Oxygen isotopic composition of veins and host rocks as tracers of fluid rock interaction in the crust

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Veins represent important indicators of fluid-rock interaction and they occur at all levels within the lithosphere. Veins associated with cooling plutons and certain classes of hydrothermal ore deposits are interpreted in terms of open system, fluid infiltration metasomatism. While these veins are important for their economic reserves of minerals, veins formed in deforming rocks are volumetrically more significant. The latter class of veins typically grow in response to the local stress regime as the rock mass deforms. Oxygen isotope analyses of veins and their coexisting host rocks provide insights into the style of fluid-rock interaction and the mechanisms of isotope exchange in the crust. In deforming rocks, veins grow from the prefolding, earliest stages of tectonic shortening through the various stages of folding and fabric development. Depending upon the post-tectonic thermal history, rocks with over 50% shortening and weak fabric development preserve vein growth events over the entire deformation and hence record the non-isothermal history of fluid-rock interaction.

In solving the general oxygen isotope exchange problem, the δ values of coexisting phases are the natural variables because temperature and mass balance constraints are coupled and can be visualised simultaneously in graphs of one δ value against another. For veins, the natural δ coordinates are the $\delta^{18}O$ values of vein quartz and the fluid. The $\delta^{18}O$ values of the coexisting host rocks or the other minerals can be represented graphically on the zero fractionation line ($\Delta = 0$). In the general n-phase exchange problem, there are n-1 kinetic exchange equations and a material balance equation which is the continuity equation of fluid mechanics. The latter equation includes the effect of externally-derived fluid. Gregory *et al.* (1989) discussed this problem for coexisting minerals from hydrothermally-exchanged plutonic rocks. For the case where dimensionless ratio of the fluid flux to kinetic rate constant is much greater than 1 ($u/k \gg 1$), the general problem is greatly simplified and adequately described by the open system fluid-rock equations of Taylor (1977). In this latter model, a fixed volume of rock incrementally exchanges with packets of pristine fluid that are allowed to equilibrate with the rock before their removal as the next packet is added to the rock. For short times, the 'closed' (fluid recycling) and the open system (one pass) solutions are similar, i.e. $\ln(1 + W/R_{closed}) = W/R_{open} \sim W/R_{closed}$.

For the problem of vein growth, it is convenient to consider the general exchange problem from the frame of reference of the fluid. In the crust where $u/k \ll 1$, the general exchange problem also greatly simplifies to the situation where a fixed volume of moving fluid exchanges progressively with small increments of pristine rock. The differential equation that describes the isotopic composition of the fluid is of the form [e.g. Gregory and Taylor, 1981]:

$$w \, d\delta w/dt = -\Sigma r_i(\delta_w + \Delta_{iw} - \delta_i^0) \quad (1)$$

where w is the mass of the fluid in oxygen units, and the r_i represent the mass of rock reacted per unit time. If all of the minerals in the layers have similar reaction rates [e.g. Walther and Wood, 1986], then $\Sigma r_i = k$R where R is the total oxygen content of the layers . Using this identity and assuming that w is small compared to the mass of rock, w transforms to x_ww, equation 1 becomes:

$$-x_w/k \, d\delta w/dt = \delta w + \Sigma x_i(\Delta_{iw} - \delta_i^0) \quad (2)$$

The analytical solution to this equation with $x_i(\Delta_{iw} - \delta_i^0)$ and x_w/k's taken as constants is:

$$\delta w = (\delta w^0 - \delta w_{steady\ state}) exp - (kt/x_w) + \delta w_{steady\ state} \quad (3)$$

The multiplicative constant of t (k/x_w) in the argument of the exponential term has units of 1/time and its inverse gives the characteristic time to reach steady state. Implicit in this analysis is that the integrated fluid rock ratio is small, because we have assumed that the mass of rock is sufficiently large so that the δ_i^0s were held constant during the integration of equation 2. For $x_w \ll 1$, the fluid-rock ratio is close to x_w, itself. Therefore, the characteristic time will depend on the fluid-rock ratio, as well as the rate constant. By suitable variable substitutions, the argument of the exponent in equation 1 can be represented also by the inverse integrated water/rock ratio, R/W, or the product of the rate constant times the distance the fluid travels divided by mole fraction of water times the fluid velocity, $(kx)/(x_w v)$.

In the crust under rock-buffered conditions, $x_w \ll \Sigma x_i$ and x_w can be neglected in the mass balance calculation of the δ values so that the steady-state fluid δ value is :

$$\delta w_{\text{steady state}} = \Sigma x_i(\delta_i^0 - \Delta \text{iw}) \quad (4)$$

The fluid $\delta^{18}O$ value consists of two parts: one term that refers to the isotopic composition of the rock mass, and the second which refers to the bulk fractionation factor associated with the exchange and reaction between silicate minerals and the fluid phase.

$$\delta w = \Sigma x_i \delta_i^0 - [\Delta \text{iw}]\text{avg} \quad (5)$$

The $\delta^{18}O$ of vein quartz is just $\delta w + \Delta qw$, where Δqw is the quartz–water fractionation at the temperature of precipitation. If ε is defined as:

$$\varepsilon \equiv \Delta \text{qw} - [\Delta \text{iw}]\text{avg}, \quad (6)$$

the $\delta^{18}O$ of vein quartz is:

$$\delta_{\text{vein quartz}} = \Sigma x_i \, \delta_i^0 + \varepsilon \quad (7)$$

Under crustal conditions, ε varies much less as a function of temperature than Δqw so that veins formed under rock rock-buffered conditions over the lifetime of a deformation event will display little change in $\delta^{18}O$ values as a function of their relative age.

On a graph of $\delta^{18}O$ values of veins (y-axis) versus the fluids, a rock-buffered system under non-isothermal conditions generates vein-fluid pairs that map out a subhorizontal trajectory (slope → 0 as u/k → 0) which points to mean of the whole-rock values on the zero fractionation line (a 45° sloped line passing through the origin). Transformation of this graph into parameters that can be measured ($\delta^{18}O$ values of veins and whole rocks; the parameters in equation 7) shows that rock-buffering of the fluid on scales larger than a single rock layer produces a steep array in a whole rock versus vein (x-axis) plot. Numerical simulations using the general exchange model indicate that this behavior will be observed for $u/k \ll 1$. This type of relationship has been observed in the Lachlan fold belt (Gray *et al.* 1991) indicating that ubiquitous veining does occur without massive fluxes of externally-derived fluids.

References

Gray, D.R., Gregory, R.T. and Durney, D.W. (1991) *J. Geophys. Res.*, **96**, 19,681–704.

Gregory, R.T. and Taylor, H.P., Jr. (1981) *J. Geophys. Res.*, **86**, 2737–55.

Gregory, R.T., Criss, R.E. and Taylor, H.P. (1989) *Chem. Geol.*, **75**, 1–42.

Taylor, H.P., Jr. (1977) *J. Geol. Soc. London*, **133**, 509–58.

Walther, J.V. and Wood, B.J. (1986) *Advances Phys. Geochem.*, **5**, 194–212.