THE MECHANICS OF METAMORPHIC FLUID EXPULSION

James A. D. Connolly (james.connolly@erdw.ethz.ch)
Dept. of Earth Sciences, Swiss Federal Institute of Technology
Claussiusstr 25, 8092 Zurich, Switzerland

ABSTRACT

Metamorphic devolatilization generates fluid and grain-scale porosity. Evidence for high fluid pressure indicates the devolatilization occurs under poorly-drained conditions. Under such conditions fluid expulsion is limited by the capacity of the reacted rocks to resist compaction or the rate at which deformation modifies the permeability of the overlying rocks. In the former case the compaction time scale must be greater than the metamorphic time scale and flow patterns are dictated by details of permeability. The alternative is that compaction processes are fast relative to metamorphism. In this case flow is compaction-driven and accomplished by waves of fluid-filled porosity.

INTRODUCTION

Typical crustal rocks lose 3-6 % of their weight during regional metamorphic devolatilization, a process that generates fluid and porosity at the expense of the solid volume (Fig. 1). These fluids are of interest because of their role in crustal rock mechanics, mineralization, and kinetics of other metamorphic reactions (Jamtveit and Austrheim, 2010). The flow of fluid generated by devolatilization is determined by the rates at which: fluid is produced, dilational deformation accommodates volumetric effects, and fluid is drained from the reacting rock. The classical view of metamorphism involving lithostatically pressured fluids implies a perfect balance between these rates and allows fluid flow in only one direction, toward the Earth’s surface (Walther and Orville, 1982). Such a balance is not only a mechanical impossibility (Connolly, 1997a), but is at odds with studies that infer a significant lateral component to
metamorphic fluid flow (Ferry and Gerdes, 1998). This paper reviews the relationship between fluid generation and expulsion and considers its implications for metamorphic flow patterns.

Phase equilibria (Holdaway and Goodge, 1990), fluid inclusions (McCuaig and Kerrich, 1998), and deformation styles (Etheridge et al., 1984, Simpson, 1998) testify that metamorphic fluid pressures are above the hydrostatic values common in the upper crust (Zoback and Townend, 2001). That these pressures are, at least sometimes, quantitatively lithostatic has been demonstrated by aseismic tremors in both subduction zone and continental settings (Peng et al., 2008, Scarpa et al., 2008). Because rocks have low tensile strength ($<\sim 5$ MPa), dilational deformation, e.g., brittle failure, provides an essentially instantaneous mechanism for regulating supra-lithostatic fluid pressures (Sibson, 1992). The mechanism for generating high fluid pressure is more complex, but admits two limiting cases: the rock pore are collapsed; or devolatilization pressurizes the porosity of a rigid rock. The miniscule grain-scale porosity of pristine metamorphic rocks (Norton and Knapp, 1977), geophysically observable densification (Hetenyi et al., 2007) and isotopic evidence that grain-scale fluid rock-interaction occurs on brief time-scales ($10^3$-$10^5$ years, Van Haren et al., 1996, Graham et al., 1998, Ague and Baxter, 2007) attest to efficient metamorphic compaction, but do not necessarily require that compaction occurs on the time scale of individual reactions.

Fluid expulsion is inseparable from rock deformation, which in turn modifies hydraulic properties, but the relevant hydraulic and rheologic rock properties are poorly constrained by laboratory data. As a supplement to this data, this paper begins by examining the order of magnitude constraints on crustal hydrology and rheology that follow from reasonable suppositions about the rate of metamorphism and the depth of the brittle-ductile transition. The paper then develops a conceptual one-dimensional model of fluid expulsion towards the earth’s surface, and concludes with discussion of a numerical model that illustrates the nature of lateral fluid flow in compacting rocks.
REGIONAL METAMORPHIC RATES

Prograde regional metamorphism is the wagging tail of a geodynamic dog. Unlike retrograde hydration and near surface processes, the rates and scales of prograde devolatilization are controlled by the energy input arising from the large-scale geodynamic process responsible for continental collision that thicken the continental crust (England and Thompson, 1984, Connolly and Thompson, 1989). Because the plate movements responsible for thickening are rapid relative to heat conduction, the thickened crust is undercooled relative to the steady state geotherm necessary to conduct mantle heat flow. Subsequent heating as the geotherm relaxes toward the steady state in combination with isostacy result in counterclockwise pressure-temperature paths for the metamorphic process. Heat conduction constrains the metamorphic time scale, $\tau_{\text{met}}$, to be $\sim l_c^2/\kappa$ where $\kappa$ is thermal diffusivity, $\sim 10^{-6}$ m$^2$/s for crustal rocks, and $l_c$ is crustal thickness. Thus a tectonic event that doubles crustal thickness to $\sim 70$ km is expected to generate metamorphism on 100 My time scale. This scale is consistent with the correlation between the thickness and age of crustal roots beneath mountain belts, which suggests metamorphic densification requires $\sim 300$ My to restore isostatic equilibrium (Fischer, 2002). For an initial continental geotherm of $\sim 15$ K/km, peak conditions of Barrovian metamorphism ($T$~600-700 C at 20-30 km depth) imply heating rates of $\sim 3$ K/My. In turn, these heating rates imply a metamorphic reaction front will advance at $\sim 200$ m/My.

Assuming a heat conduction controlled time scale and steady state vertical fluid expulsion, average metamorphic fluid fluxes are

$$q_0 = -\frac{w l_c}{\rho_f \tau_m} = -\frac{w \kappa}{\rho_f l_c} \approx -10^{-12} \text{ m/s}$$

for typical fluid densities ($\rho_f \sim 900$ kg/m$^3$) and volatile contents ($w \sim 0.06$ kg-volatile/kg-rock). The time scale for metamorphism in subduction zones and continental rifting is significantly shorter ($\sim 10$ My), but the kinematics of these settings are such that the average fluid flux is of the same order of magnitude as
for collision belts (Connolly, 1997a; Connolly, 2005). Time-averaged flux estimates derived from field studies are also in this range (Ferry and Gerdes, 1998). In detail, variability of the dehydration process (Fig 1) is sufficient to assure that metamorphic fluid production occurs within horizons of intense reaction bounded by non-reacting rocks that would limit drainage (Fig 2). For the specific geotherms illustrated in Fig 1, the vertically-integrated fluid production for a heating rate of 3 K/km corresponds to fluxes of \(~10^{-10}\) m/s, 2 orders of magnitude greater than the average flux, a result that demonstrates the temporal variability of devolatilization.

Petrologists periodically invoke advective heating by fluids and melts to explain anomalous heating relative to the classical model of England & Thompson (1984). For such fluxes to be significant the integrated fluxes must be comparable to the rock mass that is heated; thus it is implausible that the fluids can be generated by the metamorphism itself without extreme focusing (Connolly, 1997b). While these models may prove correct they leave open the troubling question of the flux source. It is noteworthy that deficiencies in the England & Thompson model with regard to temperature (Lux et al., 1986) and exhumation (Amato et al., 1999) can be explained mechanical effects, notably advective heating by the vertical displacement of large blocks of crustal material and both local and diffuse shear heating (Burg and Gerya, 2005) without substantially changing the metamorphic time scale. Deficiencies with regard to rate are more problematic and may ultimately require new tectonic models for regional metamorphism (Lister & Marnie, 2009), most prominently several lines of evidence suggest that type section for Barrovian metamorphism (Ague and Baxter, 2007; Dewey, 2005; Oliver et al., 2000) evolved 1-2 orders of magnitude faster than predicted by the conductive time scale.

HYDRAULIC PROPERTIES: PERMEABILITY AND POROSITY

Although the hydraulic properties of metamorphic fluids are reasonably well known and not strongly variable (Walther and Orville, 1982), the permeability and porosity of metamorphic rocks are poorly constrained. Indeed, it is not inconceivable that diagenetic processes are capable of eliminating all
hydraulic connectivity prior to metamorphism. Discounting this possibility, the permeability of metamorphic systems is usually derived by estimating the metamorphic fluid flux and pressure gradient (Manning and Ingebritsen, 1999). These estimates mask a dependence on both the rate of metamorphism and flow mechanisms which ultimately determine the fluid flux. This problem is avoided if the hydraulic regime of the lower crust is characterized by the flux necessary to maintain lithostatic fluid pressure rather than permeability. This flux defines a background state from which it is possible to assess the effect of local perturbations. Although this state is somewhat arbitrary, it is reasonable to expect that background fluxes are unlikely to exceed the average metamorphic fluid flux, which is dependent on the rate of metamorphism. Thus, for quantitative illustrations here, the background flux is equated to the average flux of $-10^{-12}$ m/s deduced above. The corresponding characteristic permeability is $k_0 \sim 10^{-19}$ m$^2$. While this permeability is low in comparison to permeabilities generally observed in situ in the upper crust ($10^{-13}$-$10^{-17}$ m$^2$, Ferry, 1994, Manning and Ingebritsen, 1999), they are unexceptional in view of the permeabilities of argillaceous sediments (Neuzil, 1994).

To a good approximation the background flux is a proxy for all hydraulic properties of a metamorphic system except porosity. The term porosity here includes any interconnected fluid-filled voids present on spatial scales that are much smaller than the scale for fluid flow, thus porosity includes both grain-scale porosity generated by densification during devolatilization as well as small-scale fractures induced by the consequent dilational deformation. This porosity is critical to understanding the dynamics of fluid expulsion because the hydraulic impact of metamorphic reactions is determined by how they influence permeability via porosity. Theoretical and empirical considerations (Neuzil, 1994; Wark and Watson, 1998; Xiao et al., 2006) indicate that permeability rises as a cubic or higher power of porosity. This relationship implies that the percent-level porosities generated by devolatilization reactions lead to order of magnitude increases in the permeability of the reacted rocks provided initial porosities are small, i.e., $<< 1\%$. 
On the basis of isotopic diffusion profiles, Skelton et al. (2000) infer background porosities in the range $\phi_0 \sim 10^{-3} - 10^{-6}$. These are consistent with grain-scale porosities of $10^{-3} - 10^{-6}$ measured in exhumed metamorphic rocks (Norton and Knapp, 1977). An upper bound on pre-metamorphic porosities of $\sim 10^{-2}$ is provided by the sensitivity of geophysical measurements, which generally do not indicate fluids in the lower crust except in active metamorphic settings.

**RHEOLOGY: THE BRITTLE-DUCTILE TRANSITION**

Elevated fluid pressure is commonly attributed to compaction in the ductile lower crust. This association is tenuous because the classification of the crust into an upper brittle regime and a lower ductile regime is based on its response to tectonic stress, whereas compaction occurs in response to the difference between pore fluid pressure and the mean stress. In fact, a compelling case can be made that the upper crust is characterized by hydrostatic pressures only because faulting maintains large scale permeability (Zoback and Townend, 2001). The absence of faulting in the aseismic lower crust allows processes, which include compaction, but also may include retrograde metamorphism and diagenetic processes, to eliminate large scale hydraulic structures. In the absence of such structures, the effective permeability of the crust would be limited by the vanishingly small permeability of argillaceous sediments (Neuzil, 1994).

Regardless of the significance of the brittle-ductile transition for fluid pressure, as temperature increases thermally activated time dependent compaction must become important. Current experimental models for compaction are so uncertain that they provide no practical constraints (Farver and Yund, 2000). Given this situation, an alternative is to calibrate ductile rheology in terms of the compaction time scale ($\tau_{B-D}$), formally the time to reduce porosity by $\sim 36\%$, at the depth ($z_{B-D} \sim 15$ km) and temperature $T_{B-D}$ ($\sim 623$ K) of the brittle-ductile transition via

$$\eta = \tau_{B-D} (\Delta \rho g z_{B-D})^n \exp \left( -\frac{Q}{RT} \frac{T - T_{B-D}}{T_{B-D}} \right)$$
where $\Delta \rho$ is the difference between rock and fluid density ($\sim 1900$ kg/m$^3$), $n$ is the stress exponent, $Q$ is the activation energy for viscous creep ($\sim 270$ kJ/mol, with $n=3$), and $\eta$ is the coefficient of viscous creep. From equation 1, for $\tau_{B-D} \sim 1$ My, the effective viscosity at the brittle-ductile transition is $\sim 10^{22}$ Pa s and decays to $\sim 10^{14}$ Pa s at 973 K.

**THE COMPACTION SCALES**

At near lithostatic fluid pressures, the stress that causes compaction cannot be related directly to depth, but time-dependent compaction processes develop on a natural length scale known as the compaction length (McKenzie, 1984). For crustal rheology (Connolly and Podladchikov, 2004) the compaction length is

$$\delta = \pi \left[ \frac{\eta k_0}{\phi_0 \mu (\Delta \rho g)} \right]^{n-1}$$

where $\mu$ is the fluid viscosity. In essence $\delta$ is the length scale over which pore fluids can move independently of compaction processes; thus it is intuitive that $\delta$ increases with rock bulk-strength ($\eta/\phi_0$) and the ease with which fluid can flow through it ($k_0/\mu$). Substituting Equation 1 into Equation 2, $\delta$ can be reformulated as

$$\delta = \eta^{\frac{n-2}{n}} \left[ \frac{\phi_0}{\phi_0 \mu (\Delta \rho g)} \right]^{n-1} \exp \left( \frac{Q T}{RT} - \frac{T_{B-D}}{T_{B-D}} \right).$$

Equation 3 is relatively insensitive to the parameter estimates discussed previously, but varies from $\sim 10^4$ to $\sim 10^1$ m with increasing temperature. This result suggests that at moderate temperatures compaction is likely to influence metamorphic flow patterns on observable spatial scales. The compaction time-scale for poorly drained rocks, $\tau_c = \eta (\Delta \rho g)^n$, is highly uncertain and only weakly related to the time-scale for compaction at the brittle-ductile transition ($\tau_{B-D}$), but its temperature dependence via equations 1 and 3 indicates metamorphic temperature variations are sufficient to cause a 10-fold increase in compaction rates with depth.
THE LIMITING FLOW REGIMES

Although it is widely accepted that fluid expulsion occurs during metamorphism, it is not widely appreciated that this process is mechanical and as such strongly dependent on rheology. To illustrate this dependence consider a minimal model for vertical flow in which: the fluid and rock are inelastic; the rock compacts viscously if the difference between the fluid pressure \( P_f \) and lithostatic pressure \( P \) is less than the tensile strength \( \sigma_y \sim 5 \text{ MPa} \); and the rock dilates plastically (i.e., fractures instantaneously) if \( P_f - P > \sigma_y \). The model can be further by discounting the volume change associated with devolatilization. While this effect is often attributed mechanical importance, in poorly drained systems it is largely irrelevant (Connolly 1997). Evidence for high fluid pressures during metamorphism requires that the metamorphic systems are poorly drained. Thus the essence of devolatilization is to produce a permeable horizon surrounded by impermeable rocks through which negligible fluxes are sufficient to generate lithostatic fluid pressure. Within the reacting layer, even if devolatilization involves a net volume increase, fracturing maintains near lithostatic conditions. Conservation of mass requires that in the absence of deformation the fluid flux must be equal to the drainage flux \( q_0 \) throughout the column. By Darcy’s law this flux is

\[
q_0 = -\frac{k}{\mu} \left( \frac{\partial P_f}{\partial z} - \rho_i g \right)
\]

if \( \frac{\partial P_f}{\partial z} \) is the lithostatic gradient \( (=\rho_i g) \). However, within the reacted zone permeability is \( \gg k_0 \), therefore the right hand term in equation 4 must be small, which is only possible if \( \frac{\partial P_f}{\partial z} \) is near the hydrostatic gradient \( (\rho_i g) \) regardless of the near lithostatic absolute pressures. This situation gives rise to a positive effective pressure gradient \( (~\Delta \rho g) \) that causes deformation (Fig 3).

The manner in which viscous compaction is superimposed on the foregoing scenario can be represented by the cases that the compaction time-scale \( (\tau_c) \) is much greater, or comparable to the metamorphic time-scale \( (\tau_m) \). For a constant volume devolatilization reaction, the mean fluid pressure within the rocks
behind a reaction front is identical to the mean total pressure. Thus the upper and lower halves of the reacted interval will be subject to negative (dilational) and positive (compactive) effective pressures. If $\tau_c >> \tau_m$, the rocks remain rigid on the time-scale of reaction until the vertical extent of the reaction is large enough (i.e., $2 \sigma_y/\Delta\rho/g < \sim 500$ m) to cause micro- or macro-scale fracturing at the top of the reacted rocks. Unless this fracturing produces fractures that breach the low permeability barrier formed by the overlying rocks, fracturing acts as homeostat that limits fluid pressure within the permeable zone as reaction progresses. Because the fracturing occurs at the top of the reacted rocks, the effect of continued reaction once the yield stress has been reached is to propagate fracture-generated porosity beyond the reaction front and lower fluid pressures. The propagation rate is dependent on the fracture mechanism, but because fracture permeability is also a cubic function of porosity it is unlikely that the fracture front propagates much more rapidly than the reaction front. The important feature of this limiting scenario is that metamorphic reactions generate a permeable horizon that has the potential to allow lateral fluid flow.

In the alternative scenario $\tau_c \sim \tau_m$ compaction squeezes fluid upward from the base of the reaction zone, while dilational processes at the top of the zone create porosity beyond the reaction front. If $\delta \Delta \rho g > \sigma_y$, this dilation may be accomplished by fracturing, but regardless of the dilational mechanism, the rate of dilation is limited by the rate at which devolatilization and compaction at depth supply the fluid that causes dilation. The combined effect of these processes is to propagate porosity upward relative to the reaction front. Because the rate of compaction at the base of the porous zone must increase with its vertical extent, compaction isolates the porous zone from the reaction front once the vertical extent of the reacted rocks is $\sim \delta$. The porosity then propagates upward independently of the reaction as a solitary wave. The essential features of this mode of fluid flow are that fluid pressures oscillate by $\sim \pm \delta \Delta \rho g$ about the lithostat; and fluid pressure gradients oscillate between hydrostatic and lithostatic. Since both the $\tau_c$ and $\delta$ are proportional to rock shear viscosity, the classical picture of metamorphism as an isobaric process is
recovered at high temperature when $\eta$ is low (i.e., $\tau_c<<\tau_m$ and $\delta\to0$), but porosity waves will slow and lengthen as they propagate upward into cool upper crustal rocks.

Compaction-driven fluid flow is widely appreciated in the context of both sedimentary basins and asthenospheric melting (Richter and McKenzie, 1984, Connolly and Podladchikov, 2000) and mathematical analysis has demonstrated that solitary porosity waves are a steady-state solution of the governing equations for fluid flow in compacting media (Barcilon and Lovera, 1989). This shows that the waves are stable provided the reaction-generated fluid flux is at least three times the flux that can be conducted through the unperturbed matrix. When this condition is not met, the transient evolution at the reaction front is unchanged, but the waves dissipate as they propagate into the overlying matrix (Connolly and Podladchikov, 1998). An additional requirement for the development of waves is that the vertical extent of the permeable source region must be $\sim\delta$. Both the localization of fluid production and the large resultant fluxes (Fig 2), suggest these requirements are met in metamorphic systems. Numerical analysis has shown that the 1-dimensional waves just described, decompose into elongate tube-like waves in 3 dimensions (Connolly and Podladchikov, 2007). Such waves, illustrated in the next section, do not substantially change the scales of compaction-driven fluid flow.

Chemical kinetic effects cause devolatilization to occur at higher temperature than predicted by equilibrium models, but they do not fundamentally change the equilibrium picture because chemically limited rates are proportional to the free energy change $\Delta G$ of devolatilization (Jamtveit and Austrheim, 2010). To a good approximation $\Delta G$ is related to the displacement in temperature ($\delta T$) and fluid pressure ($\delta P_f$, Dahlen, 1992) from equilibrium conditions by

$$\Delta G = \Delta V \delta P_f - \Delta S \delta T$$

where $\Delta S$ and $\Delta V$ are the entropy and volume change of devolatilization, typically $\Delta S = 3000\text{-}3500 \text{ J/K}$ and $\Delta V = -2 \cdot 10^{-5} - 8 \cdot 10^{-4} \text{ J/Pa per kg/volatile}$. These values imply that increasing temperature will increase
chemically limited kinetics rapidly until the process becomes heat supply limited. A variation in fluid pressure must be > 5 MPa to have the same effect as a 1 K change in temperature; thus during prograde metamorphism the effect of fluid pressure is to modulate the thermally controlled devolatilization rates. If fluid overpressure of ~5 MPa are sufficient to cause fracturing, in the non-compacting limit the effect of a devolatilization reaction with $\Delta V > 0$ is to reduce the vertical extent of reacted rocks necessary to induce fracturing. Counterintuitively, devolatilization reactions with $\Delta V < 0$ also induce fracturing, but the vertical extent of reacted rocks must greater than that deduced for the isochoric case.

**LOOKING FOR LARGE LATERAL FLUXES**

There is little doubt that lateral fluid flow occurs in metamorphic rocks (Ferry and Gerdes, 1998). Both lithological contrasts and reaction-generated porosities give rise to permeable horizons that promote lateral fluid flow, but in overpressured systems large lateral fluxes can only be explained by the existence of local drains into the permeable upper crust. It is conceivable that ephemeral brittle shear zones could function in this manner (Sibson, 1992), therefore to illustrate the interplay between compaction and lateral flow consider an initially poorly-drained reaction front that is punctured by a high permeability shear zone. To make this illustration less abstract, the fluid flow is modeled by solving the governing equations for compaction and heat flow numerically (Fig 5). The details of this model are described elsewhere (Connolly, 1997a) with the modifications that the matrix is weakened during decompaction to simulate fracturing (Connolly and Podladchikov, 2007) and random noise is added to the initial porosity to destabilize one-dimensional porosity waves (Fig 4).

As anticipated by the 1-dimensional scenario, in the non-compacting limit ($\tau_c >> \tau_m$) a small fringe of fracture generated porosity develops above the reaction front (Fig 5a), but other than this feature the model has no important non-kinematic behavior. With time the reaction creates an ever thicker permeable layer that conducts the reaction-generated fluid, as well as fluids flux from the adjacent rocks, to the fracture zone (Fig 5b). The pressure gradient within the layer necessary to drive lateral flow is
insignificant with the consequence that reaction front propagates uniformly upward. With time such a model evolves toward a steady state in which fluid pressures are hydrostatic above the reaction front and essentially all flow is focused into the shear zone. The flow pattern in this model is unsurprising and the main conclusion to be drawn from it is that the pattern is determined by uncertain initial conditions and kinematics.

In the compacting scenario, the shear zone drainage is less effective because compaction throttles lateral fluid flow and the shear zone must compete with drainage by tube-like porosity waves. The waves (Fig 5c) develop with a spacing comparable to the model compaction length ($\delta$=880 m), an effect that leads to focusing of reaction-generated fluxes. An unexpected feature of the flow pattern associated with the two-dimensional porosity waves is that the lateral and vertical fluxes are comparable. This convective pattern results because fluid is forced into surrounding matrix by high pressures at the top of the wave and drawn back into the lower under-pressured portion of the wave. This convective pattern is reminiscent of buoyancy-induced Rayleigh convection that develops in shallow hydrothermal systems (Norton and Knight, 1977). However, dimensional analysis (Connolly, 1997a) indicates that Rayleigh convection is unlikely in lower crustal metamorphic settings, a conclusion also be reached by more elaborate numerical modeling (Lyubetskaya and Ague, 2009).

In the compacting model, drainage by the shear zone suppresses the development of porosity waves. The lateral extent of this near instantaneous effect is quantitatively determined by the properties of the shear zone it decays rapidly as compaction seals the distal portions of the layer. This decay accelerates with time as the shear zone becomes a more effective drain for the portion of the layer with which it is in hydraulic contact. In the numerical simulation, these effects seal the shear zone from the reaction front within 60 ky (Fig 5c) and by 120 ky compaction has eliminated essentially all hydraulic contact with the shear zone. This latter effect has the consequence that subsequent devolatilization induced fluid flow occurs independently of the shear zone.
DISCUSSION

Regional metamorphism occurs in an ambiguous rheological regime between the brittle upper crust and ductile sub-lithospheric mantle. This ambiguous position has allowed two schools of thought to develop concerning the nature of metamorphic fluid flow. The classical school holds that metamorphic rocks are inviscid and that any fluid generated by devolatilization is squeezed out of rocks as rapidly as it is produced (Walther and Orville, 1982, Connolly and Thompson, 1989, Yardley, 2009). According to this school permeability is a dynamic property and fluid flow is upward. In contrast the modern school, selectively uses concepts from upper crustal hydrology that presume implicitly, if not explicitly, that rocks are rigid or, at most, brittle (Walder and Nur, 1984, Manning and Ingebritsen, 1999, Lyubetskaya and Ague, 2009). For the modern school, the details of crustal permeability determine fluid flow and as these details are poorly known almost anything is possible.

Reality, to the extent that is reflected by field studies, offers some support to both schools. In particular, evidence of significant lateral fluid flow (Ferry and Gerdes, 1998, Skelton et al., 2000) is consistent with flow in rigid media, while evidence for short ($10^4$-$10^5$ y) grain-scale fluid-rock interaction (VanHaren et al., 1996, Graham et al., 1998, Ague and Baxter, 2007) during much longer metamorphic events, suggests that reaction-generated grain-scale permeability is sealed rapidly by compaction; a phenomenon that is also essential to prevent extensive retrograde metamorphism. These observations provide a compelling argument for recognizing in conceptual models of fluid flow that metamorphic rocks are neither inviscid nor rigid, but have finite strength. The surprising consequence of this finite strength is that the steady-state solutions for fluid flow in porous compacting media require that fluid expulsion is channeled into waves of fluid-filled porosity. The waves develop on a characteristic length scale that is also the length scale for lateral fluid flow. In this context, porosity includes all hydraulically connected space present on a spatial scale $\ll \delta$. Thus, porosity waves may be manifest as self-propagating domains of fluid-filled fractures. Because $\delta$ is proportional to rock viscosity and consequently decreases exponentially with
increasing temperature, the flow regimes of the classical and modern schools are recovered at high and low temperatures.

The compaction-driven flow regime has been illustrated here under the assumptions that: compaction is time-dependent; decompaction is largely time-independent; and the far-field stress is isostatic. Near-surface sediments compact by time-independent plastic mechanisms that may well contribute to metamorphic porosity reduction. Fluid flow through porous media that compact dominantly by time-independent rheological mechanisms is also accomplished by porosity waves, but in contrast to the viscous case the waves have no intrinsic length scale (Connolly and Podladchikov, 1998, Miller et al., 2004). The assumption that decompaction occurs by fracturing is responsible for strongly channelized flow (Fig 5). If fracturing is suppressed, porosity waves in viscous rocks are equant at high temperature, but flatten as the waves propagate toward the surface (Connolly and Podladchikov, 1998). Fluid flow in compacting media is in the direction of low mean stress, in non-isostatic systems mean stress does not necessarily decay upward, an effect that may trap fluids beneath the tectonic brittle-ductile transition or draw fluids downward (Connolly and Podladchikov, 2004). In the presence of far-field stress, the Mohr-Coulomb failure criterion implies that dilational fracturing occurs at sublithostatic fluid pressures (Sibson, 1992). This effect would reduce fluid pressures and influence fracture patterns, but would not change the dynamics and scales of porosity waves limited by viscous compaction.

REFERENCES


FIGURE CAPTIONS

Fig 1. Water-content for average pelitic sediment (GLOSS, Plank and Langmuir, 1998) as a function of temperature and pressure computed assuming equilibrium with a pure H₂O fluid. This calculation indicates that more than half the initial mineral-bound water-content (7.6 wt %) is lost during diagenesis. The GLOSS composition includes 3 wt % CO₂ that is not incorporated in the model because decarbonation is dependent on fluid-rock interaction. Red and blue lines, respectively, indicate hot (20 K/km) and cold (10 K/km) metamorphic geotherms. The yellow-red-blue curve illustrates the typical clockwise depth-temperature path followed by rocks during collision belt metamorphism (England and Thompson, 1984). The tectonically controlled prograde burial segment (yellow) is rapid (~1 My); thermal relaxation, in conjunction with isostatic rebound, after burial (red) is slower (~10-100 My) and causes most prograde metamorphism; retrograde cooling (blue) does not affect the prograde mineral assemblages provided compaction isolates the mineralogy from grain-scale fluid-rock interaction.

Fig 2. Rates of metamorphic volume change with heating for metamorphic rocks (Fig 1) along hot and cold geotherms; in both cases fluid production occurs within restricted depth intervals. In the hot case, solid densification is insignificant and dilational deformation must create pore space for the fluid. Along the cold geotherm, solid densification creates much of void space necessary to accommodate fluid production, in fact, in the lower 6 km of the section the volume change of devolatilization is negative, i.e., solid densification creates more void space than necessary to accommodate the fluid at isobaric conditions. True volumetric production rates are the product of rate with respect to temperature multiplied by the metamorphic heating rate. For a metamorphic heating rate of 3 K/My, the vertically integrated fluid production generates steady-state fluxes of ∼10⁻¹⁰ m/s, 2 orders of magnitude greater than the average flux obtained assuming uniform production.
Fig 3. Conceptual model of metamorphic devolatilization neglecting minor elastic effects that quantitatively influence fluid pressure but have no important implications for its evolution (Connolly, 1997a). The reaction leaves a region of elevated porosity and permeability in its wake. Fluid flux is proportional to the permeability and the difference between the fluid pressure gradient and the hydrostatic gradient, thus the drainage flux through the lithostatically pressured overlying rocks is \( q \propto -k_0 \Delta \rho g \). In the absence of deformation, conservation of mass requires that this drainage flux must also be the flux within the reacted horizon with permeability \( k >> k_0 \); this is only possible if the difference between the fluid pressure gradient and the hydrostatic gradient is small. However, this near-hydrostatic fluid pressure gradient within the reacted rocks gives rise to an effective pressure \( (P_f - P) \) gradient of \( -\Delta \rho g \) so that pore fluids become increasingly underpressured relative to the lithostat with depth within the high porosity zone, and conversely increasingly overpressured toward the reaction front. The resultant effective pressures are the driving force for deformation and fluid expulsion.

Fig 4. Time evolution of reaction-generated porosity and fluid pressure profiles for the non-compacting \( (\tau_c >> \tau_m) \) and compacting \( (\tau_c \sim \tau_m) \) scenarios, for each profile the baseline is indicated by a vertical dotted line. The baselines for the porosity and pressure profiles correspond, respectively, to the background porosity \( \phi_0 \) and lithostatic pressure. For purposes of illustration it is assumed that dilational deformation, in the form of microscopic or macroscopic fracturing, is instantaneous if fluid overpressure exceeds tensile strength (Fig 4). The magnitude of the fluid pressure anomaly within the reaction-generated porosity is proportional to the vertical extent of the high porosity zone, thus as the reaction progresses the anomaly grows until it becomes large enough to cause deformation. In the non-compacting case at \( t=1 \), fluid pressure has just reached the failure condition. Thereafter failure acts as a homeostat requiring that any advance of the reaction front is accompanied by propagation of fracture porosity, an effect that lowers fluid pressure at the reaction front. Even in the unlikely event that such fractures should become self-propagating (Rubin, 1998), they are not a mechanism for draining the reaction-generated porosity. In the
compacting scenario, compaction squeezes fluid upward providing an independent mechanism for maintaining high fluid pressures that cause dilational deformation above the reaction front, an effect that ultimately propagates the porosity beyond reaction front. Once this occurs ($t=3$), the porous domain propagates independently of the reaction as a solitary wave of anomalous porosity (Richter and McKenzie, 1984; Connolly, 1997a).

Fig 5. Numerical simulation of the influence of a permeable ($10^{-17}$ m$^2$) shear zone on devolatilization-induced fluid flow for non-compacting ($\tau_c>>\tau_m$; a, b) and compacting ($\tau_c\sim\tau_m$; c, d) scenarios. The plots of porosity, fluid pressure, and the magnitude of the vertical and horizontal components of the fluid flux are for a 24 km wide segment of the model spatial domain, which represents a 20x40 km crustal section. In the plots of vertical flux magnitude, large domains in which flow direction is predominantly downward are bounded by white curves, smaller domains of downward flow associated with individual porosity waves are not indicated. Prior to shear zone emplacement at $t=0$, devolatilization proceeds for ~100 ky creating a 100 m thick permeable horizon overlain by a ~200 m wide fringe of fracture-generated porosity (as in Fig 4 for $\tau_c<<\tau_m$). At $t=10$ ky both scenarios are virtually identical, the surge of fluid into the shear zone causes extraordinary fluid pressures and fluxes, and the consequent lowering of fluid pressure within the reacted horizon locally accelerates devolatilization. The non-compacting scenario rapidly reaches a quasi-steady state in which negligible pressure gradients are adequate to drain fluid from both within and about the reacted layer. In contrast in the compacting scenario, by $t=60$ ky the active portion of the reaction front is drained by tube-like porosity waves and is completely isolated from the shear zone. By $t=120$ ky compaction has also eliminated the residual porosity in the inactive portion of the reaction zone; thus when dehydration resumes the resulting flow will be independent of the shear zone.
Connolly, Fig. 1

Metapelite $H_2O$ content

Weight % $H_2O$

Temperature, °C

Depth, km
Connolly, Fig 2

- Cold, \( q_m = -2.2 \times 10^{-10} \) m/s
- Hot, \( q_m = -1.3 \times 10^{-10} \) m/s

Depth, km

Volume production, m³/K
Porosity

Depth

Fluid Pressure

\[ \frac{\partial \rho}{\partial z} = \frac{\rho g}{\gamma} \]

reaction front
dilation
compaction

reaction front

heat source

lithostatic pressure + tensile strength (\(\sigma_v\))

hydrostatic gradient, no fluid flow

lithostatic, no deformation

Connolly, Fig 3
Connolly, Fig 5

(a) non-compacting, \( t = 10 \) ky

(b) non-compacting, \( t = 110 \) ky

(c) compacting, \( t = 60 \) ky

(d) compacting, \( t = 110 \) ky