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Time-Dependent Hydraulics of the Earth's Crust

AMOS NUR
Stanford University

JOSEPH WALDER
University of Washington

ABSTRACT

The deceivingly simple question posed here is how deep free water extends in the Earth's crust. This question is found to be inseparable from crustal porosity, permeability, and pore pressure and most significantly their dependence on time at depth. If we assume that porosity and permeability are time independent, it follows from *hydrological evidence* that ambient pore pressure must be hydrostatic. However, *geological evidence* suggest that rocks in situ rapidly seal hydraulically. To reconcile this conflict, the hypothesis is explored that crustal porosity, permeability, and hence pore pressure are in general time dependent due to the gradual closure of crustal pore space via healing, sealing, and inelastic deformation. It is found that when the hydraulic conductivity of the system is large, so that the ratio of porosity reduction rate ϕ to permeability k, ϕ/k, is small, the initially porous water-saturated crustal rock mass will gradually lose its porosity and fluid, until it becomes essentially dry. Pore pressure, P_p, throughout this process will remain around hydrostatic. If on the other hand, system permeability is small and ϕ/k is large, the pore fluid cannot escape fast enough and pore pressure will build up. It is envisioned that when the pore pressure reaches the level of the least compressive stress in the crust, natural hydraulic fracturing takes place, leading to some fluid release, pore pressure drop, and resealing of the system. With time P_p builds up again, leading to another cycle of hydrofracturing.

Analysis shows that tens to hundreds of such P_p cycles are possible and that the duration of one cycle may be 10^3 to 10^5 yr; the duration of the entire dewatering process can be 10^6 to 10^7 yr. If subducted lithosphere supplies additional waters, the process may last as long as 10^8 yr.

The conclusion that the crust is most likely undergoing repeated (1) cyclic episodes of pore pressure buildup to lithostatic and (2) water expulsion events associated with natural hydraulic fracturing may explain such observations as the episodic nature of some types of ore deposits and veins, the recurrence time between large earthquakes, the emplacement of foreland thrust systems, and the seismic and mechanical nature of crustal detachment zones.

INTRODUCTION

One of the simplest yet profound questions that can be asked about the state of the Earth's crust is the depth to which free water extends. The top of the free water zone in the crust, in the form of groundwater table, is present almost everywhere. The depth to the water table typically varies between 0 and 1000 m or so and is the subject of extensive hydrological exploration, especially in arid and semiarid regions. A much more difficult question is the depth of the bottom of the free water or groundwater in the crust (see Table 7.1): Is groundwater limited to the top 1 to 2 km, with the crust below being dry, or does free water extend to much greater depth, for example, the depth to which the crust is brittle (~10 to 15 km), or even deeper? And if free water is present at the greater depth, how much is present or, equivalently, what is the porosity in the crust? This question in turn is found to be inseparable from what the water's ambient, or steady-state, pore pressure P_p must be (Figure 7.1). Furthermore, as discussed in this chapter, the pore pressure question is itself inseparably linked to the question of permeability at depth.

In this chapter we consider the following three interlinked questions: What is the depth in the crust of free water and its pressure? What are typical crustal porosity and permeability? Are or can these parameters be time invariant? These simple questions are actually very important because their answers are keys to understanding a surprisingly wide range of geological and geophysical crustal phenomena. For example, the mechanisms by which crustal rocks deform tectonically are strongly influenced by the presence or absence of water as well as by the level of pore pressure (e.g., Carter, 1976; Brace and Kohlstedt, 1980). Circulation of crustal water has important effects on heat flow (e.g., Sleep and Wolery, 1978; Lachenbruch and Sass, 1980; Smith and Chapman, 1983), on the distribution of oxygen and hydrogen isotopes (e.g., Taylor, 1977; O'Neil and Hanks, 1980), and on the formation of

TABLE 7.1 Possible Indicators for the Presence of Free Water in the Earth's Crust and Estimated Depth Values

Indicator	Depth range	References
Water table	0 to 2 km	—
Deep wells	to 12 km	—
Reservoir induced seismicity	to 12 km	Bell and Nur (1978)
Crustal low velocity zones	7 to 12 km	Berry and Mair (1977)
		Feng and McEvilly (1983)
		Jones and Nur (1984)
Crustal electrical conductivity zones	10 to 20 km	Nekut *et al.* (1977)
		Shankland and Ander (1983)
Oxygen isotopes	to 20 km	Taylor (1977)
Metamorphism	>20 km	Fyfe *et al.* (1978)
		Etheridge *et al.* (1984)
Crack healing and sealing	?	Richter and Simmons (1977a,b)
		Sprunt and Nur (1979)
		Ramsay (1980)
		Smith and Evans (1984)
Formation of hydrothermal ore deposits	>5 km	Norton and Knight (1977)
		Cathles and Smith (1983)
Crustal seismic attenuation zones	7 to 15 km	Hermann and Mitchell (1975)
		Winkler and Nur (1982)
Low stress on faults	0 to 10 km?	Raleigh and Evernden (1982)

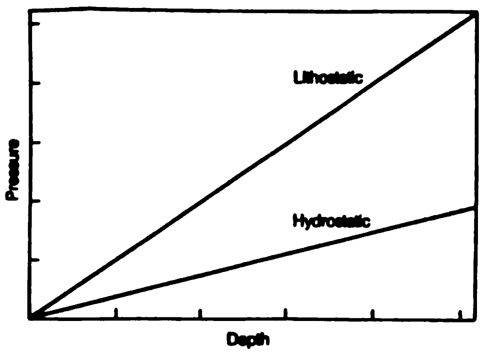

FIGURE 7.1 Rock and water pressure in porous rock in the Earth's crust. The gradients of the two lines are proportional to rock density and water density, respectively. When pore pressure is equal to hydrostatic, the difference between the two pressures gives rise to the tendency of the pores to inelastically close with time.

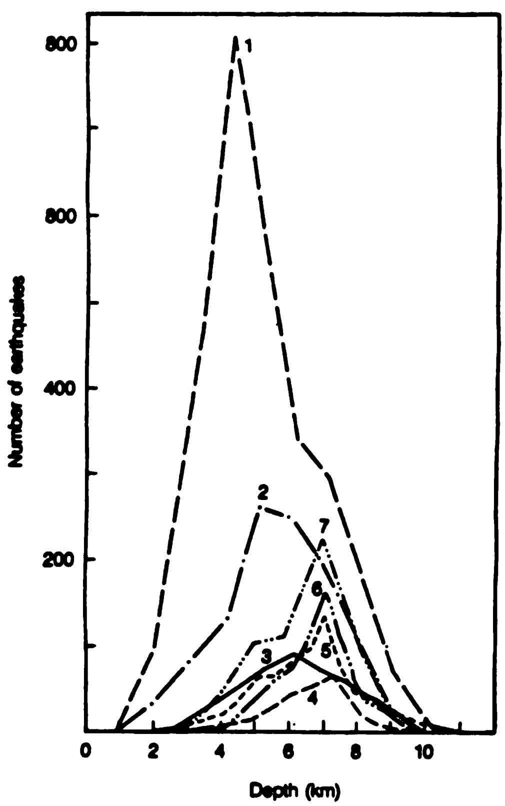

FIGURE 7.2 Example of the depth and time migration of earthquakes induced by reservoir impounding (from Chung-Kang *et al.*, 1974). As shown by Bell and Nur (1978), this induced seismicity requires the presence of free water at the depth of these induced earthquakes prior to the formation of the reservoir.

hydrothermal ore deposits (e.g., Norton and Knight, 1977). The role of pore pressure is also central in understanding the processes of the earthquake failure process (Byerlee, 1967), earthquake prediction (Nur, 1972), induced seismicity (Bell and Nur, 1978) (Figure 7.2), the mechanical processes in and below the accretionary wedge in subduction zones (Zhao *et al.*, 1986), the rate and depth of magmatic melting and volcanism associated with subduction zones (McGeary *et al.*, 1985), the nature of deep crustal seismic reflectors (Jones and Nur, 1984), and the state of stress in the crust (Zoback *et al.*, 1987).

Answers to the three questions posed must at present be based largely on indirect evidence. For example, in situ measurements of crustal hydrologic properties typically reach to depths of only 2 to 3 km (Brace, 1980). However, the presence of free water to much greater depths in the

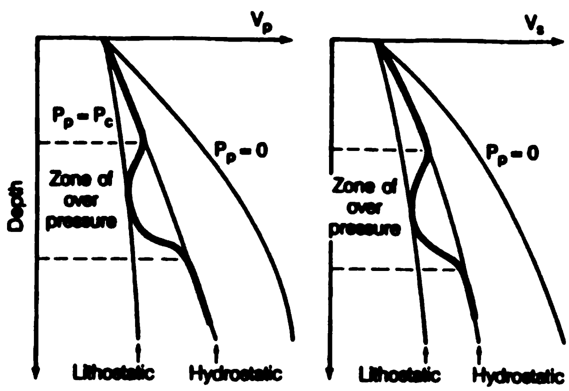

FIGURE 7.3 Schematic illustration of the development of a low compressional and shear wave velocity zone due to anomalously high pore pressure, based on extensive laboratory measurements (Nur and Simmons, 1969) and some field observations.

crust is suggested by isotopic studies of batholithic rocks (e.g., Taylor, 1977; Norton and Taylor, 1979), studies that indicate that meteoric water may circulate to depths of up to 20 km. Deep crustal electromagnetic soundings have revealed zones of relatively low electrical resistivity, which suggests the presence of a continuous water phase (e.g., Nekut *et al.*, 1977; Thompson *et al.*, 1983), as inferred from laboratory studies of the electrical properties of rocks (Olhoeft, 1981; Shankland and Ander, 1983).

Seismology has also contributed to ideas about the hydrologic character of the crust. For example, Berry and Mair (1977) argue that crustal low-velocity zones could be due to P_f locally in excess of hydrostatic. This argument is based on experimental results such as those of Nur and Simmons (1969) (Figure 7.3), who showed that even in very low porosity saturated rocks, compressional velocity

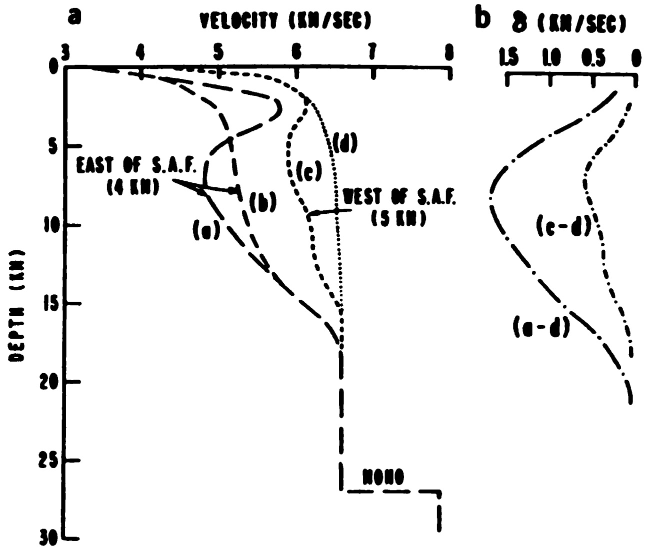

FIGURE 7.4 Compressional low-velocity zones near the San Andreas Fault in central California. The figure is taken from Raleigh and Evernden (1982), who based it on extensive crustal reflection measurements by Feng and McEvilly (1983). Raleigh and Evernden (1982) suggest that the (a) low-velocity zone and (b) velocity anomaly δ are due to high pore pressure in the depth range of 5 to 10 km or so.

drops markedly as P_f approaches the confining pressure P_c. Feng and McEvilly (1983) discovered a prominent low-velocity zone near the near San Andreas Fault in California, which has been interpreted to result from high pore pressure (Raleigh and Evernden, 1982; Figure 7.4). Similarly, crustal low Q zones, such as observed by Hermann and Mitchell (1973; Figure 7.5), are also consistent with the presence of zones with pore pressure close to lithostatic (Winkler and Nur, 1982). Using a similar line of reasoning, Jones and Nur (1984) suggested that reflections from deep crustal fault zones may be associated with elevated P_f within or below these zones. The hypothesis of elevated P_f has also been suggested by Raleigh and Evernden (1982) to explain the low deviatoric stresses thought to exist along plate boundaries such as the San Andreas Fault.

Other inferences about crustal hydrology are derived from geological evidence. For example, Fyfe *et al.* (1978) and Etheridge *et al.* (1984) have reviewed geological indicators for free water, with P_f often exceeding hydrostatic, being widespread during low- to medium-grade regional metamorphism. Two principal lines of evidence follow. (1) The ubiquity of mineralized fractures whose microstructure and orientation indicate that they formed in extension (Ramsay, 1980). On the basis of commonly accepted criteria for brittle failure, this requires that P_f exceed the minimum principal confining stress at the time of fracture formation. (2) Experimentally determined phase equilibria (with P_f equal to confining pressure P_c) are consistent with natural distributions of metamorphic mineral assemblages.

Together the above arguments suggests that free water must be fairly common at upper and mid-crustal levels. Furthermore, it appears that elevated P_f directly implies that the permeability of crustal rocks must be very low. Bredehoeft and Hanshaw (1968) and Hanshaw and Bredehoeft (1968), for example, studied simple models of crustal P_f development and concluded that, in general, maintenance of elevated P_f for geologically significant periods of time requires the presence of some crustal horizons with very low permeability, down to 10^{-21} m^2 (1 ndarcy) and lower.

The evidence and arguments favoring the presence of water at great crustal depth at high P_f and the consequent implication that crustal permeability must be low are in remarkable disagreement with the conclusion of Brace (1980), who after reviewing direct and indirect estimates of crustal permeability (Figure 7.6) argued that zones with

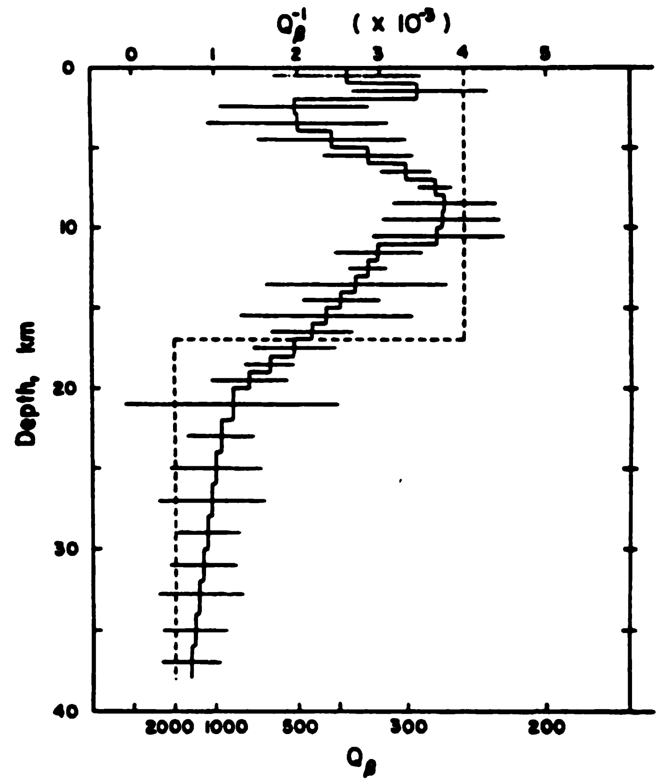

FIGURE 7.5 Estimated shear wave specific attenuation versus depth obtained from surface wave measurements in the stable North American continent (from Hermann and Mitchell, 1975).

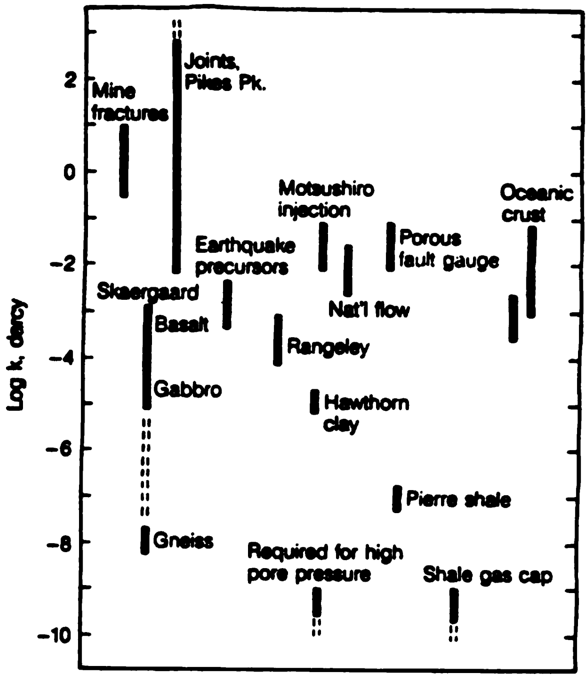

FIGURE 7.6 In situ permeability inferred from various large-scale phenomena. Numbers in parentheses refer to accompanying notes, which explain the calculation of k (from Brace, 1980).

permeability of about 10^{-15} m^2 (1 mdarcy) or higher must exist down to at least 10 km depth. Brace therefore concluded that P_f is very unlikely to exceed hydrostatic pressure in regions where crystalline rocks extend to the surface and that P_f above hydrostatic in crystalline rocks could be maintained only if a cover of very low permeability (e.g., argillaceous) rocks were present. Also, Jones (1983) examined the hypothesized relationship between seismically reflective zones in the crust and elevated P_f by combining simple models of P_f development with synthetic seismograms. Jones concluded that, although elevated pore pressure can significantly affect the existence and amplitudes of reflected waves, such effects persist for geologically significant periods of time "only for a permeability lower than that generally observed in laboratory measurements on crustal rocks."

Clearly, a model is needed that can reconcile the two conflicting lines of evidence regarding crustal hydrology: evidence for ubiquitous high pore pressure and hence low permeability on the one hand, and evidence for relatively fast flow and fast pore pressure dissipation on the other. We will outline such a model by considering two general cases of crustal hydraulics: (1) time-invariant crustal porosity (and hence permeability) and (2) time-dependent porosity, permeability, and pore pressure.

TIME DEPENDENCE OF POROSITY AND PERMEABILITY

The simplest model one can envision for the hydraulics of the crust is based on the assumption that porosity and, consequently, permeability remain unchanged with time or are time invariant. In that case hydraulic permeability or diffusivity inferred from in situ phenomena (Brace, 1980) directly provides an estimate of the typical steady-state or ambient crustal permeability. In fact, Brace's compilation of numerous such estimates (Figure 7.5), made for a wide variety of rock types and geological settings, results in typical values of crustal permeabilities ranging from as much as 10 to 100 darcy for shallow fractured rock masses to 10^{-1} to 10^{-5} darcy for most rocks down to 10^{-6} to 10^{-7} darcy for shales and some gneisses.

As pointed out by Brace (1980), pore pressure in a crust with these rock permeabilities will generally have to be close to hydrostatic, with the water at depth thus sufficiently connected to the free surface of the crust, so that the pressure at any depth is simply the weight per unit area of a column of water reaching the Earth's surface. Episodes or regions of overpressure will thus be special cases, transient in time and localized in space.

The apparent contradiction between the indirect evidence for elevated crustal P_f, on the one hand, and the inferred relatively high permeability throughout the crust, on the other hand, cannot be reconciled with this model. Suppose instead that porosity, permeability, and consequently pore pressure can vary significantly with time rather than remain static. Such variations have been inferred by D. Norton (University of Arizona) and co-workers in communication with fluid flow and mineralization associated with magmatic intrusion in the crust. Norton (Chapter 2, this volume) suggests that the pore fluid pressure induced by such intrusions sufficiently exceeds lithostatic pressure to cause natural hydraulic fracturing. Ramsay (1980) described indirect evidence for repeated episodic P_f pulses in the form of composite crack filing veins, presumably episodically precipitated from solution. In this chapter we explore the feasibility of this kind of episodic P_p buildup not as limited to magmatic intrusions but as a general behavior of the crust. Such general time dependence might allow for periods of fast fluid flow, for example, during episodes of P_f equal to lithostatic, bracketing periods of lower pore pressure, and no or little flow. Brace's geological estimates may thus represent only the periods or episodes of fast flow and not the crust in general. The immediate question that arises is: How must porosity deep in the crust decrease with time, and can this decrease be rapid enough (e.g., due to inelastic processes in porous rocks)? If so, what are the conditions under which permeability changes through time to maintain P_f well in excess of hydrostatic? Three most obvious mechanisms potentially responsible for time-dependent changes in porosity and pore space configuration are (1) inelastic pore deformation, leading to pore closure; (2) dissolution, including pressure solution, and redeposition of solutes in the pores; and (3) the creation of fractures, with their subsequent healing and sealing. As porosity changes due to these processes, so will permeability.

Possible relationships between porosity reduction and strain can be indirectly inferred from rocks that were once at considerable depths in the crust. Porosity reduction processes in rocks such as indicated by detailed studies by optical and scanning electron microscopy (e.g., Richter and Simmons, 1977a; Sprunt and Nur, 1979; Padovani *et al.*, 1982) clearly demonstrate that crack healing and sealing are quite ubiquitous in a wide variety of crustal rock types, particularly in crystalline rocks. We use the term *healing* for former cracks in which the mineral filling is the same as the host grain and the term *sealing* for cases in which the crack filling is mineralogically different from the host grain. The material source for healing is likely to be local (i.e., nearby grains), whereas sealing may require an external or more remote source of crack-filling material transported via the pore fluid.

Although there has been little laboratory work on porosity reduction in crystalline rocks, experimental studies provide some indications that porosity reduction could be relatively fast. For example, Sprunt and Nur (1976, 1977; Figure 7.7) measured appreciable porosity loss, presumably due to local pressure solution, in sandstone samples subjected to elevated temperatures (to 250°C), pressures, and macroscopic shear stress (to ~500 bars) for 2 weeks. Their results indicate that porosity reduction rates in rocks subjected to tectonically induced deviatoric stresses can be fast, suggesting that porosity reduction rates in situ may also be geologically fast.

Smith and Evans (1984) examined healing of cracks in synthetic quartz under elevated pressure ($P_f = P_c = 200$ MPa) and temperature (200° to 600°C). Morphologically, healed cracks were strikingly similar to fluid inclusions and "microtubes" commonly seen in thin sections (e.g., Richter and Simmons, 1977b). Smith and Evans found that thin cracks healed extremely rapidly (in less than 1 hr) at 400°C. It is likely, therefore, that healing and sealing are quite rapid on the geologic time scale under mid-crustal conditions.

Evans and co-workers also found that the sealing rate falls off very rapidly with increasing crack aperture, so thick cracks (>1 mm?) may seal or heal very slowly. As pointed out by Brace (Massachusetts Institute of Technology, private communication, 1988), we do not know the crack aperture versus depth in the crust. However, the ubiquitous occurrence of fully sealed cracks in exposed or exhumed mid-crustal rocks and the evidence for sealing by repeated episodes (Ramsay, 1980) suggest that, in general, sealing is widespread and fast enough to leave only bubble chains behind.

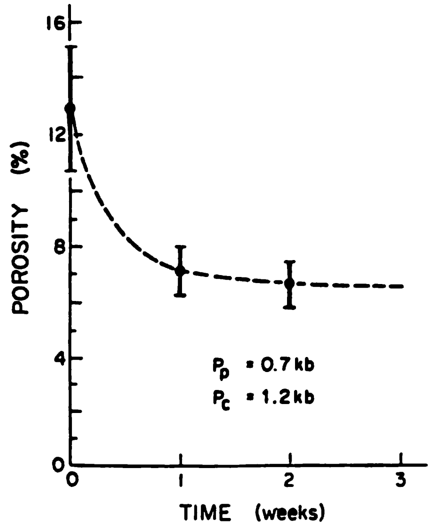

FIGURE 7.7 Experimentally induced inelastic porosity reduction in a sandstone sample, subject to an overburden pressure of 1.2 kbar and pore pressure of 0.7 kbar for a period of 2 weeks at 250°C. Note the rapid porosity decrease in a period of a week or so at this temperature (from Sprunt and Nur, 1976).

POROSITY REDUCTION WITH TIME

The discussion above implies that porosity must generally tend to decrease with time. This reduction leads to two important interrelated effects (Figure 7.8): (1) the gradual expulsion of water out of the crustal pore space (if permeability is high enough) and (2) the gradual buildup of pore pressure within the pore space of crustal rocks if permeability is sufficiently low. Accordingly, two cases (Figure 7.9) for the development of crustal porosity with time, permeability, and pore pressure can be envisioned.

Case 1

To illustrate this case, consider an element of porous crust at depth, subject to overburden stress, as shown in Figure 7.8. Due to porosity reduction processes, the pore space gradually decreases with time. If a permeable path exists between the subsurface rock element and the Earth's free surface, the fluid in the pores will gradually be squeezed out. The pore fluid pressure during this process will

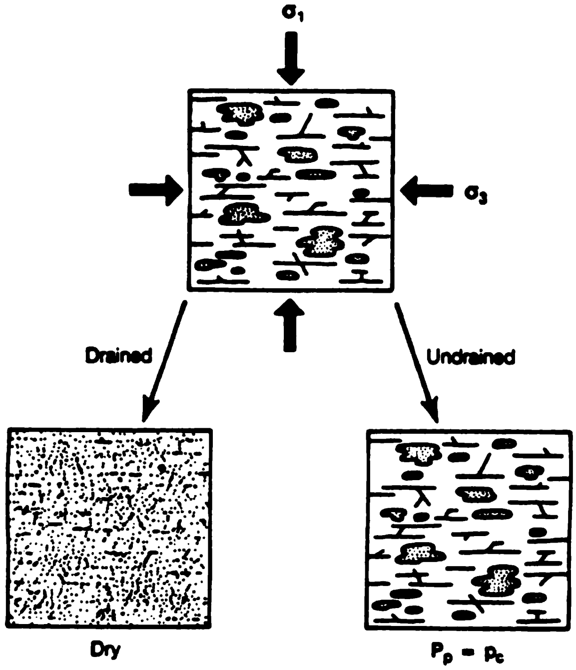

FIGURE 7.8 Cartoon illustrating the possible development of a porous fluid-saturated rock, initially saturated and subject to stress and undergoing irreversible porosity reduction. If fluid escape is sufficiently fast relative to the pore pressure buildup rate (due to porosity reduction), the rock will gradually lose its porosity and hence the pore fluid. If fluid escape is slow relative to pore pressure buildup, P_p will reach lithostatic pressure and cause natural hydraulic fractures to occur.

obviously depend on the rate of porosity reduction on the one hand and the resistance to flow through the permeable path to the surface. If the resistance is low or, equivalently, if the permeability is sufficiently high, P_p throughout the duration of the porosity reduction process will be only slightly higher than hydrostatic. Under these conditions a gradual, continuous process of porosity loss accompanied by gradual water loss will take place, ultimately leading to a dry, pore-free crust. For this condition to be satisfied, the initial permeability of the rock element must be high and the rate of permeability decrease with porosity $dk/d\phi$ must be small enough so that drainage can occur.

A process that may lead to this sort of crustal porosity and permeability reduction without major P_p buildup is crack healing, which has been investigated in the laboratory (e.g., Smith and Evans, 1984). In this process progressive elimination of pore space in the form of cracks in crystals due to mechanical closure and the healing of the bonds across the crack surfaces takes place. Under a fairly wide range of circumstances little trapped fluid remains behind and little connected pore space remains in the form of fluid inclusions.

Case 2

A drastically different development of pore pressure will take place when permeability decreases more rapidly with time than porosity due to sealing and healing (Bernabe *et al.*, 1982; Walder and Nur, 1984), so that $\dot{k}/k > \dot{\phi}$. Consider again an element of porous crustal rock at depth, again subject to overburden stress as shown in Figure 7.8. Again we expect the porosity to decrease inelastically with time. If a permeable path does not exist between the element and the free surface of the crust or the permeability of such a path is sufficiently low, the fluid in the pores will not be squeezed out fast enough, and the pore pressure of the trapped fluid will thus rise with time. If the porosity reduction rate $\dot{\phi}$ is sufficiently large, permeability k sufficiently small, and their rate of change with time or with each other are such that $\dot{k}/k > \dot{\phi}$, then pore pressure must increase with time. The rate of such pore pressure buildup will depend on the rate of porosity reduction $\dot{\phi}$ and the permeability k. Eventually P_p will reach its upper possible limit—the least normal stress acting on the element plus its cohesive strength. When P_p reaches that stress level, natural hydrofracturing will occur, involving a rapid episode of fluid release together with a sudden reduction in pore pressure. As soon as the pore pressure drops, the flow path through the hydrofracture will close again. Because the process of porosity reduction continues, P_p will again build up toward lithostatic, leading to another cycle of hydrofracturing, fluid release, and sealing, etc.

In this case, dominated by low permeability, pore pres-

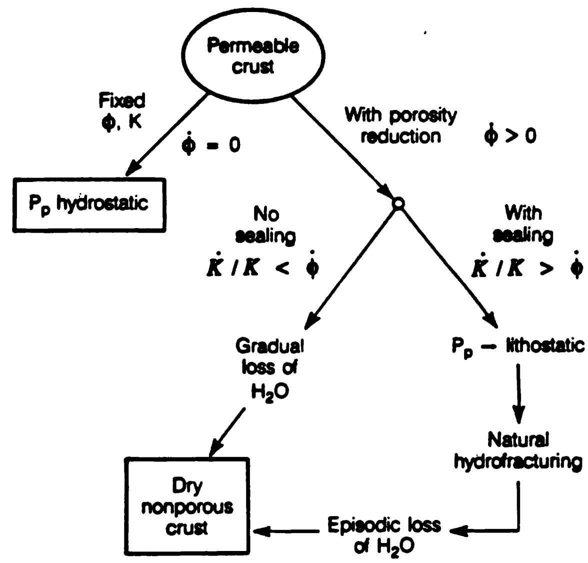

FIGURE 7.9 Flow diagram illustrating the possible paths that initially permeable, fluid-saturated crustal rocks can follow, depending on permeability, porosity, and their rates of change.

sure oscillates episodically between lithostatic and less than lithostatic, with each oscillation involving a pulse of fluid release and flow and a short episode of fracturing, followed by a longer period of little or no flow, pore pressure recovery, and sealing.

RATES AND MAGNITUDES OF PORE VOLUME STRAIN AND PORE PRESSURE BUILDUP

To determine whether the transient processes described above—the draining, drying, and porosity elimination of crustal rocks—are actually geologically important and whether they are gradual or episodic, we need to estimate four parameters: (1) the magnitude of the porosity reduction rate $\dot{\phi}$ required for P_p buildup, (2) the permeability k and its dependence on porosity to ensure P_p buildup, (3) the time required for pore pressure buildup in relevant crustal rock masses and correspondingly the duration of the P_p cycles, and (4) the time required or duration of these processes before they cease due to the elimination of connected porosity and fluid removal in situ.

Walder (1984) and Walder and Nur (1984) investigated the conditions under which lithostatic pore pressure will develop and estimated the length of time required for its buildup due to porosity reduction. By dimensional analysis they show that the controlling factor is the dimensionless grouping

$$\frac{\mu\dot{\phi}H}{\Delta\rho k g} = \dot{F}, \qquad (7.1)$$

where H is depth in the crust, $\Delta\rho$ is the difference in density between rock and fluid, μ is fluid viscosity, $\dot{\phi}$ is porosity reduction, k is hydraulic permeability, and $\dot{F}$ is the buildup index. When $\dot{F} < 1$, pore pressure development is largely unaffected by porosity reduction. For $F > 1$, fluid pressure is strongly affected by porosity reduction. Furthermore, because $\dot{F}$ increases as the depth H to which porosity reduction occurs increases, high pore pressure is more likely at greater depth. This may be further enhanced by the lower permeability k that is more likely at greater depth.

As an example, Walder and Nur (1984) considered a 10-km-thick section of granitic rock undergoing porosity reduction. For the decrease of permeability as ϕ decreases, they assumed the following relationship:

$$k = k_0 \frac{\phi^n - \phi_c^n}{\phi_0^n - \phi_c^n}, \qquad (7.2)$$

where k_0 is the initial value of permeability, ϕ_0 is the initial value of porosity, ϕ_c is the percolation threshold porosity for throughflow, and n is the exponent. Also assume that $n = 2$ and $\phi_c = 2 \times 10^{-4}$, which is well below porosities typically measured in crystalline rocks ($\phi \geq 10^{-3}$). Assuming the initial permeability of 5×10^{-20} m^2 (50 ndarcy) throughout the section (Brace *et al.*, 1968), water viscosity $\mu = 2 \times 10^{-4}$ Pa, $\Delta_\rho = 1.7 \times 10^3$ kg/m^3, Walder and Nur found that $\dot{F} = 1$ when $\dot{\phi} = 4 \times 10^{-16}$/s (Figure 7.10). In other words, if porosity reduction were to proceed throughout the 10-km-thick section at a rate higher than 4×10^{-16}/s or 1 percent per million years, excess pore pressure would be generated and maintained.

The next question we need to consider is whether the time required for pore pressure to build up to lithostatic is fast enough to be geologically important. This question is difficult to answer because few details are known about the processes involved and only little relevant experimental evidence is available. However, much of this evidence indicates that porosity and permeability reduction can be very rapid, geologically speaking, if temperatures are sufficiently high. Smith and Evans (1984) found that crack healing rates, based on a model by Evans and Charles (1977), are most likely governed by processes with activation energies around 50 to 100 kJ/mol, which in turn can be used to estimate in situ rates. Smith and Evans (1984) also found in the laboratory that crack healing in quartz requires several hours at 600°C and several days at 400°C. At the crustal depth of interest here (5 to 12 km or so) with temperatures typically ranging from 100° to 300°C, the

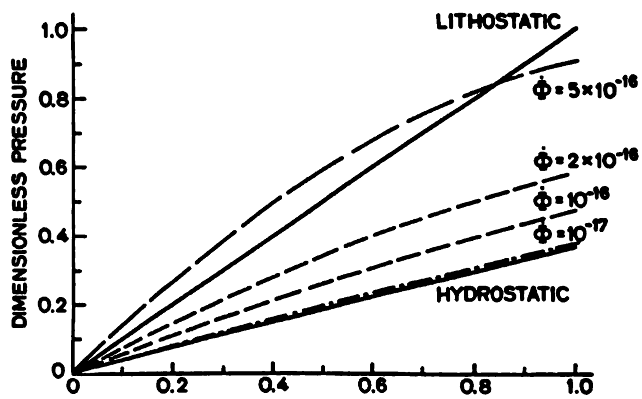

FIGURE 7.10 Fluid pressure as a function of depth for a 10-km-thick section undergoing uniform porosity reduction for several values of porosity reduction rate $\dot{\phi}$ (in s^{-1}). Initial permeability is 5×10^{-20} m^2 (50 ndarcy). Solid lines show hydrostatic and lithostatic pressure gradients. Dashed curves show fluid pressure profiles that would develop after porosity reduction for 2500 yr at indicated rates. Note that the fluid pressure would exceed lithostatic for the largest ϕ value, which slightly exceeds the "critical" value for this geometry and permeability (from Walder and Nur, 1984).

corresponding rates increase to a few hundred years. Sprunt and Nur (1977) found that porosity can be reduced significantly (by 30 to 40 percent) in laboratory experiments in sandstones subject to moderate shear stress at 250°C over a period of 2 weeks. The corresponding period of porosity reduction at 150°C would be, using the Smith and Evans activation energy, 10^2 to 10^3 longer, yielding at least several tens of years.

Walder (1984) and Walder and Nur (1984) used a different approach to estimate the minimum time for pore pressure to reach lithostatic pressure by considering the extreme case of a totally sealed rock mass. In this case the rate of pore pressure increase with time, $\Delta P/\Delta t$, is simply proportional to the rate of porosity change with time $\dot{\phi}$

$$\left.\frac{\Delta P}{\Delta T}\right|_{no\ flow} = \gamma\dot{\phi}\,, \tag{7.3}$$

where the proportionality constant γ is given by

$$\gamma = \frac{2}{\beta}\left(\frac{\beta_s}{\beta}\right)\left(\frac{1-2\nu}{1-\nu}\right), \tag{7.4}$$

where β is the compressibility of the rock, β_s is grain compressibility, and ν is the rock's Poisson's ratio, which, by using the parameters listed earlier, reduces to

$$\Delta P \approx (2 \times 10^{10}\ \mathrm{Pa}) \times \dot{\phi} \times \Delta t\,. \tag{7.5}$$

Thus, for a given porosity reduction rate $\dot{\phi}$, it is possible to estimate from Eq. (7.1) the critical permeability required to cause pore pressure to build up and from Eq. (7.5) to estimate the time for this buildup to take place. As mentioned above, it is the decrease of pore volume or porosity with time that drives the pore pressure up, when permeability is sufficiently small. It is reasonable, for example, on the basis of the results of Sprunt and Nur (1977), to expect that porosity reduction would be especially likely in porous rocks undergoing tectonic deformation (and hence subject to shear stresses). Strain rates $\dot{\varepsilon}$ associated with active crustal deformation such as in orogenesis or accretionary wedges are generally estimated at 10^{-14}/s to 10^{-11}/s (Price, 1970; Heard and Raleigh, 1972; Rutter, 1974), whereas strain rates at or near faults are much higher (Pfiffner and Ramsay, 1982; Wojtal and Mitra, 1986). If we assume that the pore volume strain rate ϕ is only a small fraction of the total strain rate (e.g., $\dot{\phi}/\dot{\varepsilon}$ = 0.01), we obtain $\dot{\phi}$ in the range of 10^{-16} to 10^{-13} s^{-1}, and for $\dot{\phi}/\dot{\varepsilon}$ = 0.10, $\dot{\phi}$ is in the range of 10^{-15} to 10^{-12} s^{-1}. Clearly, even these conservative estimates of porosity reduction rates as fractions of total strain rates are more sufficient to drive the pore pressure toward lithostatic while porosity lasts.

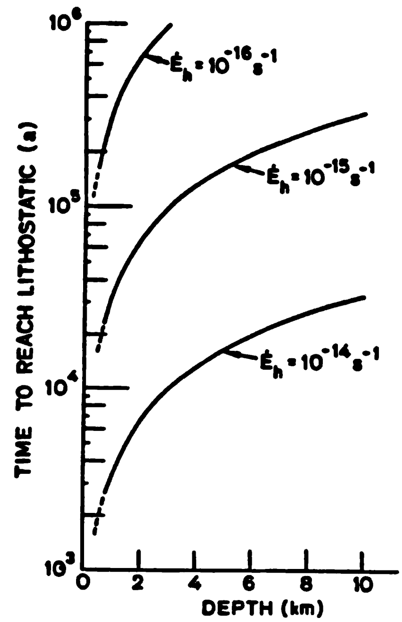

FIGURE 7.11 Time for initially hydrostatic pore pressure in a "sealed" rock mass to reach lithostatic as a result of lateral strain-induced overpressuring, as a function of depth to the rock mass, for three values of strain rate. For strain rates typical of convergent plate margins, pore pressure may reach lithostatic within only a few thousand or tens of thousands of years (from Walder, 1984).

With the above estimated ranges of $\dot{\phi}$, it is now possible to estimate the time τ_1 for P_p to reach lithostatic pressure. Figure 7.11, from Walder (1984), provides relations between time to reach lithostatic pressure τ_1, the depth, and the strain rate. Clearly, τ_1 can be surprisingly short at strain rates ϕ of 10^{-13} to 10^{-15} s^{-1}, with $\tau_1 \approx 10^3$ and 10^5 yr, respectively. Thus, if in a given crustal section subject to tectonic strain there is a sufficient thickness of porous low permeability units, lithostatic pore pressure can develop in a few thousands to tens of thousands of years.

Next we estimate how much of the total porosity must be lost to sustain these porosity reduction rates over a period τ_1. At a rate of pore volume reduction of 10^{-15} s^{-1}, porosity will decrease by 3×10^{-4} (or 0.03 percent) in 10^4 yr, by 3×10^{-3} (0.3 percent) in 10^5 yr, and by 3×10^{-2} (3 percent) in 10^6 yr. When the rate $\dot{\phi}$ is higher, the pressure buildup is faster and thus the duration τ_1 is less. From this simple analysis we can conclude that the total porosity reduction associated with one cycle of pore pressure buildup and natural hydraulic fracturing is only a small fraction of the porosity of typical crustal rocks. Consequently, maybe the porosity reduction process could involve many such cycles.

GEOLOGICAL DURATION

Next we need to determine how many cycles of P_p release can take place at a given location. For this purpose we need to estimate the duration τ_2 of the crustal drying process and the porosity elimination. One estimate can be made by assuming that the cyclic buildup and release of P_p will proceed as long as porosity remains or $\tau_2 = \dot{\phi}^{-1}$. Thus, for $\dot{\phi} = 10^{-16}$ s^{-1}, $\tau_2 \approx 300 \times 10^6$ yr, and for $\dot{\phi} = 10^{-15}$ s^{-1}, $\tau_2 \approx 30 \times 10^6$ yr. These values are probably high estimates, first because porosity may begin to lose its connectivity at some finite value, thus reducing the amount available for the reduction process considered here. Furthermore, the rate of porosity reduction may be slower with lowered porosity. But even if we allow only one-tenth of the total porosity reduction to influence pore pressure increase, we still obtain periods of 30×10^6 yr (for $\dot{\phi} = 10^{-16}$ s^{-1}) and 3×10^6 yr (for $\dot{\phi} = 10^{-15}$ s^{-1}). These shorter values for τ_2 over τ_1 yield the number of natural hydraulic fracturing episodes n

$$n \approx \frac{\tau_2}{\tau_1} \tag{7.6}$$

with $\tau_1 = 10^4$ to 10^6 yr, and $\tau_2 = 10^7$ to 10^8 yr, we obtain $n = 10$ to 10^4 events.

A different estimate of the number of P_p cycles n can be made by considering the amount of porosity reduction needed to raise P_p to lithostatic pressure per P_p cycle. Assuming that the mass of fluid in the decreasing pore space is conserved during the P_p buildup phase of each cycle and is being reduced only during the expulsion phase, we can write

$$\frac{\Delta V_f}{V_f} = \beta_f \Delta p\,, \tag{7.7}$$

where V_f is the pore fluid volume, ΔV_f is the change of pore fluid volume due to pore pressure ΔP, and β_f is the fluid's compressibility. The quantity Δp represents the magnitude of the induced P_p fluctuation during a cycle. If we ignore, as first approximation, the change of fluid density during the porosity reduction cycle, we can write

$$\frac{\Delta\phi}{\phi} \approx \frac{\Delta V_f}{V_f}\,, \tag{7.8}$$

where $\Delta\phi$ is the porosity change during a cycle. Combining Eqs. (7.7) and (7.8) we obtain

$$\frac{\Delta\phi}{\phi} \approx \beta_f \Delta p \tag{7.9}$$

taking $\beta_f = 3 \times 10^{-5}$ per bar for water, Eq. (7.9) indicates that pressure rises to lithostatic of $\Delta p = 100, 300$, and 1000 bars required 1/3, 1, and 3 percent reduction of the initial porosity, respectively. Accordingly, if these Δp values represent realistic P_p fluctuation, somewhere between 30 and 300 episodes are possible. This value is in reasonable agreement with the estimate of n based on the duration ratio of Eq. (7.6) and is consistent with crack seal layers documented by Ramsay (1980).

WATER REPLENISHED BY SUBDUCTION

The duration τ_2 of the time over which the crust dewaters could be significantly prolonged if additional waters are supplied to the crust from the lower crust or mantle below (Figure 7.12). The most obvious possible source of such waters is associated with subducted oceanic slabs. It is generally thought that the oceanic slab is rich in water and hydrous minerals and that this water plays major roles in controlling the onset, amount, and rate of melting in and above the downgoing slab. A recent study by McGeary *et al.* (1985) showed, for example, that most of the prominent gaps in the circum Pacific active volcanic chains are associated with the subduction of anomalous oceanic crust, usually in the form of thick-rooted oceanic rises. One explanation for these gaps is that the rises somehow reduce the supply of water to the melt zones, thus raising the

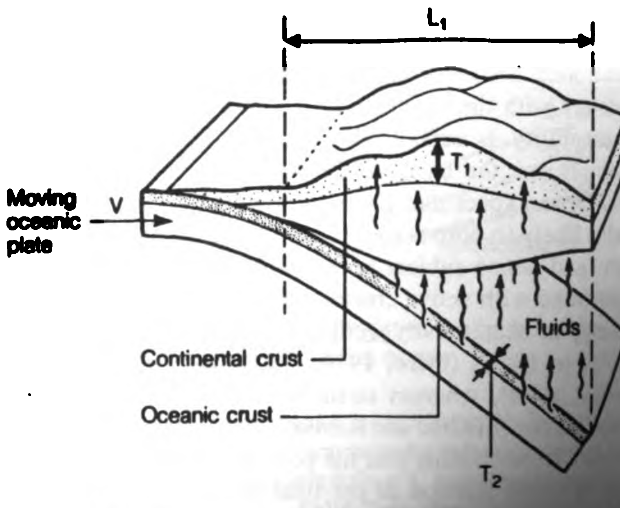

FIGURE 7.12 Cartoon illustrating the supply of s... into the crust overlying a subduction zone ... parameters that determine the magnitude o... its effects on prolonging the episo... crust.

TABLE 7.2 The Main Time Constants for Time-Dependent Hydraulic Behavior of the Crust

Duration of a P_p cycle	10^3 to 10^5 yr
Time for crust to dry up	10^6 to 10^7 yr
Duration of subduction	10^7 to 10^8 yr

melting temperature and causing a temporal cessation or reduction of volcanism. This model suggests that the subduction of normal oceanic crust does provide for a continuous supply of water into crustal regions below which subduction is taking place. Most probably this water is initially stored in the heavily fractured oceanic lithosphere, some of which migrates upon subduction into the overlying lithosphere and crust.

Constraints on the amount of water that might be added this way to the crust overlying a slab may be obtained from a comparison between the pore fluid volume in the crust and the pore fluid volume that passes underneath while subduction lasts. The volume of fluid in the continental crust V_c per unit length of subduction zone (Figure 7.12) is roughly $V_c = T_1 D\phi$, where T_1 is thickness, D is the width of the crustal region overlying the subducted slab and tectonically affected by the subduction process, and ϕ is the average porosity over the thickness T.

In the continental crust we consider the pore fluid to be present in the top $T_1 = 12$ km, with the average porosity $\phi = 1$ percent. With $D = 300$ km as a representative length of the crust overlying the subducting slab and which is tectonically affected by the subduction process, the volume of water present per unit length of the crust is on the order of 30 km^2. The total volume of fluid that is subducted with the oceanic crust per unit subduction zone over the duration of subduction is $V_0 = \mathrm{v} \times \tau \times T_2 \times \phi_0$, where v is subduction rate, τ_3 is the subduction duration (and $\mathrm{v} \times \tau_3 = L_2$ is the total length of slab subducted), T_2 is the thickness of the porous oceanic crust, and ϕ_0 is its average porosity. For the subducted oceanic crust we take crustal thickness $T_0 = 6$ km, average porosity also 1 percent, duration of subduction of 100 Ma, and subduction rate of 5 cm/yr, which yields a volume of subducted free water per unit width of slab on the order of 300 km^2. This value of V_0 is about 10 times V_c, which suggests that the waters released from the slab at depth can replenish dissipated crustal water, adding as much as several times the volume of initial fluid to the continental crust. As a result, the duration of the cycles of pore pressure buildup, natural hydrofracturing, and sealing may continue for longer periods of time up to the duration of the subduction process, on the order of 10^3 yr (see Table 7.2).

DISCUSSION AND CONCLUSIONS

The original question posed in this chapter was whether free water is generally deep in the crust. As it turns out this question is inseparable from the question of the nature of porosity, permeability, and pore pressure in the crust and, most significantly, their dependence on time, as illustrated in Figure 7.9. If we *assume* that porosity and, consequently, permeability and pore pressure are time invariant, and consider the hydrological evidence for easy fluid flow in the crust, it follows that crustal pore pressure must be generally hydrostatic to a depth of 10, 15, or even 20 km. This conclusion is in conflict with geological and laboratory evidence that implies that permeable paths in rock tend to clog very rapidly by healing, sealing, and inelastic deformation, all of which become very effective in crustal rocks even at moderately elevated temperatures.

One way to reconcile the conflict between rapid flow on the one hand and rapid clogging on the other is to consider crustal hydrological behavior as varying with time. Although direct data are very sparse, the analysis in this chapter suggests that the most likely state of crustal porosity and water content is transient, with both porosity and water content as well as permeability decreasing with time. It is quite unlikely that these quantities are constant over geological time because inelastic deformation below a depth of a few kilometers in the crust must tend to cause pore closure. Such a tendency will be accompanied by squeezing of water out of the crust due to induced pore pressure. If crustal permeabilities are high enough and the rate of permeability decreases with time or porosity is sufficiently small, the dewatering of the crust will be a gradual, monotonic process. However, if permeability is low and its rate of reduction with porosity is relatively fast, the pore pressure of the trapped fluid will rapidly rise to overcome the least principal stress, leading to natural spontaneous hydrofracturing accompanied by the pulsed release of water and the loss of a little porosity (Figure 7.13). This is followed by a drop in pore pressure, a prolonged buildup period, another hydrofracturing episode, etc. It is especially intriguing that the episodic release process is most probable when tectonic deformation, even at fairly low strain rates, is taking place.

Estimates of the time required for P_p to reach lithostatic and the amount of porosity reduction, especially in tectonically active areas, suggest that this kind of episodic hydrological and mechanical behavior of the crust is quite probable. The number of cycles expected at a given site depends on strain rate, permeability, depth, and other rock parameters, but simple analysis suggests that tens to hundreds of cycles may be expected. The duration of each cycle is estimated at 10^3 to 10^5 yr, whereas the duration of the crustal dewatering process is estimated at 10^6 to 10^7 yr if no additional waters are supplied to the crust. Such supply may be common where the ocean lithosphere was or is being subducted underneath the system. In such situations the geologic duration of the cyclic pore pressure behavior may last as long as subduction does, on the order of 10^8 yr.

The likelihood that the Earth's crust experiences cyclical pore pressure buildup as suggested by Norton and Knight (1977) in conjunction with igneous intrusions, with P_p magnitude oscillating somewhere above hydrostatic to lithostatic, has profound implications for crustal processes and our understanding of these processes. One such process is the formation of hydrothermal ore deposits and mineralized veins (Cathles and Smith, 1983; Vrolijk, 1987). Extensive geological evidence indicates that several types of ore deposits are formed by periodic precipitation from brines. Similarly many types of veins in rocks have been shown to have formed from repeated episodic precipitation from fluids in fractures (Ramsay, 1980). The cyclic pore pressure behavior described in this chapter provides a very simple and compelling mechanism for the formation of these bodies. The cyclic fluctuation in crustal pore pressure may also play an intriguing role in the control and initiation of large earthquakes, which tend to be cyclic in time. In the past it has often been assumed that the repeat time of large earthquakes is basically controlled by the rate at which tectonic strain is accumulating, with fault strength essentially constant. More recently it has become apparent that some sort of time-dependent fault strength is required, if only to reconcile laboratory rock failure results with in situ fault rupture. The increase of P_p toward lithostatic during a pore pressure cycle as described above provides a simple mechanism for cyclic fault weakening, which in turn leads to cyclic failure. Unlike pore pressure changes associated with elastic and elastic dilatant deformation (e.g., Nur and Booker, 1971; Nur, 1972; Rice, 1975, 1979; Rudnicki, 1977) in which P_p buildup is very sensitive to geometrical details, pore pressure buildup due to the inelastic process considered here is fairly insensitive to the details of the stress field buildup around the impending failure zone. Rather it is a robust process of sealing and trapping of the pore fluid in and around the fault zone and the gradual buildup of P_p due to porosity decrease until fault rupture begins. A related effect of this process has been suggested by Sibson *et al.* (1975) in which inelastic deformation and dilatancy work together to induce seismic pumping, which enhances the instability during failure.

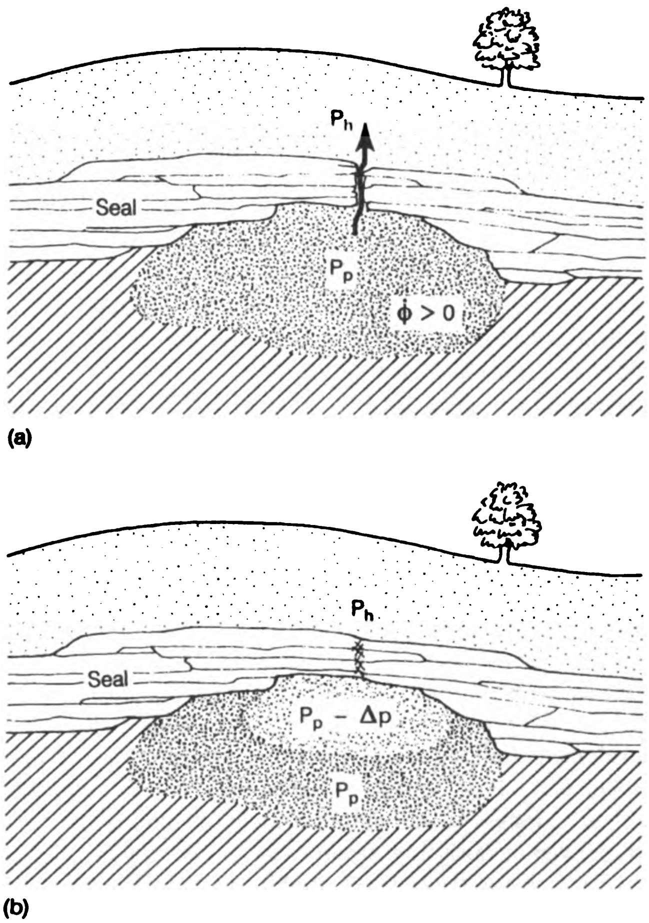

FIGURE 7.13 Cartoon illustrating the pore pressure buildup and release envisioned in this chapter. (a) Pore pressure in the region of porosity reduction reached the minimum compressive stress and caused natural fractures to occur in the sealed rock surrounding this region. Some fluid escapes through these fractures into the overlying lower pore pressure (P_h) region, causing a drop ΔP in pore pressure (dotted region) before (b) the induced fractures close again.

A third manifestation of the presence of high P_p may be deep crustal reflectors. As discussed by Eaton (1980), laboratory results (Nur and Simmons, 1969; Todd and Simmons, 1972) and field observations (Berry and Mair, 1977) together have suggested that high pore pressure zones in the crust would show up as seismic low-velocity zones. Eaton further suggested that such zones may come about when crustal waters are trapped under a permeability barrier or seal. Jones and Nur (1984; Figure 7.14) and Walder and Nur (1984) showed that such a seal can be effective but only when its permeability is maintained at very low values. Because such a seal is most likely to be broken repeatedly in areas subject to earthquakes, it can be effective only if healing and sealing processes are continuously active. It is quite possible that the process we outlined in this chapter is therefore responsible for dynamically trapping high pore pressure zones in mid-crustal depth. These zones will last as pronounced seismic reflec-

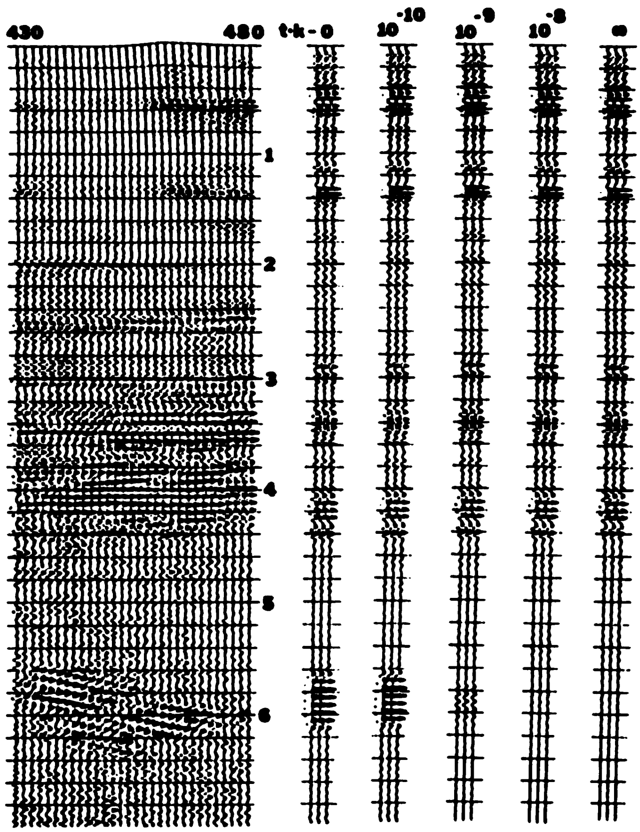

FIGURE 7.14 The possible role of high pore pressure zones on crustal reflections (from Jones and Nur, 1984). Portion of the Wind River line 1 on the left, showing reflections from the Pacific Creek thrust at 6 s. Synthetics computed for a series of zones 110 m in thickness, alternating ones that have an initial lithostatic pore pressure. Several cases show the change in reflectivity as the pore pressure diffuses out as the product *tk* increases.

tors as long as enough water remains in the pore space. The possibility that crustal seismic reflectors may be high pore pressure zones and hence mechanically weak is especially intriguing in view of the growing evidence that these reflectors may represent subhorizontal crustal detachment zones. Such detachments are mechanically very difficult to explain unless high pore pressure is actually involved. Finally, Oliver (1986; Chapter 8, this volume) suggested that deep crustal fluids can migrate horizontally over very large distances away from consumption or collision zones. Our proposed processes of pore pressure development, and its consequent episodic behavior, especially in continental crust overlying subducted slabs, provide a mechanism that can drive fluids horizontally, the pore pressure needed to sustain such flow, and the low-permeability barrier needed to prevent the escape of pressurized brines upward.

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