuming that the phases containing $\mathcal{G}$ were in equilibrium with the aqueous fluid, concentration in the fluid can be deduced from standard thermodynamic data, phase relations in activity–activity coordinates, and estimates of the temperature and pressure at these positions.

CONCLUSIONS

The transport theorem has been used as a basis from which to derive finite-difference formulas for fluid velocity and permeability as functions of metasomatic alteration and fluid buoyancy. These equations provide a theoretical basis for using metasomatic alteration and the orientation and chronology of veins to predict average values of fluid velocity and permeability. The combination of field mapping, laboratory analyses, and computations provides a highly effective technique for describing the transport properties of natural metasomatic systems.

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