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Metamorphic fluids and anomalous porosities in the lower crust

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ABSTRACT

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Because of the plastic behavior of rocks at high temperature and confining pressure, the porosity, ϕ, of lower crustal rocks is a variable dependent on the amount of free fluid present. On the basis of hydrologic theory it is predicted that the permeability, k, of such rocks can be expressed by a relationship of the form $k \approx \omega\phi^n$, where ω and n are material specific constants. The exponent n is likely to vary between 3 and 6, whereas ω, which is ultimately related to grain size and shape, probably varies over several orders of magnitude. Consequently, substantial contrasts between the ϕ–k relationships of lower crustal rocks are to be expected. The case in which thin crustal layers, "aquitards", are characterized by relatively low values of ω or n, with respect to the crust in general, is of particular interest. Such layers would acquire high porosities in the presence of miniscule steady state crustal fluid fluxes and are a possible explanation for anomalous seismic "brightspots" and high electrical conductivities in the lower crust. Modelling of collision belt metamorphism suggests that even long after the cessation of active metamorphism and isostatic reequilibration, relict metamorphic fluid fluxes could generate aquitard porosities, between 0.1 and 0.4%, adequate to explain such geophysical anomalies.

Introduction

Metamorphic hydrology and the possibility of deep crustal aquifers and aquitards has become particularly important in recent years for several reasons. Petrographic observations supplemented by stable isotope analysis have revealed that specific rock layers have acted as aquifers for metamorphic fluids (Beach, 1976; Fyfe et al., 1978; Rumble and Spear, 1983; Ferry, 1987; Rye and Bradbury, 1988). Most of these studies have concentrated on rocks metamorphosed at shallow crustal depths (< 15 km). Results from deeper crustal levels have proved more difficult to interpret because of uncertainty in the timing of the fluid–rock interation events (Valley, 1986; Frueh-Green, 1987, 1990). However, the existence of deep crustal fluids is suspected because magma generation by crustal anatexis apparently requires, in some cases, ingress of external fluid (Wickham and Taylor, 1987). Electrical conductivity studies generally, and seismic reflection studies in particular cases, have also led to the suggestion that fluid accumulation may be a common phenomenon in the lower crust.

Fluids percolate upwards through the crust by porous-media or fracture flow. In this paper we focus on the former mechanism as a means of accumulating fluids in the lower crust. Given the plastic rheologies of lower crustal rocks, and the likely amounts of free fluid in these rocks, lower crustal porosity probably consists of grain boundaries, microcracks, and disconnected pores filled by fluids at, or near, lithostatic pressure. Despite this disconnected porosity, porous-media fluid fluxes can probably still be maintained due to fluid buoyancy and the plastic behaviour of crustal rocks at high temperature and confining pressure.

Lower crustal fluids may be introduced from

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the mantle, or they may be generated by dehydration reactions during metamorphism. Because the former process is not easily constrained, we consider only internal generation of crustal fluids here. Such fluids are released from hydrous minerals when the pressure–temperature–time (P–T–t) path of a rock passes through the P–T conditions for a relevant devolatilization reaction. The subsequent fate of upwards percolating metamorphic fluid depends upon whether crustal melting occurs (the solubility of water being very large in granitic melts in the middle and lower crust), whether retrograde reactions remove fluid into hydrous minerals, and the porosity (ϕ)–permeability (k) characteristics of overlying crustal layers. We will concentrate on this latter aspect in this paper, and specifically attempt to identify geological circumstances which will lead to accumulation of metamorphic fluid in parts of the lower continental crust.

We consider first lower crustal porous media fluid flow and a model for the release of fluids during dehydration metamorphism subsequent to tectonic thickening. We then present the crustal porosity distributions obtained from the modelling.

A model for porous media flow in the lower crust

In contrast to upper crustal rocks, rocks in the lower crust are expected to be plastic on short time scales so that porosity can only be maintained if rock pores are filled with fluid at, or close to, lithostatic pressure. Lower crustal fluids, therefore, tend to follow a lithostatic pressure gradient. Under a lithostatic pressure gradient, fluids are inherently mechanically unstable and will always tend to migrate upwards (Walther and Orville, 1982; Walther and Wood, 1984; Valley, 1986). The boundary between the upper and lower crustal hydrologic regime varies depending on rock competency. Geologically, this appears to be a function of temperature and hence tectonic age. In general, the transition, which occurs over 2 or 3 km, takes place at depths of 6 to 10 km (Fyfe et al., 1978; Sibson, 1983; Wood and Walther, 1986). For present purposes it is assumed that lower crustal vertical fluid fluxes (q_f) can be estimated by Darcy's law:

$$q_f = -k/\mu_f (\mathrm{d}P_f/\mathrm{d}z) \tag{1}$$

where μ_f is the fluid viscosity, $(\mathrm{d}P_f/\mathrm{d}z)$ is the fluid pressure gradient, and k is the hydraulic permeability. For the conditions of interest, both viscosity and fluid pressure gradient are essentially constants ($\mu_f = 1.5 \cdot 10^{-3}$ g cm^{-1} s^{-1}, and $(\mathrm{d}P_f/\mathrm{d}z) = g(\rho_r - \rho_f) = 1800$ g cm^{-2}s^{-2}). Thus, any variability of the fluid fluxes from eqn. (1) is due to permeability. Lower crustal permeabilities are unknown, but it is expected on the basis of hydrologic theory that permeability is proportional to a power of porosity (Bear, 1972; Dullien, 1979; Ungerer et al., 1987):

$$k \approx \omega\phi^n \tag{2}$$

In eqn. (2), both ω and n are material-dependent constants that are approximately independent for low values of porosity ($< 1\%$). The exponent n, varies between 3 and 6, and increases with increasing content of nonequidimensional grains in a rock. The constant ω, may vary over orders of magnitude and is dependent grain size, shape distribution, and wetting properties.

In the models of Connolly and Thompson (1989) for fluid migration in a homogeneous crustal column, the hydrology of the crust was characterized by $\omega = 2.9 \cdot 10^{-10}$ cm^2 and $n = 3$ in eqn. (2). Consequently, their models generate a relatively uniform distribution of porosity with depth except in the immediate vicinity of active fluid producing reaction fronts. In this work we will illustrate the effect of crustal layers with contrasting ϕ–k relationships, i.e., aquitards. Ordinarily, aquitards and aquifers are defined to have a low and high permeabilities, respectively, but with variable porosity–permeability relationships this definition becomes meaningless. For by varying the porosity of a rock it is possible to obtain any permeability. Instead, we use the term aquitard here to describe a rock that must have a relatively high porosity to achieve a given permeability, i.e., an aquitard is characterized by a low value of ω or n in eqn. (2).

The effect of an aquitard, characterized by the values ω_a and n_a, embedded within a crust, characterized by ω_c and n_c, is easily understood because, through eqns. (1) and (2), fluid fluxes are proportional by ω to porosity raised to the nth power. Thus, in the presence of a steady state crustal fluid flux, the porosity of an aquitard will increase, by inflation of microcracks and pores,

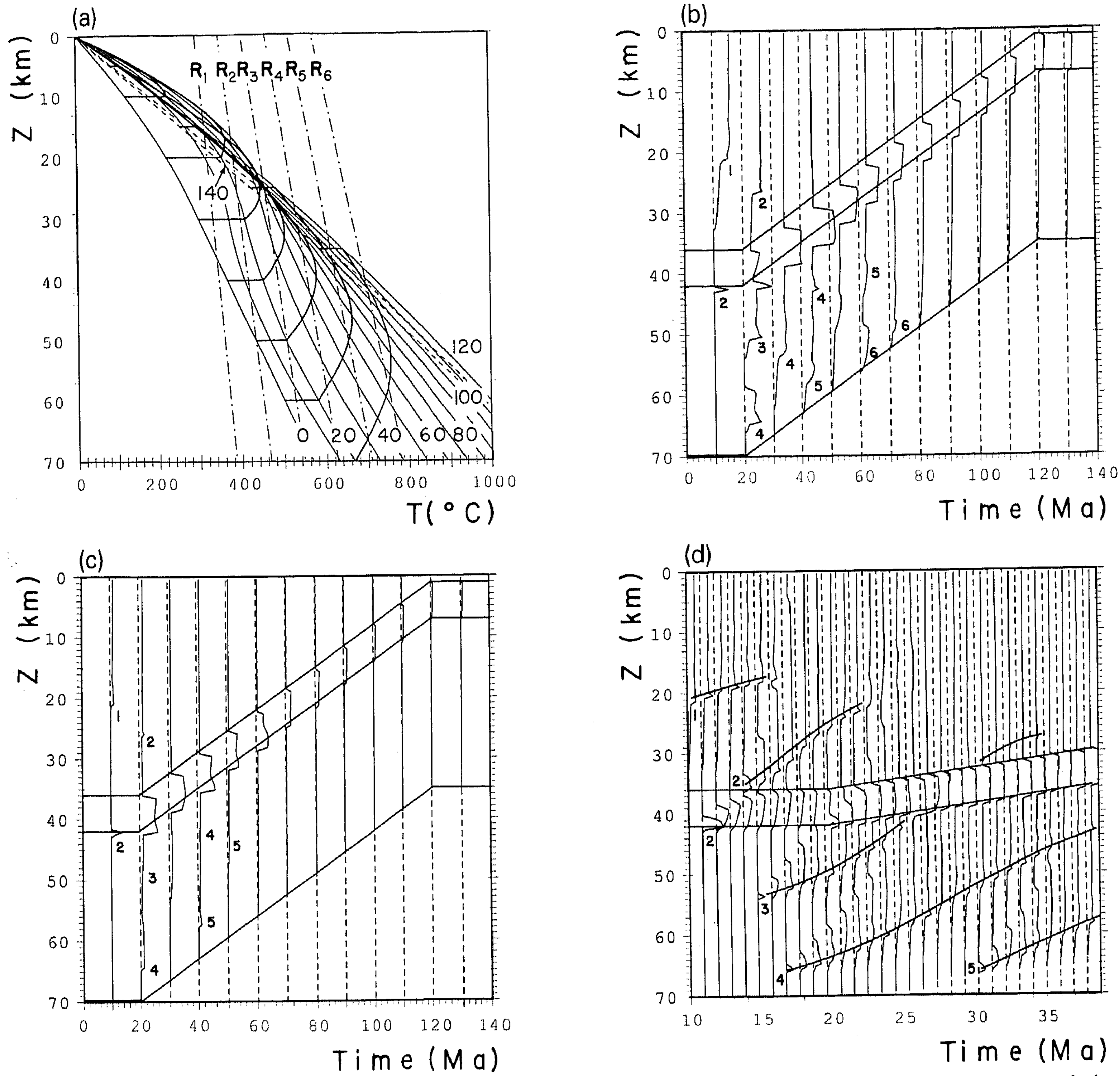

Fig. 1. Results of the homogeneous thickening model. (a) Depth, z (km)–Temperature, T (°C) diagram showing geotherm evolution (thin solid curves, labelled at 20 Ma intervals) and P–T–t paths of rocks at 10 km intervals (heavy solid curves). Model pelite dehydration reaction grid consists of reactions R_1–R_6 (uneven dashed lines). (b) Crustal porosity profiles at 10 Ma year intervals from 0 to 140 Ma for the homogeneous crustal thickening model with an aquitard at initial depths from 36 to 42 km. Porosity–permeability relations for the crust (c) and aquitard (a) are determined from eqn. (2) given the assignments: $\omega_c = 3 \cdot 10^{-10}$ cm^2 and $\omega_a = 3 \cdot 10^{-11}$ cm^2. The zero baseline for each profile is shifted by 0.01 porosity units for every 10 Ma of model time. Alternate profiles are labelled by time in Ma and positions of the base of the crust and the aquitard in each profile are indicated by a solid lines. Position of active reaction fluid producing fronts are indicated by the reaction numbers. Note that even after isostatic equilibration crustal fluid fluxes still maintain substantial porosities in the aquitard layer. (c) As for (b), but calculated with the parameter values of $\omega_c = 3 \cdot 10^{-8}$ cm^2, and $\omega_a = 3 \cdot 10^{-10}$ cm^2 for the crust and aquitard respectively. These values result in lower porosities for the models, but a larger contrast between the aquitard and the crust in general. (d) Porosity profiles at 1 Ma intervals from 9.86 to 38.86 Ma for the model shown in (b). The spacing of adjacent porosity profiles corresponds to an an absolute porosity of 0.01 or 1%. The time of the first profile, 9.86 Ma, is the time at which fluids first penetrate crustal rocks immediately below the aquitard. Reaction fronts and their effects on the aquitard porosity are easily identified.

until its permeability is such that it can maintain the fluid flux. From eqn. (2), the porosity of the aquitard, ϕ_a, is related to the porosity in the adjacent crustal rocks, ϕ_c, by the equation:

$$\phi_a = (\omega_c/\omega_a)^{1/n_a}\phi_c^{n_c/n_a} \tag{3}$$

For simplicity, we consider here only $n_a = n_c = 3$, in which case equation (3) becomes:

$$\phi_a = (\omega_c/\omega_a)^{1/2}\phi_c \tag{4}$$

From eqn. (4), an order of magnitude difference in ω between an aquitard and subjacent crustal rocks will lead to an approximate twofold ($10^{1/3} = 2.15$) increase in the porosity within the aquitard. Likewise, a two order of magnitude difference increases the aquitard porosity fivefold ($100^{1/3} = 4.64$). Inasmuch as substantial ϕ–k contrasts can be expected between crustal rocks, eqn. (4) provides a basis for predicting that aquitards could maintain substantial porosities even with negligible crustal fluid fluxes. It should be noted that ratios of n_c/n_a greater than one, or higher values for n_a, would diminish the effect of aquitards.

A model for fluid production in collision belts

The foregoing discussion suggests a mechanism that could create anomalously high crustal porosities. To demonstrate the magnitude of the porosities and the time scale of this mechanism we present here two models developed by Connolly and Thompson (1989) to simulate the generation of metamorphic fluid in response to crustal thickening. In these models, a 35 km thick hydrous crust is doubled in thickness to 70 km by homogeneous thickening or overthrusting. The initial prethickening geotherm in the models, as dictated by the thermal conductivity, basal heat flux, and distribution of radiogenic material, is a typical stable continental geotherm of 15 K/km (England and Thompson, 1984). Thickening is assumed to occur instantaneously, followed by a static period of 20 Ma before the onset of erosion (to simulate the time period required for thermal relaxation to effect the density of the thickened crust). Erosion then uniformly restores the crust to its original thickness, and isostatic equilibrium, in a period of 100 Ma. The models are one-dimensional, and this, coupled with the assumption of instantaneous thickening, tends to exaggerate the amount of metamorphic fluid generated. Nonetheless, the intensity and timing of metamorphism generated in the models is not unreasonable and serves as a useful basis for discussion.

The dehydration of the crust is based on an equilibrium model for the dehydration of pelitic lithologies. That the entire continental crust is composed pf pelitic material is unlikely, but it is used as a model because it maximizes fluid generation. More plausible lithologies (e.g., amphibolites, gneisses, and marbles) produce less fluid. To further maximize the amount of fluid present in the models, we do not consider the effects of retrograde hydration, which tends to consume fluid in the upper portions of the crust (Connolly and Thompson, 1989).

The initial hydrate content of the crust in the models, which averages 2.94 wt.%, is determined by intersections of the prethickened crustal geotherm with the model dehydration reaction grid. Subsequent to thickening, the evolution of the geotherm and the consequent P–T–t paths, in combination with heat flow constraints, determine the time, depth, and rate at which water is released by dehydration. The evolution of the geotherms and the P–T–t paths for the two thickening models, homogeneous thickening and thrust thickening, are shown in Figs. 1a and 2a, respectively. Connolly and Thompson (1989) found that vertical fluid advection had negligible effect on the thermal development of the models. Thus, in the following discussion, we consider hydrologic variations without specifically assessing their effect on the thermal development of the metamorphic column.

Initial porosity, i.e., free water, in the crustal models is taken to be zero and it is assumed that the fluid released by dehydration is able to create porosity (in the general sense used earlier) by inflation of the rock.

Porosity evolution of continental crust with an aquitard

To illustrate the effects of contrasting ϕ–k relationships, we consider here the effect of an 6 km

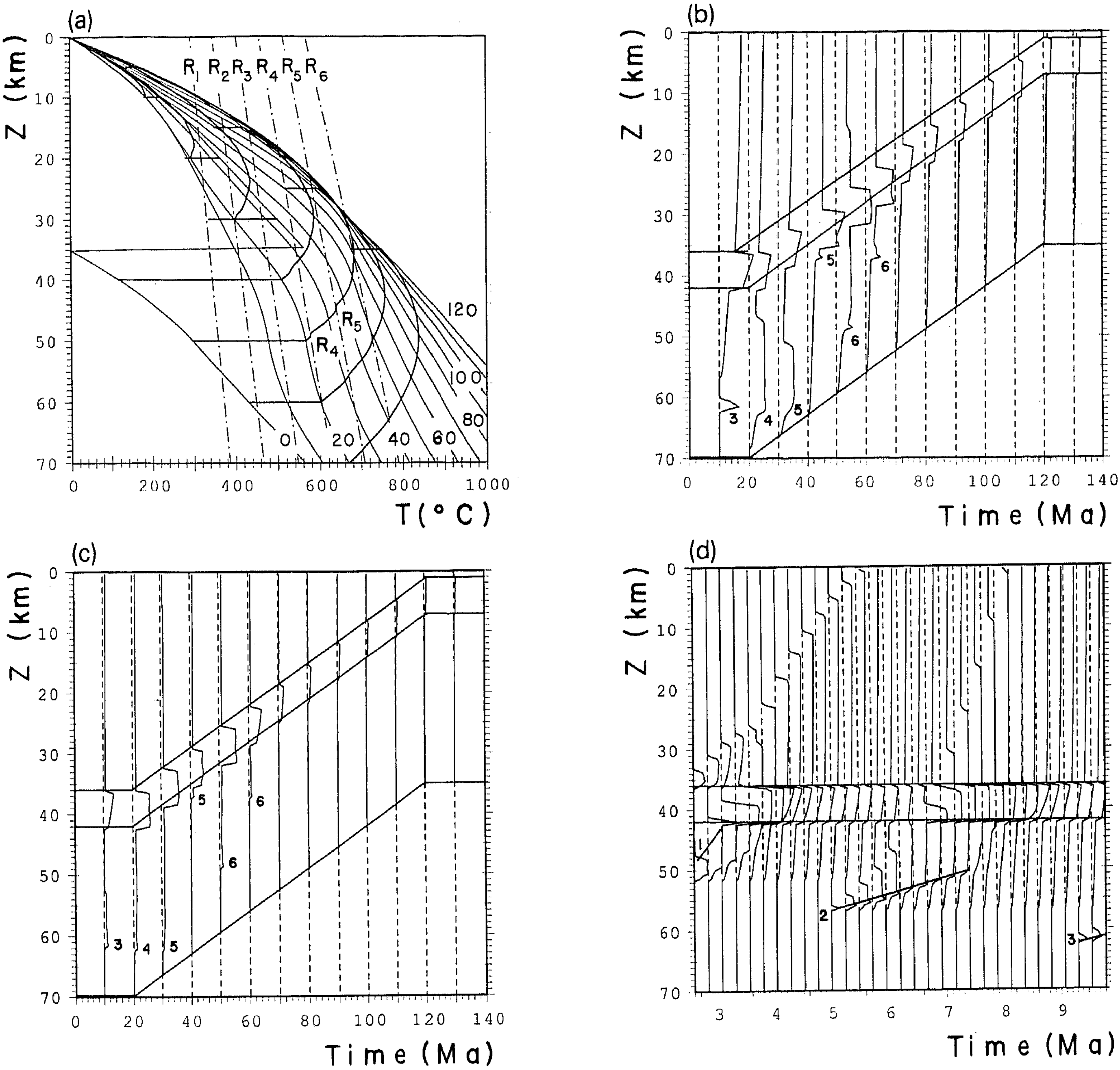

Fig. 2. Results for the thrust thickening model. In comparison to the homogeneous thickening model (Fig. 1), the thrust model generates a larger thermal perturbation, consequently higher porosities are obtained because of more rapid and extensive dehydration. Refer to the respective captions for Fig. 1 for descriptions of (a–c). (d) Porosity profiles at 0.25 Ma intervals from 2.59 to 9.84 Ma for the model in (b). The time of the first profile, 2.59 Ma, is the time at which fluids first penetrate crustal rocks immediately below the aquitard. Compared to Fig. 1d, the maximum porosities are larger because of more rapid dehydration.

thick aquitard at the depth interval from 36 to 42 km in the otherwise homogeneous thickened crust. This position was chosen because rocks at these depths have the highest fluid–rock ratios and are not removed by erosion during isostatic equilibration. To simplify our results it is assumed that the aquitard itself does not undergo dehydration, therefore, fluid is introduced into the aquitard only from dehydration reactions at greater depths.

Connolly and Thompson (1989) have suggested that their value for ω, $3 \cdot 10^{-10}$ cm^2, in eqn. (2) leads to low estimates for crustal permeability (e.g., for $\pi = 10^{-3} = 0.1\%$, $k = 10^{-18}$ cm^2; and for $\pi = 10^{-2} = 1.0\%$, $k = 10^{-15}$ cm ^{2}) To show order of magnitude variations here we present two cases of ω values for an aquitard within the crust: (i) $\omega_c = 3 \cdot 10^{-10}$, $\omega_a = 3 \cdot 10^{-11}$ cm^2 and (ii) $\omega_c = 3 \cdot 10^{-8}$, $\omega_a = 3 \cdot 10^{-10}$ cm^2. For the thermal conditions realized in the lower crust (depth > 12 km) for the two models (Figs. 1a and 2a), the density of water can be reasonably assumed to be a constant 0.95 g/cm^3. Thus, the amount of free water present in the crustal rocks can be measured in terms of porosity because of the model assumption that pores are filled by water.

The distribution of free water, as measured by porosity, with depth in the crust for the different models is shown in Figs. 1b, 1c, 2b, and 2c at 10 Ma intervals. In these figures, the zero porosity baseline for each depth vs. porosity profile is drawn at the time ordinate for the profile. The value of the porosity at any depth is scaled such that the spacing between successive profiles corresponds to a porosity of 0.01 or 1%.

Because fluid advection has negligible effects on the thermal development of the models, the variations in the porosity profiles reflect essentially only variation in the parameter ω. As anticipated with eqn. (3), in the crustal porosity profiles of Figs. 1a nd 2, porosities in the aquitard are greater than in the adjacent crust. The irregularity of the early porosity profiles in Figs. 1b, 1d, 2b, and 2d reflects the low ω values for these models and the fact that multiple reaction fronts are simultaneously active at different depths in the crustal column. The position of these fronts is indicated by peaks in the profiles and are particularly easy to identify in Figs. 1d and 2d. During this stage, the crustal porosity is primarily dictated by the rate of fluid production, i.e., porosity increases until the permeability permits fluid to drain as rapidly as it is produced. For the models shown in Fig. 1b and 2b, this porosity is about 0.5%, except within the aquitard where it reaches values around 1%. The high porosities that develop in the first 70 Ma could explain the high electrical conductivities reported for layers within the crust (e.g., Hyndman, 1988). At model times beyond 70 Ma, active metamorphism ceases and porosities gradually decrease.

The models shown in Figs. 1c and 2c, differ from those shown in Figs. 1b and 2b in that they are characterized by higher values of ω, and by a larger contrast between the aquitard and crust ($\omega_c/\omega_a = 100$). Because of high ω values, lower model porosities are necessary to accommodate the reaction generated fluid fluxes. Lower porosities, in turn, mean that there is less of a lag between the onset of metamorphism and the achievement of steady state profiles, hence the more uniform porosity distributions in Figs. 1c and 2c. Although, the porosities are lower for these models, the contrast between the crust and aquitard porosities is greater. Contrasts of this kind, due to lithologic layering within the crust, are probably capable of creating certain deep crustal geophysical anomalies.

An interesting feature of the models is that relict metamorphic fluid fluxes and substantial crustal porosities are maintained long after the end of dehydration metamorphism and fluid production ends at about 70 Ma (Figs. 1b, 1c, 2b, 2c). Aquitard porosities in excess of 0.1% persist even at a model times of 130 Ma, 10 Ma after tectonic equilibrium. This suggests that aquitards could produce geophysical anomalies in terrains which otherwise appear to be normal, in terms of seismic or thermal activity. A counterintuitive result of the variable ϕ–k models is that the rocks with the highest porosity are aquitards rather than aquifers. This result applies only in the steady state case which applies on the long time scales illustrated in Figs. 1 and 2. On short time-scales, there is a period during which metamorphic fluids accumulate below the aquitard layers, but negligible porosities are obtained within the aquitards (Figs. 1d, 2d).

Variations and limitations of the modelling

The porosity of the aquitards in the models presented here is determined by the steady state fluid flux in the crust and is not significantly affected by the thickness of the aquitard. Because the metamorphic fluid fluxes decrease with depth, the porosity of an aquitard will decrease with its depth. The magnitude of such variations, as well as variations in the ϕ–k relationship of either the aquitard or crust, can be determined through eqns. (1) and (3). This modelling does not apply at depths less than 12 km, because at such depths fluid pressure becomes independent of lithostatic pressure with the consequence that both substantial lateral and downward fluid fluxes are possible.

Under these conditions it is more likely that anomalously high concentrations of free water develop within aquifers or below aquitards. Moreover, at intermediate crustal depths, retrograde metamorphic processes, which were not considered here, may consume the metamorphic fluid flux leaving the rocks completely dry (Connolly and Thompson, 1989).

If eqn. (2) is examined in more detail, some suggestions can be made as to the likely characteristic of metamorphic aquitards. The term ω in the Carman–Kozeny formulation (cf., Dullien, 1979) contains expressions for grain size,and tortuosity (grain shape). In general, rocks with smaller grain size and grains with high surface area (platy minerals) would exhibit a lower permeability for a given porosity. Thus, in terms of oil reservoir materials, compacted shales and tight limestones are less permeable than sandstones, marls or fractured rocks (Ungerer et al., 1987). In terms of lower crustal materials, schists and other rocks with micas or large abundance of other platy minerals, are likely to be less permeable than equigranular well-annealed coarse-grained rocks such as granites or granulites.

The early history of fluid accumulation near an aquitard needs to be investigated in more detail, by different hydrological modelling. The present model assumes that fluid is always at lithostatic pressure appropriate to any depth. In rocks where the rheological response of the matrix operates on different time scales than porosity inflation (Rice, 1979; Rice and Cleary, 1976), failure of an aquitard by fracture is likely to occur. Fractures are propagated in part, by small gradients in the fluid pressure (Weertman, 1971; Spence and Turcotte, 1985; Spera, 1987).

Relevance of the modelling to geophysical observations

High electrical conductivity (low electrical resistivity) in certain crustal layers and some seismic low-velocity layers (and bright spots) have been ascribed by some workers, to the presence of free fluid. Shankland and Ander (1983) suggest that observed normal lower crustal electrical resistivity of 100–1000 Ω m could be due to 0.1 to 0.01 vol.% of free H_2O ($\phi = 10^{-3}$–10^{-4}). Haak and Hutton (1986) have suggested ways in which this electrical resistivity can be accommodated in thinner layers, but with higher fluid porosity, during tectonic development of continental crust.

A direct connection between high electrical conductivity and lower seismic velocity assigned to particular crustal layers has been proposed as being caused by accumulated free H_2O (Hall, 1985, 1986, 1987; Matthews, 1986; Matthews and Cheadle, 1986; Klemperer, 1987; Pavlenkova, 1987; Hyndman, 1988). Matthews (1986, p. 19) notes that for a reduction in seismic velocity from 6.5 to 6.4 km/s assuming this is achieved by a pore fluid with sound speeds of 1.5 km/s, requires a fluid-filled porosity of about 0.4%($\phi = 0.004$).

Seismic bright spots at 15 km depth beneath southeast Georgia have been ascribed by Brown et al. (1987) to fluids accumulated along an Appalachian suture; and that at 9.5 km beneath the Schwarzwald to a fluid trap in a Variscan thickening sequence reactivated by Mesozoic extension by Lueschen et al. (1987).

Our present models of porosity evolution have direct bearing on these observations. The required porosities could be explained by the presence of low ω aquitards which could remain inflated by the almost negligible fluid fluxes that persist long after (> 100 Ma) the cessation of active metamorphism. It should also be noted that in the opinion of some workers (e.g., Hurich and Smithson, 1987) many anomalous seismic reflection features (such as bright spots or lamellae) could originate by interference of reflections in complexly deformed and lithologically diverse terrains to be expected in the lower crust, and should not automatically be taken to indicate that a higher fluid-filled porosity is present.

Seismic bright spots or lower-velocity layers occurring around 10 km (KTB-Research Grove Black Forest, 1987; Pavlenkova, 1987) might require a special explanation not provided by our modelling. Depending upon temperature (and hence tectonic age) and the rheological–hydrological behaviour of actual lithologies, this depth range will correspond to the transition zone of rheological behaviour in continental crust. In simplified terms, the transition from average ductile to average brittle behaviour will correspond to a change in the average fluid pressure gradient, i.e., from lithostatic to hydrostatic. How this change in hydrological behaviour of upward migrating fluids will influence the accumulation characteristics, e.g., from the formation of water sills (Fyfe et al., 1978, p. 294), remains to be further investigated in terrains of different tectonic age and hence temperature distribution.

Conclusions

Crustal layers with hydrologic characteristics which contrast with the rest of the lower crust can maintain anomalously high porosities for long periods of time in the presence of small fluid fluxes. Modelling of the fluid fluxes generated by the metamorphism that occurs in response to collision belt metamorphism suggest that adequate fluxes can be maintained more than 70 Ma after the end of metamorphic activity. The resultant porosities, which are in excess of 0.1%, appear to be sufficient to account for seismic bright spots, some lower crustal relective lamellae, and high-electrical-conductivity layers. The model porosities appear to be inadequate to explain anomalous magnetotelluric resistivities unless the porosity is connected, and the pore fluids highly saline. During active metamorphism, the resulting high crustal porosities could generate rheological instabilities and hydrofracturing. This would considerably modify the long-term porosities estimated here. The process of pore dilatation, in response to fluid generation, needs to be examined in detail to determine if it is consistent with current rheological models.

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