Regional metamorphic dehydration and seismic hazard

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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 pre-existing 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 $10^8$–$10^9$ 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; Hanks and Bredehoef, 1968; Schols, 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 ($M = 5$–$7$), but damaging, earthquakes nucleate [Smith and Arabasz, 1991; Steen 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; Nishiya, 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 ($P_o$) in order for repeated faulting to occur [Sibson, 1992]. However, mechanisms that can repeatedly generate $P_f$ in excess of $P_o$ 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 ($\Delta V_s$) is small and negative, so that the total volume change of reaction is positive. If the reaction proceeds very near the equilibrium $P_f=0$ 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 modeling studies, overstepped reactions are characterized by 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].

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

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

(1)

where $R_r$ is the reaction rate; $m_\theta$ is the moles of species $\theta$ per unit volume of rock; $t$ is time; $k^\theta$ is the intrinsic reaction rate constant at a convenient reference temperature $T^o$; $E_a$ is the activation energy; $R$ is the gas constant; $T$ is the absolute temperature; $\nu_\theta$ is the stoichiometric coefficient of species $\theta$; $A_s$ is the surface area
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 $\Delta G_r$ is negative and -1 otherwise. We used the least-squares method to fit the experimental rate data of Wegener and Ernst [1983] for the serpentine dehydration reaction to the form of equation (1). We took $E_a=20,000$ calories/mole [Lasaga and Rye, 1993] and $T_0=633$ 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). $A$ was then evaluated assuming an initial grain radius of $5 \times 10^{-4}$ cm [Wegener and Ernst, 1983] and the expressions of Lasaga and Rye [1993]. $\Delta G_r$ was calculated using the data of [Berman, 1991]. The best-fit values of $k^0$ and $n$ are $4.08 \times 10^{-18}$ moles/cm$^2$/yr/(calories/mole)$^n$ and 3.64, respectively. We note that $k^0$ 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]

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

(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 $\phi$ due to the volume change of reaction and "pore compressibility" $\beta_\phi$ [Walder and Nur, 1984]; and Darcian fluid flow, governed by permeability $\kappa$ and fluid viscosity $\mu$. The fluid density ($\rho_f$) and compressibility ($\beta_f$) 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, $\Delta V_f$. 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 $\phi$–$\kappa$ feedback thought to be appropriate for relatively low permeability rocks: $\kappa = \kappa_0 (\phi/\phi_0)^3$; where $\kappa_0$ and $\phi_0$ 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](image_url)
failure is likely once $P_f$ exceeds solid pressure $P_s$ by this amount. Low permeabilities ($\kappa_s = 1 \times 10^{-21} \text{ m}^2$) 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; \( \sim 11.5 \text{ km} \)), Figure 3a indicates that tensile failure (hydrofracture) is possible on decade timescales for $T$-oversteps ($\Delta T$) 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 $\Delta T$, primarily because $|\Delta G_r|$ increases with $\Delta T$. $T$ 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 $\Delta T$ 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 $\phi_s = 0.001$ (Figure 3b). Moreover, it is critical to point out that increases in $P_f$ will decrease the rock strength dramatically [Hubbert and Rubey, 1959; Hanshaw and Bredheoef, 1968; Scholz, 1990; Rice, 1992] regardless of whether the nominal threshold for tensile failure is reached via this mechanism. 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 \( \sim 0.1 \) and \( \sim 5 \) 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 [Coz, 1995].

On the other hand, if the fault goes unsealed and $\kappa$ 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 $\kappa = 10^{-17} \text{ m}^2$, $P_f = P_s = 5000$ bars, and $\Delta T = 5$ °C, then $P_f$ is only \( \sim 20 \) bars $\sim 50$ years after $\sim 300$ 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 $\sim 60,000$ years, in contrast to classical metamorphic models in which dehydration proceeds slowly and continuously over million year time scales.

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 [Tsunougai 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 transcurent motion is accompanied by suffi-
cient vertical displacement (either by faulting or basin subsidence) to transport shallower rocks into the nuclea-
tion zones. These scenarios raise the important possi-
bility that metamorphism may itself be a partial cause,
not just an effect, of orogenic events. We speculate that
feedbacks between deformation and vertical fault move-
ment encourage thrust and normal faulting in thickly-
sedimented continental crust along transient plate boundaries, e.g., the San Andreas Fault system.

The dehydration hypothesis for crustal earthquakes
appears best suited for the middle crust, near the brit-
tle–ductile transition. In this depth range, fault permea-
ability should be limited by the incipient plastic de-
formation 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 care-
ful assessment of midcrustal mineralogy in seismogenic zones, either contemporary or exhumed by later erosion.

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References

Berman, R. G., Thermobarometry using multi–equilibrium
calculations: A new technique, with petrological applica-

Byerlee, J., Model for episodic flow of high-pressure water in

Cox, S. F., Faulting processes at high fluid pressures: An
example of fault valve behavior from the Wattle Gully Fault, Victoria, Australia, J. Geophys. Res., 100, 12841–

Dahlen, F. A., Metamorphism of nonhydrostatically stressed

Davis, T. L., and J. S. Namson, A balanced cross-section of
the 1994 Northridge earthquake, southern California, 

Ellsworth, W. L., and G. C. Beroza, Seismic evidence for an

Etheridge, M.A., V.J. Wall, S.F. Cox, and R.H. Vernon,
High fluid pressures during regional metamorphism and
deformation: Implications for mass transport and de-
formation mechanisms, J. Geophys. Res. 89, 4344–4358, 
1984.

Fraser-Smith, A. C., A. Bernardi, P. R. McGill, M. E. Ladd,
R. A. Helliwell, and O. G. Villard, Low-frequency mag-
netic field measurements near the epicenter of the Ms 7.1
Loma Prieta earthquake, Geophys. Res. Letts., 17, 1465–
1468, 1990.

Hacker, B. R., Diagenesis and fault valve seismicity of

Hansshaw, B. B., and J. D. Bredehoeft, On the maintenance of anomalously fluid pressures, II. Source layer at depth,

Holloway, J. R., Igneous fluids, in Thermodynamic Modeling 
of Geological Materials: Minerals, Fluids, and Melts, Min-
eralogy Society of America Reviews in Mineralogy, 17, edited by I. S. E. Carmichael and H. P. Eugster, Chelsea, 

Hubbert, M. K., and W. W. Rubey, Role of Fluid Pressure 
in mechanics of overthrust faulting, I. Mechanics of 
fluid-filled porous solids and its application to overthrust 

Jennings, C.W., compiler Geologic Map of California, Cali-
ifornia Division of Mines and Geology, 1977.

Lasaga, A. C., and D. M. Rye, Fluid flow and chemical re-
action kinetics in metamorphic systems, Am. J. Sci. 293, 

Lees, J. L., and G. T. Lindley, Three dimensional attenuation 
tomography at Loma Prieta: Inversion of $t^*$ for Q, J. 

Nishiyanami, T., Kinetics of hydrofracturing and metamor-

Putnis, A. and T. J. B. Holland, Sector trillings in cordierite 
and equilibrium overstepping in metamorphism, Contrib. 

Rice, J. Fault stress states, pore pressure distributions and 
the weakness of the San Andreas Fault, in Fault Mechan-
ics and the Transport Properties of Rocks, edited by B.
Evans and T.-F. Wong, pp. 475–503, Academic Press, 

Ridley, J. and A. B. Thompson, The role of mineral kinet-
ics in the development of metamorphic microtextures, in 
Fluid–rock Interactions During Metamorphism, edited by 
J. V. Walther and B. J. Wood, pp. 154–193, Springer-
Verlag, New York, 1986.

Scholz, C. H., The Mechanics of Earthquakes and Faulting, 

Sibson, R.H., Implications of fault valve behavior for rupture 
nucleation and recurrence, Tectonophysics 211, 283-293, 

Smith, R. B., and W. J. Arhabsh, Seismicity of the Intermontane 
Seismic Belt, in Neotectonics of North America: Geol. Soc. Am. Decade Map, Vol 1, edited by D. B. Slem-
mons, E. R. Engdahl, M. D. Zoback, and D. D. Blackwell, 

Stein, R. S., and M. Lisowski, The Homestead Valley earth-
quake sequence, California: Control of athershocks and 
postseismic deformation, J. Geophys. Res., 88, 6477–6490, 
1983.

Stein, R. S., G. C. P. King, and L. Lin, Stress triggering of the 
1994 M=6.7 Northridge, California earthquake by its 

Tsunogai, U., and H. Wakaia, Precursory chemical changes in 
ground water: Kobe earthquake, Japan, Science, 269, 

Walder, J. and A. Nur, Porosity reduction and pore pressure 

Wegner, W. W. and W. G. Ernst, Experimentally determined 
hydration and dehydration rates in the system 

Wong, T., S. Ko, and D.L. Olgaard, Generation and mainte-
nance of pore pressure excess in a dehydrating system 2. 

Zhao, D., H. Kanamori, H. Negishi, and D. Weins, Tomog-
raphy of the source area of the 1995 Kobe earthquake: 
Evidence for fluids at the hypocenter? Science, 274, 1891–
1894, 1996.

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