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Migration of metamorphic fluid: some aspects of mass and heat transfer

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ABSTRACT

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Minute amounts of free fluids contained along grain boundaries or in fluid inclusions in the rocks of the lower continental crust, mostly remain trapped until dislodged by thermally-induced hydrofracturing, deformation, or tectonic uplift. Volatile-components of minerals in the lower continental crust, are released at rates depending upon the changes in metamorphic heat-supply during tectonic processes, and at depths depending upon appropriate petrogenetic grids at near lithostatic pressure ($P_{fluid} \sim P_{rock}$). Pervasive flow of metamorphic fluid does not result in significant heat advection nor metasomatic mass-transfer. Layers of different permeability, strength and fertility (in the sense of amounts of metamorphic fluid released from minerals), control the local metamorphic hydrology. Only by focussing fluid-flow, by at least 10^3, can heat and mass advection become significant. The most favourable geological mechanisms to focus fluid-flow seem to be the reactivation of inherited heterogeneiities such as lithological contacts, old faults and shear zones, close to a deep-crustal aquitard or a fertile layer.

1. INTRODUCTION

Fluid penetration from the surface is mostly restricted within the upper crust. However, tectonic burial and subduction of sedimentary rocks and hydrated oceanic crust can transport mineralogically-bound volatiles to lower crustal and upper mantle depths. Likewise, magmatism adds volatiles to the lower crust from the mantle. Our purpose here is to examine physical constraints on the length- and time-scales of metamorphic fluid migration in the lower continental crust. We try to identify geological circumstances where heat and mass advection by metamorphic fluid could have been significant. This requires information on the likely ranges for rock-permeability and sizes of lower-crustal fluid reservoirs. Some fluids can remain locked in disconnected grain-boundary networks (e.g. fluid-inclusions); others will migrate upwards along transiently connected grain-boundary networks, due to buoyancy forces. Fluid release from fluid-inclusions and from metamorphic minerals is episodic and depends upon deformation rates and mechanisms, and the evolving metamorphic heat supply. We specifically wish to separate metamorphic events in the lower crust that do not require involvement of mantle heat and fluids from those that do. We also need to consider which aspects of metamorphic hydrology require local rather than regional explanation. It is to be expected that metamorphic fluid migration processes need not be identical (in the sense of efficiency of mass and heat advection), in quite different tectonic settings – continental collision zones, continental extension zones, oceanic speading centres, subduction zones and accretionary-wedge complexes.

We have tried to identify which areas in the topic of metamorphic hydrology have developed since the 1978 publication of *Fluids in the Earth's Crust* by Fyfe, Price and Thompson. We have attempted to detail where outstanding problems remain, and to suggest some ways in which partial solutions may be found. While metamorphic fluid migration is a transient phenomenon, much can be learnt about lengthscales and timescales of fluid advection processes by considering the simplifications permitted by steady-state considerations.

2. SOURCES OF METAMORPHIC FLUIDS

Meteoric waters and sea water penetrate the upper crust down cracks, but presumably not deeper than the "rheological transition zone", which separates brittle upper-crust from plastic lower-crust. This zone is located somewhere between 10 and 15 km in crystalline continental crust, but its position will vary depending upon lithology, temperature and tectonic age (e.g. Sibson, 1983; Nesbitt and Muehlenbachs, 1989).

The main source of metamorphic fluid is structurally-bound volatiles in minerals and in fluid-inclusions in metasediments and altered magmatic rocks (Newton, 1989). In some cases such rocks have been tectonically buried in collision or subduction zones to more than 100 km, a depth deduced from coesite occurrences in metasediments and blueschists (e.g. Chopin, 1984). Such volatiles remain stored in metamorphic minerals until released by thermal or tectonic perturbations.

Many mantle-derived magmas transport dissolved volatiles and, during cooling can release such fluids at their solidus. It is necessary to attempt to distinguish whether magmatic fluids represent dehydration-melting of the mantle magma source region, or whether they could have been crustal metamorphic fluids adsorbed by hygroscopic magmas during their rise through the crust.

Fluid inclusions in minerals could reflect any of these sources – depending when they became trapped in the burial-uplift history of continental fragments.

3. LOWER CRUSTAL HYDROLOGY

Fluid motion in the upper crust is reasonably well studied (e.g. see the recent summary by Cathles, 1990). Many thermal springs reflect heat advection caused by rapid fluid-flow up fractures – even from a "normal" geothermal gradient. Fluid motion in fault zones can often be related to earthquake activity (Sibson, 1987; Gold and Soter, 1985; Segall, 1989). Convective hydrothermal systems are mostly driven by near-surface magmatic intrusions. They are often the locations of hydrothermal mineral deposits when fluid flow has become focussed into fault/fracture systems (e.g. Fyfe and Kerrich, 1985).

However, on account of the distinct weakening of geological materials in rheological response to increased temperature, the upper and lower continental crust represent fundamentally different hydrological regimes (Fig. 1). The upper crust being cooler is on average brittle, and able to sustain fluid-filled cracks to some depth. Thus upper crustal fluid pressure can be approximated by a "hydrostatic" gradient, with $P_{\mathrm{fluid}} \sim 1/3P_{\mathrm{rock}}$.

In view of normally higher temperatures in the lower crust, the rheological response at the higher confining pressures is plastic, at least at geological strain-rates (e.g. Sibson, 1983; Kirby, 1984, 1985). On average the lower crustal porosity is not connected, thus fluids here are transient and fluid pressures approach those of a near lithostatic gradient ($P_{\mathrm{fluid}} \sim P_{\mathrm{rock}}$). The transitional zone between these distinct hydrological regimes (Fig. 1) will lie at different depths in continental crust (~ 3 km for indurating sediments to > 15 km for crystalline rocks); depending upon the interrelated entities – tectonic age, geothermal gradient, and the rheological behaviour of the specific minerals in a local rock-type.

Metamorphic fluid located in fluid inclusions, pore space, grainboundary films, or microcracks is probably disconnected except transiently. Despite this disconnected porosity, an upward fluid flux is still maintained by two-phase flows of buoyant fluid through the lowporosity rock matrix. Fluid can be redistributed by volumetric changes of mineral reactions, and by local deformation that changes potential pore-volume and hence fluid pressure (Fischer and Paterson, 1989). Fluid distribution in the lower crust also depends upon where and when fluids are being released by mineral reactions, or from crystallising magma.

Deep crustal fluids will flow pervasively upwards, albeit very slowly; and occasionally this flow becomes focussed resulting in lower crustal veins and mineral deposits, and perhaps small (< 10 km^2) metamorphic thermal anomalies. We need to consider therefore two distinct hydrothermal regimes in the lower crust – the low-permeability porous media and the higher-permeability focussed flow channels, and then the transitional processes connecting them.

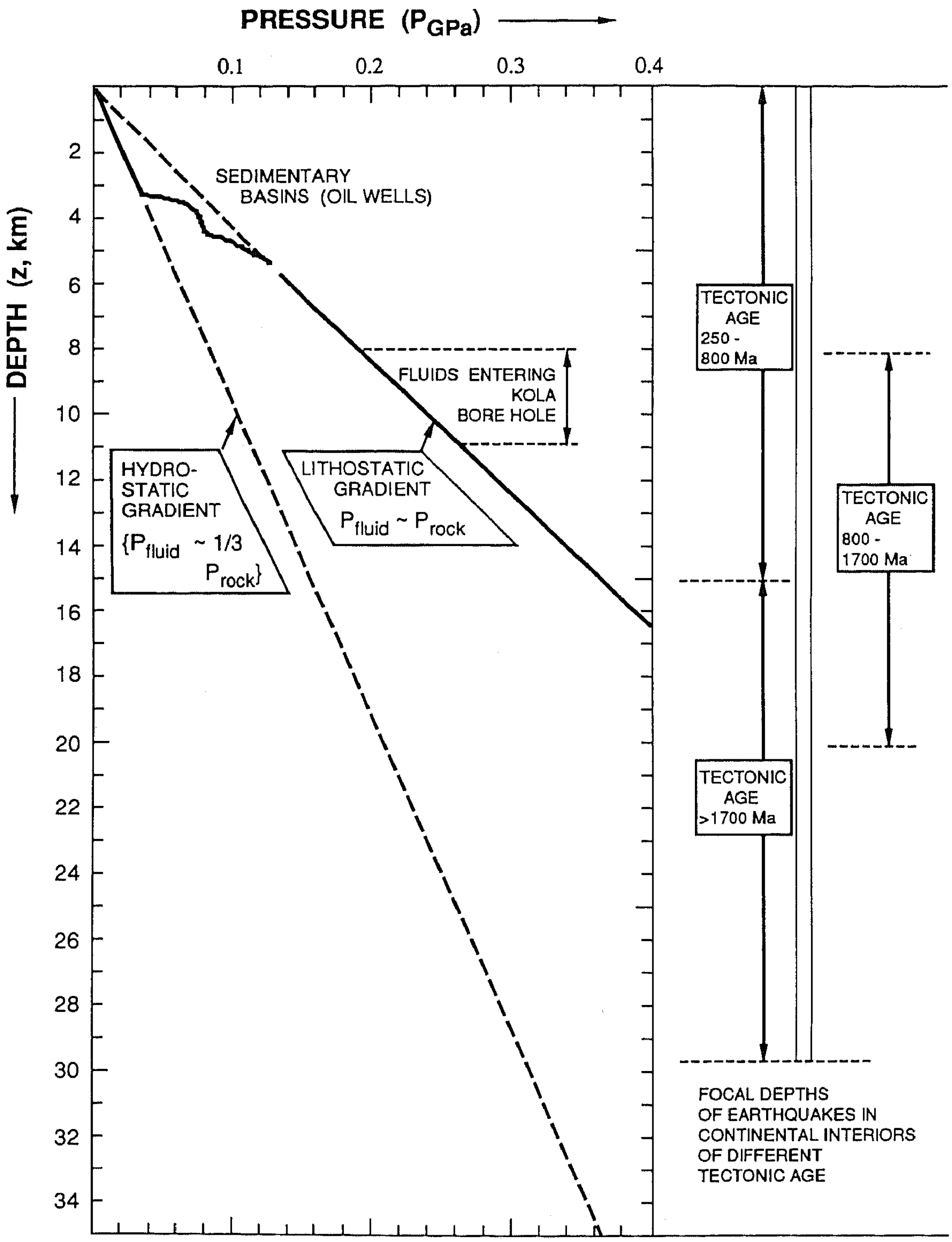

Fig. 1. Pressure–depth relations for litho- and hydro-static gradients. The transition zone observed for oil-wells in sedimentary basins is from Fyfe et al. (1978) and Wood and Walther (1986); for fluids entering the deep Kola Borehole from Kozlovsky (1987). The focal depth ranges of earthquakes in continental interiors of different tectonic-age is from Chen and Molnar (1983) and show deeper brittle behaviour in older and colder lithosphere.

4. GEOCHEMICAL AND PETROLOGICAL SIGNATURES OF METAMORPHIC FLUID

Several observations from metamorphic hydrology suggest the occurrence of deep crustal aquifers and aquitards. Petrographic inferences, supplemented by stable isotope analyses, have revealed that specific rock layers have acted as flow-concentrators of metamorphic fluid (Rumble and Spear, 1983; Ferry, 1987; Rye and Bradbury, 1988; Rumble, 1989). Lithlogical contacts and previous faults, or shear zones, have acted as higher permeability channelways. Most of these studies have concentrated on rocks metamorphosed at mid-crustal depths (<15 km). Results from deeper crustal levels have proved more difficult to interpret because of uncertainty of the timing of the fluid-rock interaction events in the (usually) polymetamorphic history (Kerrich, 1986; Valley, 1986; Frueh-Green, 1987).

Magma generation by crustal anatexis requires, in some cases, ingress of external fluid (e.g. Wickham and Taylor, 1987), and is thus additional evidence for transient fluid in the lower crust.

5. FLUID GENERATION DURING REGIONAL METAMORPHISM

Fluid is released from hydrous minerals at a rate depending mainly upon the metamorphic heat supply and to a lesser extent upon reaction and deformation kinetics (e.g. Ridley, 1985). Tectonic processes that change crustal thickness (such as thrusting or homogeneous thickening, and erosional or tectonic uplift), induce thermal perturbations which in turn cause fluid release from fertile rocks (fertile in the sense of amount of mineralogically-bound fluid). The timing of this release occurs when the pressure-temperature-time (PTt) path passes through the P-T conditions for the relevant devolatilisation reactions. Timing varies with depth and lithology in the tectonically thickened- or thinned-crustal column (Connolly and Thompson, 1989).

Walther and Orville (1982) calculated the advective behaviour of an average metamorphic fluid released by devolatilisation reactions in the metamorphic pile. Consider a crustal column 1 $m^2 \times 25$ km and assume that 5 wt% volatiles are released. Such a column therefore contains about 72×10^6 kg rock and thus 3.6×10^6 kg H_2O (3.6×10^3 m^3 H_2O). Over 20 Ma the integrated fluid flux rate (q_f) is about 180 $m^3\,m^{-2}\,Ma^{-1}$ ($= 5.8 \times 10^{-12}$ $m^3\,m^{-2}\,s^{-1}$) and over 100 Ma, $q_f \sim 36$ $m^3\,m^{-2}\,s^{-1}$ ($\sim 1.1 \times 10^{-12}$ $m^3\,m^{-2}\,s^{-1}$) and over 100 Ma, $q_f \sim 36$ $m^3\,m^{-2}\,s^{-1}$ ($\sim 1.1 \times 10^{-12}$ $m^3 m^{-2}\,s^{-1}$). Because the actual metamorphic heat supply is strongly controlled by heat advection during changes in crustal thickness, and the boundary condition of $T_{surface}$ being close to $0\,°C$ it is thus necessary to consider the ways in which the metamorphic heat supply controls fluid flux.

Consider continental crust doubled by thrusting at 35 km (see fig. 3e of England and Thompson, 1984; or fig. 2b of Connolly and Thompson, 1989). Very roughly, 25 km of the lower plate is heated from below $400\,°C$ to over $600\,°C$ during the first 20 Ma following thrusting (see Brady, 1988: p. 1197). For these selected time and length scales and temperature ranges, the zero- and one-dimensional estimates coincide – giving a metamorphic dehydration flux rates (q_f) of order 10^{-11} to 10^{-12} $m^3\,m^{-2}\,s^{-1}$ (see also Peacock, 1989). It is appropriate to consider what such low fluid flux rates might mean in terms of possible lower crustal hydrology, and the efficiency of advected fluids for heat and mass transfer.

6. POROUS-MEDIA FLOW OF METAMORPHIC FLUID

Pervasive (porous-media) flow continuously drains fluid released during devolatilisation reactions in the lower crust. Assuming that Darcy's law is valid [$q_f = -(K_\phi/\mu).\partial P/\partial z$], free fluid will migrate upwards due to buoyancy with a flux depending mainly upon rock permeability (K_ϕ). Lower crustal permeabilities are unknown, but it is expected, on the basis of laboratory experiments and in-situ well tests (Brace, 1980) that they range from 10^{-17} m^2 to less than 10^{-23} m^2. Permeability (K_ϕ) is proportional to porosity (Φ) raised to a high power (^{3} to ^{5}, Dullien, 1979).

For a fluid flux rate (q_f) of 10^{-11} to 10^{-12} $m^3\,m^{-2}\,s^{-1}$, it is possible to estimate the average equivalent Darcian permeability (K_ϕ, m^2) for the crust. For the optimum driving pressure gradient with fluids at lithostatic pressure, $\partial P/\partial z = \Delta\rho\, g = (2800\text{–}900\ \mathrm{kg\,m^{-3}}) \times (9.81\ \mathrm{m\,s^{-2}})$. With a value for fluid viscosity (μ) of 1.5×10^{-4} Pa s ($\mathrm{kg\,m^{-1}\,s^{-1}}$), then $K_\phi = 4.6 \times 10^{-20}\ m^2$. Permeabilities of 10^{-17} m^2 and 10^{-23} m^2, lead to respective Darcian metamorphic dehydration fluxes (q_f) of order 10^{-9} to 10^{-15} $m^3\,m^{-2}\,s^{-1}$ (Fig. 2). Several authors (e.g. Etheridge et al., 1984; Brady, 1988) consider a reference crustal permeability of $K_\phi = 10^{-18}$ m^2, with a corresponding Darcian fluid flux (q_f) of 1.3×10^{-10} $m^3\,m^{-2}\,s^{-1}$. This reference fluid flux rate is one to two orders-of-magnitude higher than the dehydration fluid-flux rate regulated by the regional metamorphic heat supply.

7. HEAT-ADVECTION DURING POROUS-MEDIA FLOW

Fluid advected heat can be viewed locally as heat-production and, to determine its significance in regional heating, can be com-

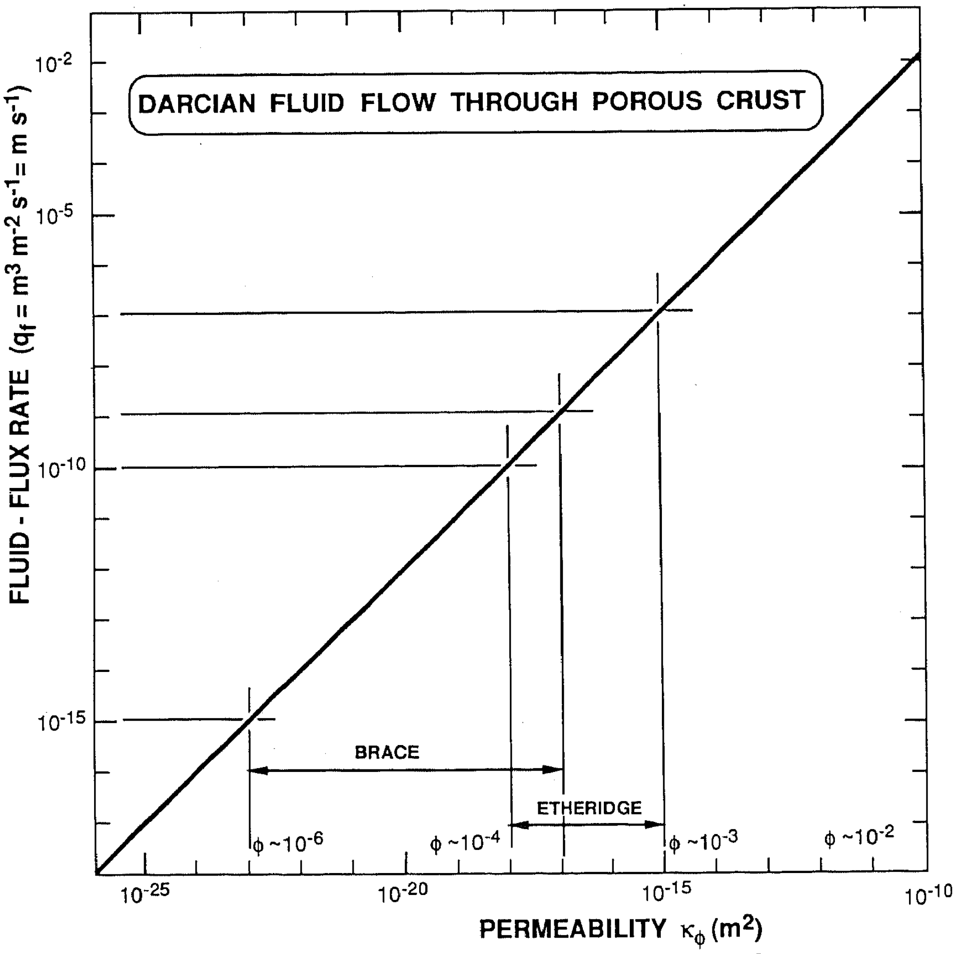

Fig. 2. A plot of Darcian fluid-flux, q_f ($m^3\,m^{-2}\,s^{-1}$), against permeability, K_ϕ (m^2), for constant $\mu = 1.5 \times 10^{-4}$ Pa s, and $\partial P/\partial z = \Delta\rho g = 1.9 \times 10^4$ $\mathrm{kg\,m^{-2}\,s^{-2}}$. Laboratory experiments and in-situ well tests Brace (1980) give K_ϕ values that range from 10^{-17} m^2 to less than 10^{-23} m^2. Etheridge et al. (1984), consider crustal permeability of $K_\phi = 10^{-15}$ m^2 to 10^{-18} m^2, to be more appropriate. The reference permeability of 10^{-18} m^2 corresponds to a fluid-flux rate of about 10^{-10} $\mathrm{m\,s^{-1}}$.

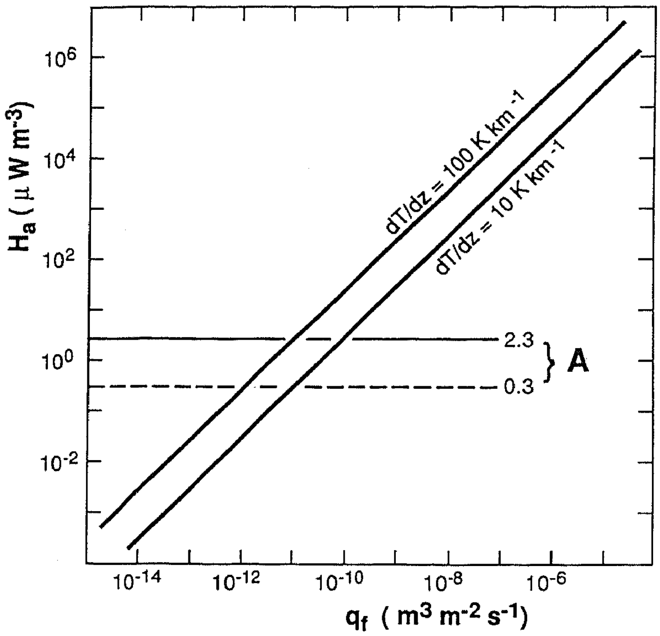

Fig. 3. Heat production (Ha, μwm^{-3}) from advected fluid for various fluid-flux rates (q_f, m^3 m^{-2} s^{-1}) for $\rho Cp = 3.5 \times 10^6$ J m^{-3} and temperature gradients ($dT/dz = 10, 100°$ km^{-1}). Radiogenic heat production for crustal rocks lies in the range 0.03 to 2.3 μW m^{-3}. Values of q_f in excess of 10^{-10} m^3 m^{-2} s^{-1} are needed for advective heat production to exceed radiogenic heat production.

pared with radiogenic heat production – a low-level long term heat source in the crust.

Any fluid flow up a positive temperature gradient will advect some heat (e.g. see Fig. 6 by Jamtveit et al., 1990: p. 191). The amount of heat advected (Ha, W m^{-3}) by metamorphic fluid flow through a porous medium can be evaluated from the product of fluid-flux rate (q_f, m^3 m^{-2} s^{-1}), fluid heat capacity (ρC_p, ~ 3.5×10^6, J m^{-3} K^{-1}) and geothermal gradient (dT/dz, K m^{-1}). Figure 3 shows the amount of advected heat for limiting geothermal gradients of 10 K km^{-1} (0.01 K m^{-1}) and 100 K km^{-1} (0.1 K m^{-1}) for a range of fluid-flux rates, compared with that from radiogenic heat production from crustal rocks (0.3 to 2.0 μW m^{-3}, e.g. England and Thompson, 1984). It can be seen that fluid fluxes in excess of 10^{-10} m^3 m^{-2} s^{-1} are required for heat produced by fluid advection to be more significant than radiogenic heat production. The figure also shows that the result is not particularly dependent upon the magnitude of the geothermal gradient.

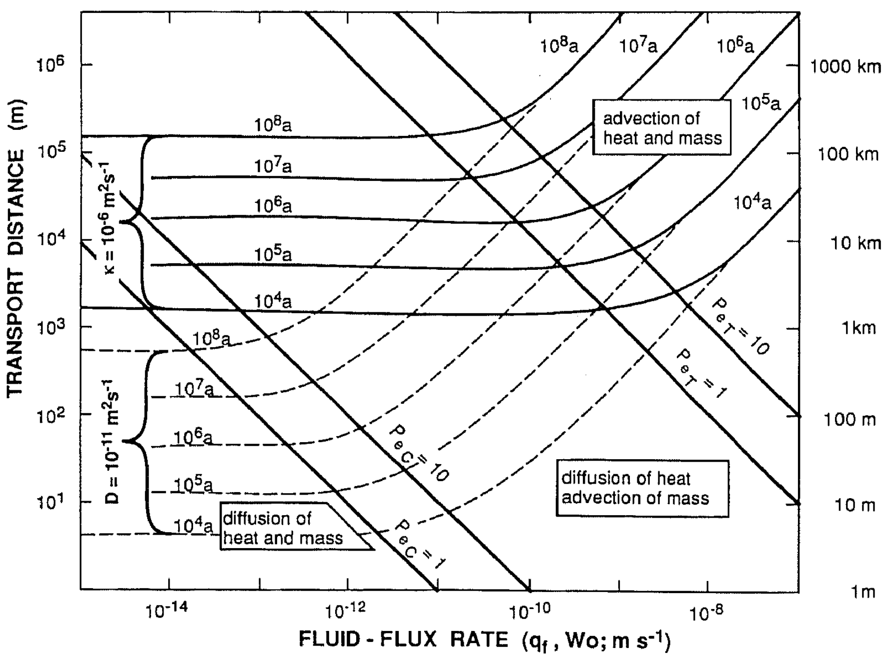

Fig. 4. Transport distances for heat and mass transfer as a function of fluid-flux rates, for advection events of 10^4 a, 1 Ma and 10^8 a duration (modified from Bickle and McKenzie, 1987: fig. 6). Simple thermal and chemical Peclet numbers are defined as $Pe_T = zW_0/\kappa$ and $Pe_C = zW_0/D$, where z is lengthscale (m), W_0 = fluid velocity (m s^{-1}), κ is a typical thermal-diffusivity (10^{-6} m^2 s^{-1}), and $D = 10–11$ m^2 s^{-1} is a sample chemical-diffusivity (10^{-8} m^2) for a porosity of 10^{-3}. Such Peclet numbers are smaller by a factor of π^2 (~ 10) than those defined by Bickle and McKenzie (1987: fig. 6).

These considerations indicate that normally, lower crustal single-pass fluid advection will not increase rock temperatures by more than a few degrees, unless vast quantities of deep crustal fluid can somehow become flow-focussed in quite short times (Brady, 1988; Peacock, 1989). In general, fluids are probably only important transporters of heat in the upper crust where convective recirculation can occur using faults/fractures as flow concentrators, or in the vicinity of cooling water-rich magmatic bodies (e.g. Cathles, 1990). However, as can be seen from Fig. 4 when large fluixes do occur, advection is an extremely efficient process for heat and mass-transfer (Bickle and McKenzie, 1987). Obviously, more rapid advection leads to earlier exhaustion of the fluid reservoir. The steady-state picture in Fig. 4 can easily be rescaled to simplify any modelling of transient processes on different timescales and lengthscales.

8. FLUID CONVECTION IN THE LOWER CRUST

Free convection of fluid in a continuously porous medium heated from below can occur depending upon the value of the critical permeability, K_c. For example, Straus and Schubert (1977) showed that:

$$K_c = 4\pi^2 \nu\kappa / g\alpha\Delta T d$$

For $\nu = 5.5 \times 10^{-8}$ m^2 s^{-1}, $\kappa = 10^{-6}$ m^2, $\alpha = 10^{-3}$ K^{-1}, $g = 9.81$ m s^{-2}, $\Delta T = 600\,°$C and $d = 30$ km (values chosen by England and Thompson, 1984: p. 922, to illustrate the case of whole-crustal fluid-convection), a value of the critical permeability, $K_c = 10^{-17}$ m^2, is required for free convection of pore fluids in a 30 km thick crust with a "normal" geothermal gradient (20° C km^{-1}). This permeability value is several orders-of-magnitude greater than those deduced from metamorphic Darcian fluid fluxes. It is to be concluded that free convection can only occur in the lower crust if rocks are able to sustain enhanced permeability, e.g. by keeping connected fluid-filled pores or fractures open at these depths. Turcotte and Schubert (1982: fig. 912, p. 405) have shown the dependence of K_c upon $\partial T/\partial z$ as a function of the thickness of a layer with continuously-connected porosity in which convection can occur. The inverse proportionality of K_c and d means that the K_c for thinner layers is much larger than for thicker layers.

If the lower crust behaves plastically there is only transient connected porosity, very little possibility for fluid downflow and hence no large-scale convection (Yardley, 1986; Wood and Walther, 1986: p. 92; Valley, 1986). However, for short-term rigid rheological rock behaviour, it been proposed that fluid convection cells could only develop in isolated permeable layers where locally the fluid pressure is hydrostatic, sealed above and below by impermeable layers in which the fluid pressure is near-lithostatic (see Guzzetta and Cinquegrana, 1987; also Yardley (1986: p. 123) for an example for crystalline rock; and Cathles (1990: fig. 3) for an example from a sedimentary basin). This would require quite unusual hydrological behaviour in the crust.

Density variations in H_2O-CO_2 fluids is small even over the range of lower crustal pressures and temperatures. However, the brines found in deep crystalline rocks (Fritz and Frape, 1982, 1987) will exhibit critical-phenomena extended to much high pressures and temperatures than for H_2O-CO_2 alone. Thus fluid unmixing into NaCl-H_2O brines coexisting with CO_2-rich vapour, should be common in the lower continental crust. This higher pressure "boiling" could induce widespread mineral precipitation in lower-crustal hydrothermal veins, and the consequent density variations in unmixed brines is likely to influence local fluid migration especially for perturbed geotherms in continental crust.

Forced convection of fluids could occur in the lower crust due to fluid pressure gradients. Such situations are likely to be transient and small-scale, e.g. beneath thrust-faults. In any case confined convection is not likely to induce much crustal-scale heat transport.

9. MASS-TRANSPORT DURING POROUS-MEDIA FLOW

Transport of dissolved material by advecting fluids requires high fluid fluxes, in view of the low solubility of minerals in aqueous fluids at pressures below about 1.0 GPa. Because of the enormous range of diffusivities of various species in aqueous solutions, there will be corresponding differences in metasomatic profiles across or along rock layers (e.g. Bickle and McKenzie, 1987). Anomalous profiles of certain elements or isotopes have been used to show where fluid infiltration occurred (Baumgartner and Rumble, 1988; Baker et al., 1989, Harris and Bickle, 1989), but cannot give actual estimates of the quantities of fluid that have flowed.

One example of mass-transport regimes is shown in Fig. 4, where the relative importance of diffusion versus advection of heat or mass is clearly illustrated for a wide-range of fluid flux rates. Transport distances are shown for geological time-scales for an "effective" mass-diffusivity of about 10^{-11} m^2 (the value used in Bickle and McKenzie, 1987, fig. 6; i.e. mass-diffusivity of 10^{-8} m^2 and porosity of 10^{-3}). Mass-transport for other values of diffusivity can be easily seen by combining fig. 4 with fig. 1 of Rubie and Thompson (1985: p. 37). Because the thermal diffusivity is typically orders of magnitude greater than mass diffusivity, there is a large region of fluid-flux rates and length scales, in the middle of Fig. 4, when diffusive mass transfer occurs in the absence of advective thermal perturbations (Bickle and Mckenzie, 1987: p. 390).

10. FLUID–ROCK RATIOS IN METAMORPHIC ROCKS

So called "fluid–rock ratios" have become popular parameters to evaluate, in attempts to assess the extent of metasomatic mass transport as a widespread compared to localised phenomenon in metamorphic rocks. Such ratios are commonly deduced from extent of progress of metamorphic reactions (Ferry, 1986; Wood and Graham, 1986), or from measured shifts in stable isotope ratios in coexisting minerals (e.g. Rumble, 1989; Valley, 1986). Typical values so deduced are about 1 to 10. Considerations of the amount of minerals in hydrothermal veins and the very low solubility of minerals in natural fluids, imply fluid–rock ratios of the order 10^4–10^5. These discrepancies reflect the ambiguous nature of the concept of fluid–rock ratios, mainly because time-integrated fluid-fluxes are often incorrectly equated with these ratios (see Connolly and Thompson, 1989: p. 359; Harris and Bickle, 1989: p. 155).

Extremely large fluid–rock ratios, up to 250 000, can develop as a consequence of dehydration metamorphism, simply because rocks higher in the column experience passage of fluids migrating from beneath (Connolly and Thompson, 1989: fig. 9). It is useful to consider the relationship between petrologic fluid–rock ratios (F) and time-integrated fluid-fluxes ($q_f = F d \rho_f / t$) for fluid flowing across, or along, a layer of distance d. As fluid flows across this section, either d or F increases. The time to alter one unit-volume of rock is t/d. For an average value of $q_f = 5 \times 10^{-13}$ $m^3 m^{-2} s^{-1}$ and $\rho_f = 1000$ $kg m^{-3}$, for $t = 50$ Ma, for values of d of 1100 and 1000 m, volumetric fluid–rock ratios, $F = (q_f t / d \rho_f) = 750, 75, 0.75$, are obtained respectively.

However, even in systems with these large fluid–rock ratios, the single pass equilibrium fluid flow likely in the lower crust can precipitate only small amounts of dissolved material e.g. silica (< 0.1 vol.%) in quartz-bearing rocks (Connolly and Thompson, 1989; Wood and Walther, 1986). Fluid–rock ratios of 10 000, deduced for mineralised fractures (Rumble, 1989), emphasises the much more efficient process of fracture- versus pervasive-flow.

Clearly not all fluids will leave an obvious geochemical signature of their infiltration, and when close to chemical equilibrium, large quantities of external fluid can pass through the system without leaving a detectable geochemical trace.

The above arguments have been developed for a lower-crust uniform in it's permeability(K_{ϕ})–porosity(Φ) relationships. By analogy with upper crustal aquifers and aquitards, it is to be expected that the various metamorphic lithologies will exhibit contrasting K_{ϕ}–Φ relationships in the metamorphic pile.

Fluid–rock ratios can be considered in terms of either thermal or chemical effects. Figure 5 shows examples of the variation of volumetric fluid–rock ratio with initial temperature of the crust before infiltration of a higher temperature fluid.

11. ACCUMULATION OF METAMORPHIC FLUID METAMORPHIC AQUIFERS AND AQUITARDS

Electrical conductivity studies generally (Shankland and Ander, 1983), and seismic reflection studies in particular cases (Matthews, 1986; Matthews and Cheadle, 1986; Hurich and Smithson, 1987; Lueschen and Forest, 1987; Hyndman, 1988), have led to the suggestion that fluid accumulation is a common phenomenon in the lower crust. High electrical conductivity and some low-velocity seismic layers (and "Bright Spots") could reflect the presence of metamorphic aquitards" (Thompson and Connolly, 1990). These geophysical observations require fluid-filled porosities only on the order of 0.1 and 0.4%, respectively (Φ of 10^{-3} and 4×10^{-3}; Matthews, 1986; Matthews and Cheadle, 1986; Shankland and Ander, 1983).

The existence and location of metamorphic "aquifers" and "aquitards" (e.g. Ferry, 1987; Guzzetta and Cinquegrana, 1987) is dependent mainly upon permeability-porosity contrasts within the crust. During regional meta-

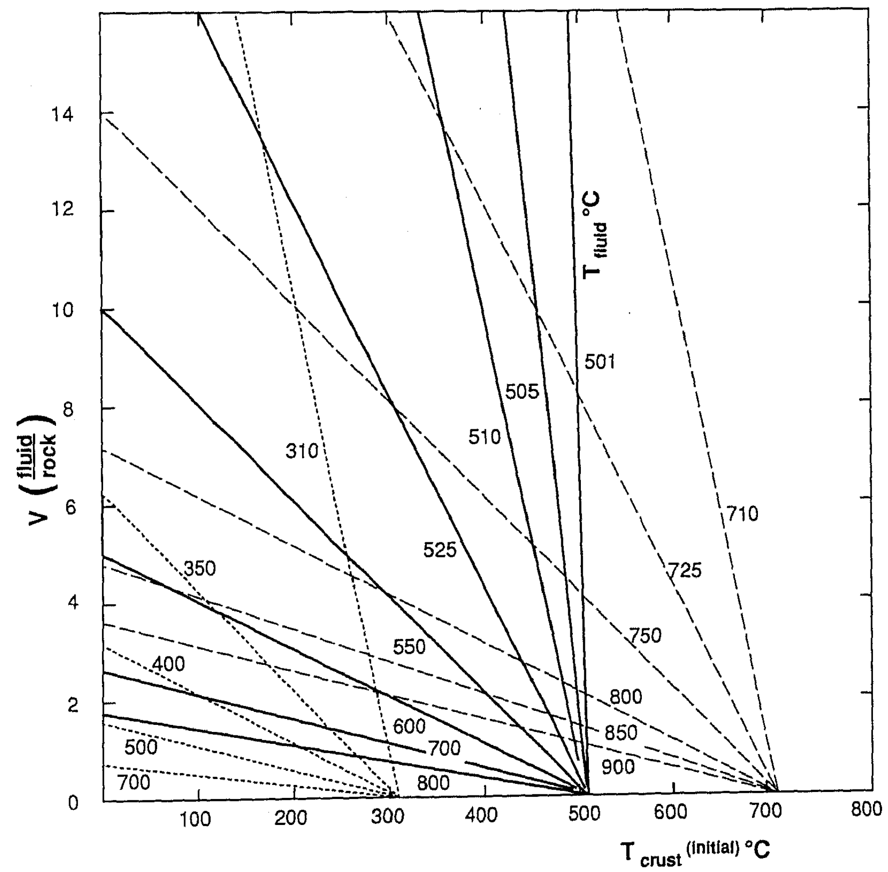

Fig. 5. Variation of volumetric fluid-rock ratios required to cause crustal heating, for different initial values for T_{crust}. Various values of initial T_{fluid} are shown for selected values of T_{final} of 700 °, 500 ° and 300 ° C. Heat conservation was assumed to be given by: $C_{p^{fluid}}/C_{p^{rock}} = \text{Volume}(^{fluid}/_{rock})$. $\Delta T^{fluid}/\Delta T^{rock}$, with equal values for volumetric heat-capacities.

morphism, relatively uniform and pervasive but transient porosities of 0.5 to 1% are predicted (e.g. Connolly and Thompson, 1989). Such uniform porosity distributions probably do not explain the above-mentioned layer-localised geophysical anomalies. Thompson and Connolly (1990) have shown that a 6 km "theoretical aquitard" [a layer with permeability–porosity (K_ϕ–Φ) relationships only one- or two-orders of magnitude lower than the rest of the crust], can maintain fluid-filled porosities up to 1.5% for time periods long in excess of those of isostatic readjustment (> 60 to 120 Ma) following crustal thickening.

Apparently, "theoretical aquitards" can store accumulated metamorphic fluid for millions of years after thickened crust has isostatically been returned to it's normal thickness. Alternatively, the geophysical manifestations may be readily referred to multiple seismic reflections (e.g. Hurich and Smithson, 1987) or interlayered lamellae of different lithologies (e.g. Meissner, 1986; Pratt et al., 1990). Seismic refraction studies in ancient crust would help to decipher whether "deep fluids" recognised by electromagnetic soundings (e.g. Bailey et al., 1989) are recent or ancient metamorphic aquitards.

Having examined aspects of metamorphic fluid-migration by porous media flow, it is necessary to attempt to constrain to what extent the pervasive flow needs to be focussed into fractures to lead to measureable, or recordable, heat- or mass-advection.

12. FRACTURE-FLOW

Fluid flow along fractures is an efficient way of heating rocks in shallow level geothermal systems, especially where shallow magmatic heat sources induce heating of meteoric waters. Although the long-term rheological behaviour of the lower continental crust is ductile, high fluid pressures the lower crust can also fracture—as witnessed by emplacement of mantle-derived magmas, and expelled fluids (e.g. Spera, 1987). It is useful to examine the possible thermal effects of fracture flow of metamorphic fluid in the deep crust (e.g. Wood and Walther, 1986; Brady, 1988; Connolly and Thompson, 1991), in terms of fracture width and spacing and for the quantities of fluid likely to be involved.

Fractures need to be widely spaces to produce thermal effects different from pervasive porous-media flow, (e.g. Peacock, 1989: p. 484). If we consider heat to be conducted laterally from the channel, characteristic diffusion distances (l, m) can be calculated for particular time scales (t, s) from $l^2 = \kappa t$.

For a typical thermal diffusivity (κ, 10^{-6} $m^2\,s^{-1}$), then $l = 5.6$ and 17.8 km for $t = 10^6$ a and 10^7 a, respectively. Thus for noticeable thermal perturbations to result from fluid-flow up fractures, channels need to be spaced more than 11 km and 35 km, apart for these characteristic time scales.

To produce a 10 K increase in temperature next to a fracture over a period of 10^7 a, Brady (1988, fig. 4) has concluded that an advected fluid heat flux rate of about 10^{-3} ($J\,m^{-2}\,s^{-1}$) is required. It can be seen from Fig. 3 that a fluid-flux rate greater than 10^{-8} $m^3\,m^{-2}\,s^{-1}$ is therefore necessary. As this flux rate is 10^3 to 10^4 greater than the estimated metamorphic fluid-flux then large amounts of fluid focussing is required, from at least 1 to 10 km^2 into a km long channel. The fluid-focussing ratio can be defined by the ratio of the horizontal cross sectional area of the source region to that of the channel (Connolly and Thompson, in prep.).

13. FOCUSSED VERSUS PERVASIVE FLUID FLOW

The physical mechanisms by which pervasively low-flux flowing fluids become focussed into high-flux fractures or channellways, are poorly understood, mainly because of lack of clear geological evidence. Likely mechanisms include hydrofracturing of brittle material, shear failure in ductile material, layers undergoing devolatilisation with significant potential volume changes; or the reactivation of pre-existing heterogeneiities in the crust. Such inherited heterogeneiities include lithological contacts, old faults or shear zones (Fyfe and Kerrich, 1985; Connolly and Thompson, 1991).

If build-up of fluid pore-pressure can induce hydrofracturing (even rapidly in the ductile-regime of the lower continental crust, e.g. Fischer and Paterson, 1989), then fluid flow will become focussed. Even then, a faster fluid escape rate means that the supplying reservoir (normally the mineral reactions in a particular reacting layer in a slowly evolving thermal regime) will need time to catch up to the required rate of fluid supply. The problem then is mainly one of discovering which geological phenomena require focussed – as opposed to pervasive fluid-flow, and then looking in specific field localities for local lithological, or structural, causes for the focussing (Brady, 1988). Notwithstanding these difficulties, we can still usefully explore the magnitudes of fluid focussing required to produce specific thermal anomalies (Connolly and Thompson, 1990, in prep.).

14. METAMORPHIC HOT-SPOTS AND OTHER REGIONS OF FOCUSSED FLUID-FLOW

Types of environments where high fluid fluxes have been proposed, include magmatic boiling (Burnham, 1967), mantle degassing (Spera, 1987), suduction-zone metamorphism (Peacock, 1987, 1990; Bebout and Barton, 1989), flow in accretionary wedges (Cloos, 1984; Moore et al., 1987; Reck, 1987; Vrolijk, 1987; Moore, 1989; and potential metamorphic hot-spots (Brady, 1988; Chamberlain and Rumble, 1988, 1989).

Even though crustal- or mantle-derived magmas can contain large percentages of dissolved volatiles, advective thermal effects will only be realised for very short time scales – much shorter than conductive-cooling timescales (a 10 km thick intrusion can cool in less than 1 Ma, depending on temperature contrast and magma temperature; Carslaw and Jaeger, 1959). Focussed flow of magmatic fluids could possibly leave recognisable thermal and geochemical signatures (see Brady, 1988; Chamberlain and Rumble, 1988, 1989).

Although subduction-zone metamorphism is frequently proposed to be a source of devastatingly large fluid production, more constrained considerations (e.g. Peacock, 1987, 1990), reveal that thermal effects are only likely if the fluid-flow becomes extremely focussed. Similar constrained viewing of processes in accretionary wedges (Reck, 1987), suggest that fluid-flow (or the thermal effects thereof) is not pervasive but must become focussed (Hoisch, 1987; Peacock, 1987).

Metamorphic "hot-spots" have been suggested by Chamberlain and Rumble (1988, 1989), as being responsible for measured shifts in oxygen-isotopes, and apparent temperature increases, in small locations in New Hampshire. Taken at face-value their deductions require enormous focussed influx of very high-temperature fluids into 10^2 km locations over a wide geographical area. Connolly and Thompson (1990) have provided, by way of 2- and 3-dimensional model calculations, examples of the optimum temperature-fluid flow conditions required to produce the thermal contrast proposed by Chamberlain and Rumble (1988, 1989). To produce an increase of temperature above ambient-500 °C to greater than 700 °C in this example, requires fluids at magmatic temperatures. Even then the fluxes of fluids needs to be high ($\sim 10^{-10}$ $m^3 m^{-2} s^{-1}$) and so requires focussing-factors of 100 to 1000. Such fluxes can seemingly only reflect magmatic boiling, and even then would require sequential multiple injection of magma bodies, intruded rapidly, to overcome the time constraints of conductive cooling (e.g. Hanson and Barton, 1989). Without pre-xisting fracture systems along which high-fluxes of hot magmatic fluid can be focussed, the fluids need to induce their own crack propagation (e.g. Secor and Pollard, 1975; Knapp and Knight, 1977; Segall, 1984; Nishiyama, 1989), otherwise the fluids will rapidly thermally equilibrate with lower crustal rocks.

There appear to be several geological instances where focussed flow of deep-crustal fluid has left geochemical signatures (e.g. Fyfe and Kerrich, 1985; Hoisch, 1987; Oliver and Wall, 1987; Gilotti, 1989; Philippot and Kienast, 1989; Jamtveit et al., 1990), and very fertile areas of research, by which possible mechanisms of fluid–rock interaction may be deciphered.

15. DEFORMATION AND FOCUSSED-FLUID FLOW

Intra- and sub-crustal detachment by sub-horizontal displacement appears to be facilitated by the way in which aqueous fluids can influence the deformation mechanisms of many crustal rocks (e.g. Etheridge et al., 1983; Brodie and Rutter, 1985; Bell and Cuff, 1989; Fischer and Paterson, 1989). It is mechanically easier to reactivate old heterogeneities than to initiate new ones in homogeneous rock. Such inherited heterogeneiities include old lithological contacts and layers of different permeability, strength and fertility (in the sense of being able to produce metamorphic fluid). The reactivation of structural heterogeneities is a complex function of the location of the most recent tectonothermal events; the local stratigraphy – which is also partly tectonically controlled, and the much earlier tectonic history that first introduced the heterogeneities into the crustal column.

Deep crustal shear-zones are likely candidates for zones of focussed fluid flow (e.g. Eisenlohr et al., 1989). Unfortunately, stable-isotope and other geochemical evidence is largely ambiguous as to whether migrating deep-crustal fluids used existing shear-zones to escape, or whether deep-crustal fluids caused the shear-zones to initiate and propagate (Fyfe and Kerrich, 1985; Frueh-Green, 1987).

16. FLUIDS AND CRUSTAL MELTING

The subsequent fate of upwards percolating metamorphic fluid depends upon whether crustal melting occurs, whether retrograde reactions remove fluid into hydrous minerals, and the porosity (Φ)–permeability (K_ϕ) characteristics of overlying crustal layers.

Because the solubility of H_2O is very large in granitic melts in the middle and lower crust, if crustal anatectic melts can migrate, then they can transport volatiles to shallower crustal levels. Upwards migration of crustal anatectic melts is limited because of the restricted (less than 25 vol.%) amounts of melt likely to be generated without a thermal input from the asthenosphere or mantle (Thompson and Connolly, 1990). Extensive crustal melting indicates either heat or fluid input from the mantle – geological reconstruction will indicate whether such regimes are related to continental collision, continental extension or subduction zones (Thompson, 1990).

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