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Regional metamorphic dehydration and seismic hazard

J.J. Ague, Jeffrey Park, and D.M. Rye

Department of Geology and Geophysics, Yale University, New Haven, Connecticut

Abstract. We use a one-dimensional model of coupled heat transport, fluid transport, porosity feedback, and metamorphism to investigate how dehydration reactions, proceeding during burial of hydrous rocks in the footwall of thrusts or the hanging wall of normal faults, may influence the temporal evolution of pore fluid pressure (P_f). Based on a new interpretation of existing kinetic data for the dehydration of model serpentinite, we advance the hypothesis that P_f buildup to tensile failure may be possible within decades of the onset of dehydration in low permeability metamorphic rocks of the middle crust. This model contrasts with both: 1) typical earthquake nucleation models which involve preexisting fluids that play a passive role, responding to tectonic stresses rather than generating them and 2) conventional treatments of P_f generation based on equilibrium thermodynamics and constant rates of dehydration, which require 10^3–10^6 years to produce significant P_f increases. Our results suggest that repeated cycles of dehydration–induced seismicity are plausible in low permeability areas of the mid–crust, and can be sustained throughout an orogenic episode if fresh hydrous minerals descend into the metamorphic zone along the downgoing face of a thrust- or normal-fault system.

1. Introduction

High pore fluid pressure (P_f) decreases rock strength and reduces frictional resistance to sliding along faults [*Hubbert and Rubey*, 1959; *Hanshaw and Bredehoeft*, 1968; *Scholz*, 1990; *Rice*, 1992]. Meteoric fluids are common along upper–crustal faults and may aid earthquake nucleation if P_f is high [*Scholz*, 1990; *Smith and Arabasz*, 1991; *Rice*, 1992], but are unlikely to penetrate to midcrustal depths, near the brittle-ductile transition, where many moderate ($M \sim 6-7$), but damaging, earthquakes nucleate [*Smith and Arabasz*, 1991; *Stein et al.*, 1994; *Davis and Namson*, 1994; *Zhao et al.*, 1996]. At these depths (10–20 km), however, metamorphic dehydration is a major source of fluids [*Etheridge et al.*, 1984; *Nishiyama*, 1989].

A cornerstone of many current earthquake nucleation models is that high pore-fluid pressure P_f (in excess of the hydrostatic gradient) results in rock failure and initiation of fault movement [*Stein and Lisoski*, 1983; *Walder and Nur*, 1984; *Sibson*, 1992; *Byerlee*, 1993]. Moreover, for certain types of faults, such as steep reverse faults, P_f must exceed the solid (or overburden) pressure (P_s) in order for repeated faulting to occur [*Sibson*, 1992]. However, mechanisms that can repeatedly generate P_f in excess of P_s over time scales relevant for the recurrence of significant earthquakes (10^2–10^3 years) have proven elusive. We propose a mechanism for generating elevated P_f and hydrofracture in metamorphic systems that is based on a kinetic description of the dehydration process that requires no infiltration of exotic fluids from outside the rock mass.

2. Model

Deep burial of hydrous rocks in the footwall of thrust faults or the hanging wall of normal faults will result in heating and associated prograde metamorphism and volatile release (Figure 1). We consider the water released by dehydration reactions among silicate minerals using the example serpentine dehydration reaction shown in Figure 2. This reaction is a reasonable model for the dehydration of serpentinite, a rock common to active fault zones worldwide, including the San Andreas fault, California, USA [*Jennings*, 1977]. We emphasize, however, that our results are general and applicable to metamorphism of a wide variety of hydrous protoliths including clay–rich sediments and hydrothermally altered volcanic rocks. Like many prograde reactions, the change in fluid volume for serpentine dehydration is large and positive, the change in solid volume (ΔV_s) is small and negative, so that the total volume change of reaction is positive. If the reaction proceeds very near the equilibrium $P_f - T$ condition, the rates and amounts of fluid pressure generation are limited because small increases in P_f drive the reaction back to thermodynamic equilibrium, halting dehydration (Figure 2, Point A). Reactions can proceed, however, at temperatures significantly greater than the equilibrium T, i.e., "T overstepping" [*Putnis and Holland*, 1986; *Lasaga and Rye*, 1993]. T overstepping may result from, for example: 1) energetic barriers to mineral nucleation and/or dissolution [*Putnis and Holland*, 1986; *Ridley and Thompson*, 1986] that cause the reaction to begin at temperatures in excess of the equilibrium T, 2) rapid heating following introduction of hydrothermal fluids or emplacement of intrusions [*Lasaga and Rye*, 1993], and 3) P_f decrease following a fracturing and fluid release event. Based on field, experimental, and theoretical evidences, the amount of overstepping probably ranges from a few °C to as much as 80 °C [*Putnis and Holland*, 1986; *Ridley and Thompson*, 1986; *Lasaga and Rye*, 1993]. In the overstepped scenario, P_f in the reaction zone can increase dramatically if fluid escape is locally impeded by low permeability (Figure 2, Point B).

We adopted the following general rate law to describe the kinetics of dehydration [*Lasaga and Rye*, 1993]

$$R_r = \frac{\partial m_\theta}{\partial t} = k^o \exp\left(\frac{-Ea}{R}\left(\frac{1}{T} - \frac{1}{T^o}\right)\right) \nu_\theta \bar{A} s |\Delta G_r|^n \tag{1}$$

where R_r is the reaction rate; m_θ is the moles of species θ per unit volume of rock; t is time; k^o is the intrinsic reaction rate constant at a convenient reference temperature T^o; Ea is the activation energy; R is the gas constant; T is the absolute temperature; ν_θ is the stoichiometric coefficient of species θ; $\bar{A}$ is the surface area

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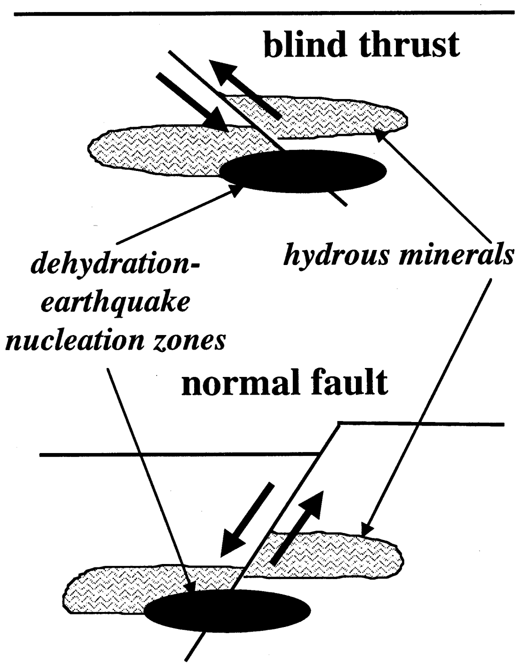

Figure 1. Cartoon illustrating possible links between metamorphic dehydration and earthquake faulting. Hydrous minerals descending on the hanging wall of a normal fault or footwall of a thrust undergo dehydration. If the permeability of the rocks in the reaction zone is low and kinetic reaction overstepping is important, then pore fluid pressure will increase rapidly in the reaction zone, the rock strength will decrease, and hydrofracture and fault rupture are possible.

of the rate-limiting mineral; $|\Delta G_r|$ is the absolute value of the Gibbs free energy change of the reaction at the T and P_f of interest [*Dahlen*, 1992]; n is the reaction order; and s is, by convention, +1 if ΔG_r is negative and -1 otherwise. We used the least–squares method to fit the experimental rate data of *Wegner and Ernst* [1983] for the serpentinite dehydration reaction to the form of equation (1). We took Ea=20,000 calories/mole [*Lasaga and Rye*, 1993] and T^o=633 K (the equilibrium T at 1 kbar). We assumed that the rate limiting mineral was brucite (the reactant with the least surface area in the experiments). $\bar{A}$ was then evaluated assuming an initial grain radius of 5×10^{-4} cm [*Wegner and Ernst*, 1983] and the expressions of *Lasaga and Rye* [1993]. ΔG_r was calculated using the data of [*Berman*, 1991]. The best–fit values of k^o and n are 4.08×10^{-18} moles/cm^2/yr/(calories/mole)^n and 3.64, respectively. We note that k^o and n values determined for other dehydration reactions [*Lasaga and Rye*, 1993] produce results that are comparable to those we report below.

Using standard finite–difference methods we model in one space dimension (1–D; x horizontal) the temporal and spatial evolution of P_f and T in a horizontal rock layer which contains an x = 150 m wide reaction zone. The layer is assumed to be sealed on both top and bottom in this geometry, but fluid is allowed to escape horizontally along the layer away from the reaction zone. *A posteriori* considerations indicate that a 2-D or 3-D geometry, though more complex computationally, would exhibit behavior similar to this 1-D case. The evolution of P_f follows a form of the hydraulic diffusion equation [*Walder and Nur*, 1984; *Wong et al.*, 1997]

$$c\frac{\partial P_f}{\partial t} = -\left(\frac{\partial \phi}{\partial t}\right) + \frac{1}{\rho_f}\left(\frac{\partial M_f}{\partial t}\right) + \frac{\partial}{\partial x}\left(\frac{\kappa}{\mu}\frac{\partial P_f}{\partial x}\right) \quad (2)$$

Here, $c = \phi(\beta_f + \beta_\phi)$. Our model accounts for kinetically–controlled changes in the mass M_f of fluid per unit volume of rock; changes in porosity ϕ due to the volume change of reaction and "pore compressibility" β_ϕ [*Walder and Nur*, 1984]; and Darcian fluid flow, governed by permeability κ and fluid viscosity μ. The fluid density (ρ_f) and compressibility (β_f) were computed using a modified Redlich–Kwong equation of state [*Holloway*, 1987]. The term on the left hand side of equation (2) describes the reversible change in the volume of fluid per unit volume of rock and incorporates the elastic compressibilities of fluid and pore space. The first term on the right represents the effects of irreversible changes in pore volume due to, in our model, ΔV_s. The second term represents the contribution of fluid volume from dehydration. The third term represents the Darcian fluid flux. Changes in the reactive surface area were calculated using the expressions of *Lasaga and Rye* [1993]. We take $\mu = 5 \times 10^{-4}$ Pa s. Following *Wong et al.* [1997], we use a simple model of ϕ–κ feedback thought to be appropriate for relatively low permeability rocks: $\kappa = \kappa_0 (\phi/\phi_0)^3$; where κ_0 and ϕ_0 are the initial permeability and porosity, respectively. *Wong et al.* [1997] estimated $\beta_\phi = 5 \times 10^{-9}$ Pa^{-1} based on experiments with gypsum samples. Because metamorphic rocks are likely to be significantly less compressible than gypsum aggregates [*Etheridge et al.*, 1984], we set $\beta_\phi = 1 \times 10^{-9}$ Pa^{-1}. As the rate of reaction is temperature–dependent, the effects of coupled chemical reaction, conduction, and convection on the temperature field were evaluated following *Lasaga and Rye* [1993].

3. Results

We report computations for end-member initial permeabilities and for lithostatic and hydrostatic initial conditions in the mid-crust. Given that the tensile strength of most schistose rocks such as serpentinites is <100 bars [*Etheridge et al.*, 1984], we presume that

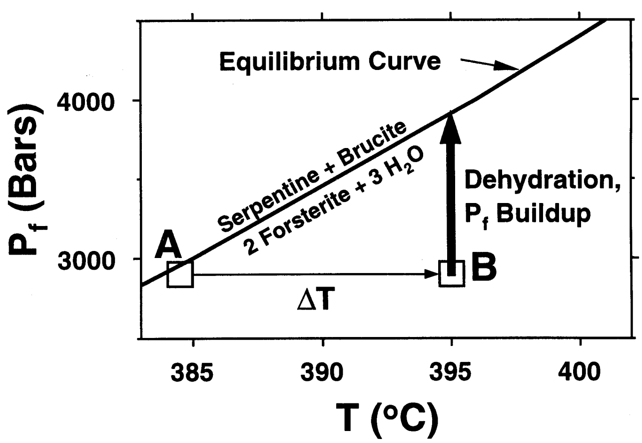

Figure 2. Pore fluid pressure (P_f) generation during dehydration of low permeability rock. Thick black line, equilibrium curve for the reaction: serpentine + brucite = 2 forsterite + 3 H_2O. If P_f and temperature (T) are near the equilibrium condition (Point A), then P_f generation is limited. If the equilibrium boundary is overstepped by an amount ΔT (Point B), then P_f can increase rapidly as dehydration proceeds.

failure is likely once P_f exceeds solid pressure P_s by this amount. Low permeabilities $(\kappa_o) = 1 \times 10^{-21}$ m^2 in the dehydrating reaction layer and its surroundings are appropriate for many types of crystalline metamorphic rocks [*Walder and Nur*, 1984; *Zhao et al.*, 1996; *Wong et al.*, 1997]. Starting with P_f equal to P_s at 3000 bars (300 MPa; $\sim$ 11.5 km), Figure 3*a* indicates that tensile failure (hydrofracture) is possible on decade timescales for T-oversteps (ΔT) as little as 5 °C. The rise of P_f diminishes as P_f approaches the equilibrium pressure and $|\Delta G_r|$ goes to zero. The rate of P_f increase is greater with larger ΔT, primarily because $|\Delta G_r|$ increases with ΔT. T changes in the reaction zone are small, have a negligible impact on reaction progress, and are not discussed further. Although considerable uncertainties surround the kinetics of individual metamorphic reactions, the serpentine-olivine system illustrates the general principle that significant fluid overpressures ($P_f > P_s$) by dehydration go hand-in-hand with kinetic reaction overstepping.

In the second example, initial P_f and P_s are 1000 and 3000 bars, respectively; this approximates the situation where the reaction zone starts at hydrostatic P_f prior to sealing. Here, larger ΔT values are required to reach $P_f > P_s$ conditions. Nonetheless, with $\Delta T = 40$ °C, the nominal threshold for hydrofracturing is exceeded over decade time scales for $\phi_o = 0.001$ (Figure 3*b*). Moreover, it is critical to point out that increases in P_f will decrease the rock strength dramatically [*Hubbert and Rubey*, 1959; *Hanshaw and Bredehoeft*, 1968; *Scholz*, 1990; *Rice*, 1992] regardless of whether the nominal threshold for tensile failure is reached via this mechansism. If the crust is already under tectonic stress, this weakening may be enough to trigger an earthquake.

When failure occurs, porosity is created, rock permeability increases, and fluid flows out of the fractured zone [*Walder and Nur*, 1984]. P_f will tend to drop, the degree of T overstepping will increase, and metamorphic dehydration will proceed at an increased rate. If permeability returns rapidly to low values due to, for example, mineral precipitate seals [*Walder and Nur*, 1984], then P_f will increase and promote renewed rock weakening. Consequently, dehydration may take an *active* role in re–pressurizing the rock after fault rupture [cf. *Hacker*, 1997]. For the cases shown in Figure 3, only small amounts of the reactants (between ~0.1 and ~5 per cent) are consumed during 300 years of reaction. Little reactant is consumed thereafter because reaction rates go to zero as equilibrium is approached. Thus, a given rock volume may produce hundreds of episodes of elevated P_f, consistent with observations of metamorphic veins, which often show clear evidence of repeated fracturing and healing [*Cox*, 1995].

On the other hand, if the fault goes unsealed and κ remains high, then P_f buildup will be limited and the reaction zone will tend to drain relatively rapidly. For example, if the initial conditions are $\kappa_o = 10^{-17}$ m^2, $P_f = P_s = 3000$ bars, and $\Delta T = 5°$ C, then P_f is only ~20 bars $> P_s$ after ~300 years of reaction, and about 0.5% of the reactants are consumed. If P_f remains low, the reaction may be stranded far from equilibrium in P–T space and the reaction zone could dehydrate completely in only ~60,000 years, in contrast to classical metamorphic models in which dehydration proceeds slowly and continuously over million year time scales.

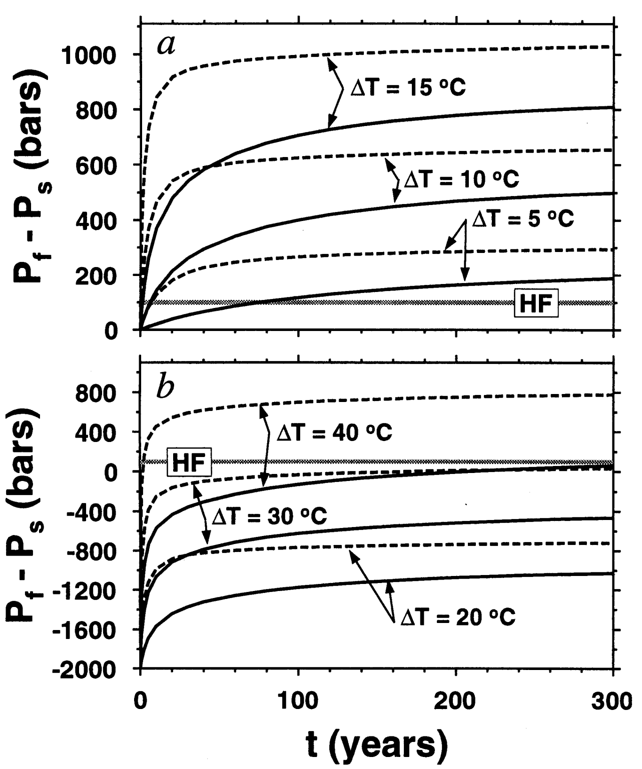

Figure 3. Model pore fluid pressure (P_f) versus time (years) curves at center of 150 m wide reaction zone. Rock initially contains 1:1 molar ratio of serpentine [chrysotile; $Mg_3Si_2O_5(OH)_4$] and brucite [$Mg(OH)_2$] crystals having grain radii of 0.01 cm. Solid and dotted curves denote calculations for initial porosities of 0.01 and 0.001, respectively. ΔT is the initial temperature overstep. Approximate minimum P_f required for tensile failure (hydrofracturing) assuming rock tensile strength of 100 bars denoted by gray line marked HF. *a*, P_f and solid (or overburden) pressure (P_s) are initially 3000 bars (300 MPa). *b*, P_s is 3000 bars, P_f is initially 1000 bars.

4. Discussion

Because the porosity of metamorphic rocks is low and the fluid is nearly incompressible, only a small volume of fluid is necessary to trigger a seismic event in low permeability rocks with a rapid increase in P_f. Consequently, there need be no observable trace of an initial volume increase in seismic records. However, our metamorphic nucleation mechanism predicts a fluid-filled mid-crustal hypocentre, as has been inferred for both the 1995 Kobe [*Zhao et al.*, 1996] and 1989 Loma Prieta [*Lees and Lindley*, 1994] earthquakes from tomographic inversion. Our model could also help explain certain precursory phenomena related to preparatory moment release [*Ellsworth and Beroza*, 1995], fluid-discharge [*Tsunogai and Wakita*, 1995] and electromagnetic signals [*Fraser–Smith et al.*, 1990], by relating them to the release of saline fluid and pressure caused by rock failure preceding complete rupture of the fault zone.

On longer time scales, repeated earthquakes on a single fault system could be triggered as long as hydrous minerals are available for dehydration (Figure 1). If a thrust or a normal fault intersects a stack of hydrous metapelites within a sedimentary basin, fresh rock for mid-crustal dehydration will be transported to earthquake nucleation depths on the downdropping side of the fault. Metamorphism along a strike-slip fault might consume any pre-existing supply of hydrous minerals, unless its transcurrent motion is accompanied by sufficient vertical displacement (either by faulting or basin subsidence) to transport shallower rocks into the nucleation zones. These scenarios raise the important possibility that metamorphism may itself be a partial cause, not just an effect, of orogenic events. We speculate that feedbacks between dehydration and vertical fault movement encourage thrust and normal faulting in thickly–sedimented continental crust along transcurrent plate boundaries, e.g., the San Andreas Fault system.

The dehydration hypothesis for crustal earthquakes appears best suited for the middle crust, near the brittle–ductile transition. In this depth range, fault permeability should be limited by the incipient plastic deformation of rocks surrounding the reaction zone, the infiltration of meteoric fluids is greatly restricted, the crustal geotherm intersects many dehydration reactions with steep dP/dT slopes, and T may be sufficiently low for energetic barriers to mineral nucleation and growth to be considerable and encourage significant reaction overstepping. Testing the hypothesis, however, will require more extensive experimental data on the kinetics of the suspect dehydration reactions, as well as a careful assessment of midcrustal mineralogy in seismogenic zones, either contemporary or exhumed by later erosion.

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J.J. Ague, J. Park, and D. M. Rye, Department of Geology and Geophysics, Yale University, P.O. Box 208109, New Haven, CT 06520–8109 USA. (e-mail: jay.ague@yale.edu, park@hess.geology.yale.edu, rye@hess.geology.yale.edu)

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