Chapter 19
Hydraulic Pulses in the Earth's Crust

Amos Nur[1] *and Joseph Walder*[2]
[1] *Department of Geophysics, Stanford University, Stanford, CA 94305, U.S.A.*
[2] *U.S. Geological Survey, Cascades Volcano Observatory, 5400 MacArthur Blvd., Vancouver, WA 98661, U.S.A.*

Abstract

The analysis in this chapter attempts to quantify the coupling between pore space deformation and pore pressure in the earth's crust. Clearly, the state of free water deep in the crust is inseparable from the state of porosity, permeability, and pore pressure. If porosity and permeability are time-invariant, crustal pore pressure at depth must be close to hydrostatic. This is in conflict with evidence for ubiquitous high pore pressure at depth. Consequently crustal porosity, permeability, and water content must decrease with time, as pores close inelastically in the crust. When crustal permeability is low, and its rate of decrease with porosity is large, pore pressure P_p of trapped fluids must rise with time until spontaneous hydrofracturing, pulsed release of water, and the loss of some porosity occurs, followed by a drop in pore pressure and sealing of the system. After a P_p buildup period, another hydrofracturing episode occurs. Because inelastic pore deformation is enhanced by shear stress, this episodic process is most probable in tectonically active areas. Dimensional analysis shows that tens to hundreds of P_p episodes or 'burps,' each lasting 10^3 - 10^5 years, can occur before the crust dewaters in 10^6–10^7 years. Oceanic crust subduction may prolong this process to 10^8 years or more.

High pore pressure can explain deep crustal seismic reflectors and detachments, and large-scale horizontal fluid migration in the crust. Episodic pore pressure buildup and release in the crust can explain the formation of hydrothermal ore deposits and mineralized veins, and possibly some aspects of the mechanics of very large earthquakes.

1. Introduction

It was W.F. Brace's (1980) paper on crustal permeability that brought into focus one of the simplest yet most profound questions about the state of the earth's crust: what

FAULT MECHANICS AND TRANSPORT PROPERTIES OF ROCKS
ISBN 0-12-243780-2

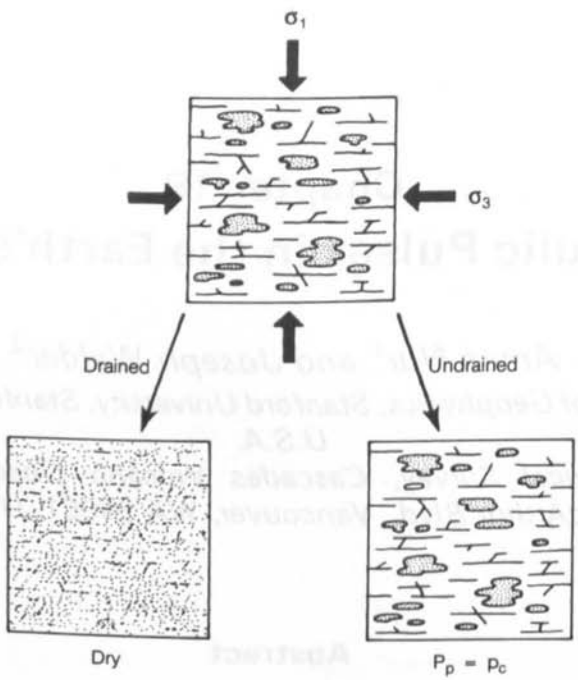

Figure 1. An initially stressed, water-saturated rock mass, with pore pressure P_p less than confining stress $P_c = \sigma_{ii}/3$, can either dry up and lose its porosity gradually if drainage is fast enough (lower left block), or, if drainage is too slow, pore pressure will with time reach the least confining stress.

is the depth of free water in the crust? This porosity is in turn inseparable from the water's ambient pore pressure P_p (Figure 1) which, as discussed below, is in turn inseparably linked to permeability.

In this chapter we attempt to answer quantitatively the three interlinked questions: (1) What is the depth and pressure of free water in the crust? (2) What are typical crustal porosity and permeability values? and (3) How do these parameters vary with time and over what timescale do they vary?

Because direct measurements of crustal hydrologic properties are confined to shallow depths, the answers to these questions can at present be based only on indirect evidence, some of which is listed in Table 1. For example, oxygen isotopic data (e.g., Taylor, 1977; Norton and Taylor, 1979) indicates water circulation down to 20 km. Deep zones of low electrical resistivity, inferred from crustal electromagnetic soundings, also suggest the presence of water (e.g., Nekut et al., 1977; Thompson et al., 1983). Seismic low-velocity zones (Feng and McEvilly, 1983) and low-seismic Q zones in the crust (Herrmann and Mitchell, 1975) are also consistent with pore pressure close to lithostatic (Nur and Simmons, 1969; Berry and Mair, 1977; Winkler and Nur, 1982). Reflections from deep crystal fault zones, especially subhorizontal detachments, may be due to zones of high P_p (Jones and Nur, 1984).

Fyfe et al. (1978) and Etheridge et al. (1984) reviewed geologic evidence for widespread free water with high P_p during low- to medium-grade regional metamorphism. The

Table 1. Possible indicators for the presence of free water in the earth's crust as suggested by the various referenced investigators. Also given are estimated depth values associated with these indicators

Indicator	Depth range (km)	Reference
Water table	0–2	Kozlovsky (1986)
Deep wells (i.e., the Kola Well)	to 12	Kozlovsky (1986)
Reservoir-induced seismicity	to 12	Bell and Nur (1978)
Crustal low-velocity zones	7–12	Berry and Mair (1977)
		Feng and McEvilly (1983)
		Jones and Nur (1984)
Crustal electrical conductivity zones	10–20	Nekut et al. (1977)
		Shankland and Ander (1983)
Oxygen isotopes	to 20	Taylor (1977)
Metamorphism	>20	Fyfe et al. (1978)
		Etheridge et al. (1984)
Crack healing and sealing	?	Richter and Simmons (1977)
		Sprunt and Nur (1976, 1977, 1979)
		Ramsay (1980)
		Smith and Evans (1984)
Formation of hydrothermal ore deposits	>5	Norton and Knight (1977)
		Cathles and Smith (1983)
Crustal seismic attenuation zones	7–15	Herrmann and Mitchell (1975)
		Winkler and Nur (1982)
Low stress on faults	0–10	Raleigh and Evernden (1982)
		Rice (this volume, Chapter 20)

ubiquity of mineralized fractures (Ramsay, 1980) also requires P_p to exceed the minimum principal stress during fracture formation and suggests widespread high P_p.

In order to maintain elevated P_p in the crust for geologically significant periods of time, the permeability of crustal rocks must be very low, down to 10^{-21} m^2 (1 ndarcy) or less (e.g., Bredehoeft and Hanshaw, 1968; Hanshaw and Bredehoeft, 1968). Such low permeability values are in remarkable disagreement with the conclusion of Brace (1980), who showed, on the basis of numerous geologic indicators, that fluid flow in the crust must be rapid and that crustal permeability must consequently be 10^{-5} m^2 (1 mdarcy) or higher to depths of 10 km or so. He concluded that P_p generally is not likely to exceed hydrostatic pressure at depth in the crust. This conclusion is in apparent conflict with the evidence for widespread, high crustal P_p.

In this chapter we model crustal hydraulics specifically to try to reconcile the two conflicting lines of evidence regarding crustal hydrology: ubiquitous water, high pore pressure, and the consequent low permeability on the one hand, and evidence for deep circulation and rapid flow on the other.

2. Time Dependence of Porosity and Permeability

The most common approach to the hydraulics of the crust is based on the (often unstated) assumption that porosity and permeability are time-invariant. Under this assumption, hydraulic permeability measured in situ is used to directly estimate typical ambient crustal permeability values. Brace's (1980; Brace and Walsh, 1984) extensive compilations thus lead to typical values of crustal permeabilities ranging from 100 darcy for shallow fractured rock masses, to 10^{-5} or 10^{-6} darcy at depth. With these permeabilities, crustal pore pressure can generally not be sustained above hydrostatic for geologically significant time.

The apparent contradiction between the indirect evidence for elevated crust P_p and the high permeability throughout the crust can therefore not be reconciled if porosity and permeability are assumed constant with time. However, if porosity, permeability, and consequently pore pressure, vary significantly with time, the two seemingly conflicting lines of evidence may be reconciled. There is in fact abundant, although indirect, evidence for episodic P_p pulses, associated with fluid flow during magmatic intrusions (Norton and Taylor, 1979), composite crack-filled veins, presumably episodically precipitated from solution (Ramsay, 1980), and isotopic variations across veins (Shemesh et al., 1992). For porosity, permeability, and pore pressure to vary with time, inelastic pore space variations are required. Tectonically or pore-pressure-induced fractures are the most likely mechanism for porosity to increase with time and pore pressure to decrease. The question that immediately arises is whether there is enough porosity deep in the crust, and whether it can decrease rapidly enough with time to produce large, lasting pore pressure increases. And how does permeability change through time to maintain P_p well in excess of hydrostatic?

Three obvious classes of mechanisms for porosity reduction, and consequently permeability reduction, are (1) plastic pore closure; (2) stress-induced dissolution and redeposition in the pores ('pressure solution') (e.g., Richter and Simmons, 1977; Sprunt and Nur, 1976, 1977, 1979; Padovani et al., 1982); and (3) crack healing (where mineral filling is the same as the host grain) and sealing (where crack filling is mineralogically different from the host grain).

Laboratory results strongly suggest that porosity reduction can be relatively fast (Smith and Evans, 1984) under mid-crustal conditions, perhaps especially so when rock is subject to deviatoric stress (Sprunt and Nur, 1976, 1977). However, two drastically different paths for the decrease of crustal porosity and permeability with time and the associated pore pressure changes can be envisioned, depending on the rate of porosity reduction and on the permeability (Figure 1).

Case 1. High crustal permeability. Owing to porosity reduction processes, the pore space at depth gradually decreases with time. If a permeable path exists to the earth's free surface, the fluid in the pore space will gradually be squeezed out so that P_p during this process will remain close to hydrostatic. Ultimately, this process leads to a dry, pore-free crust. Clearly, for this gradual dewatering at hydrostatic P_p to occur, the initial permeability of the rock system must be sufficiently high, and the rate of permeability decrease with decreasing porosity relatively small.

Case 2. Low permeability. In contrast, when porosity at depth decreases with time, but the hydraulic connectivity to the free surface is sufficiently low, the pore fluid cannot escape fast enough, and its pressure rises. However, this buildup of P_p is limited by the least normal stress (plus the tensile strength) of rock at depth. When P_p reaches this limiting stress level, natural hydrofracturing will occur which creates a short lived but significant hydraulic path. This in turn will cause a short episode of fluid expulsion, or

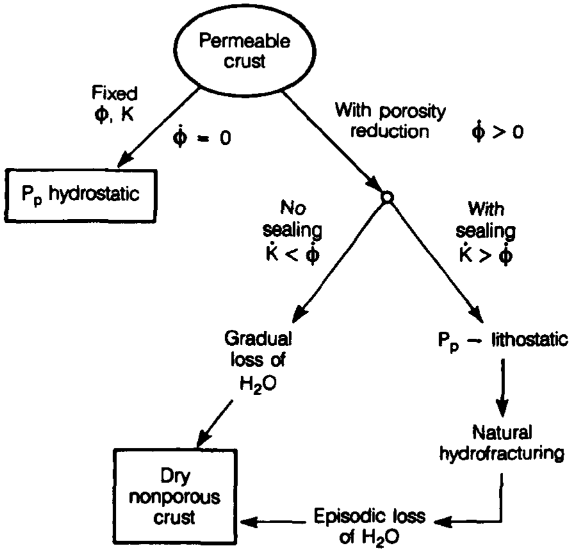

Figure 2. Different pore space/pore pressure paths for crustal rock mass depending on the rates of (1) porosity reduction $\dot{\phi}$ and (2) permeability reduction $\dot{K}$. When porosity ϕ and permeability K are fixed ($\dot{\phi} = 0$ and $\dot{K} = 0$), crustal pore pressure remains hydrostatic. It is much more likely, however, that porosity and hence permeability decrease with time. If permeability is low and its reduction rate is fast, episodic pore pressure buildup and release by repeated hydrofracturing is most likely, eventually leading to a dry, nonporous crust.

a pressure pulse, accompanied by a rapid partial pore pressure drop. Because of this drop, the hydrofracture closes, and consequently the system returns to its low prefac permeability state. Following that, as porosity reduction continues, P_p again builds up gradually to initiate another episode of hydrofracturing, fluid release, sealing, etc.

Thus porosity reduction in low-permeability systems involves episodic variation of pore pressure between lithostatic and less than lithostatic values in which each episode leads to a pulse of fluid release and rapid flow during a short period hydraulic fracturing and fluid release, followed by a long period of no or little flow, during which pore pressure recovers. The two different paths that pore pressure may follow in crustal rock is summarized in Figure 2.

3. Rates and Magnitudes of Pore Volume Strain and Pore Pressure Buildup

Next we try to determine quantitatively (1) whether actual crustal porosity reduction and its rate are hydrologically important, and (2) whether porosity reduction is typically gradual or episodic. Specifically, we need to somehow estimate four parameters: (1) the rate of porosity reduction $\dot{\phi}$ required for P_p buildup; (2) the permeability values k and their rates of change with porosity, $dk/k\phi$, in situ; (3) the time required for pore pressure buildup and hence, the duration of the P_p cycles; and (4) the time required for the elimination of all connected porosity.

Walder and Nur (1984) have derived the following equation for pore pressure P_p:

$$\frac{1}{c}\frac{\partial P_p}{\partial t} = \Delta^2 P_p + \frac{\mu}{k}\dot{\phi} \tag{1}$$

where c is the hydraulic diffusivity $c = k/\mu\phi(\beta_f + \beta_\phi)$, β_f is pore fluid compressibility, β_ϕ is the *elastic* pore volume compressibility, μ is fluid viscosity, ϕ is porosity, $\dot{\phi} = \partial\phi/\partial t$ is the *inelastic* porosity reduction with time, $k = k(\phi)$ is hydraulic permeability, and t is time.

In the simple case of flow in the vertical direction, z, only, and using the scaling $P_p = P_0\bar{P}$, $t = t_0 \cdot \bar{t}$, and $z = h \cdot \bar{z}$, where P_0 is the maximum allowable P_p, $t_0 = h^2/c$, h is the depth to which porosity reduction occurs, and dimensionless variables are denoted by bars, eq. (1) is reduced to (Walder, 1984)

$$\frac{\partial \bar{p}}{\partial \bar{t}} = \frac{\partial^2 \bar{p}}{\partial \bar{z}^2} + \frac{\mu h}{\Delta\rho g}\left(\frac{\dot{\phi}}{k}\right)$$

It follows that the time for P_p buildup due to porosity reduction is controlled by the factor F,

$$F = \frac{\mu h}{\Delta\rho g}\frac{\dot{\phi}}{k} \tag{2}$$

where $\Delta\rho$ is the difference in density between rock and fluid. When $F < 1$, pore pressure remains close to hydrostatic during porosity reduction. In contrast, when $F \gg 1$, fluid pressure increases significantly above hydrostatic during porosity reduction. Furthermore, because F increases with depth h, induced high pore pressure is more likely at greater depth. This is further enhanced by the lower permeability k that is common at greater depth.

As an example, Walder and Nur (1984) considered a 10-km section of granitic crust undergoing porosity reduction, with the permeability-porosity relationship

$$k = k_0\left\{\frac{\phi^2 - \phi_c^2}{\phi_0^2 - \phi_c^2}\right\}$$

where k_0 and ϕ_0 are the initial permeability and porosity, respectively, and ϕ_c is the percolation threshold porosity at which flow can first occur. With $\phi_c = 2 \times 10^{-4} \cdot k_0 = 5 \times 10^{-20}\,\text{m}^2$ (50 ndarcy), $\mu = 2 \times 10^{-4}\,\text{Pa}$, $\Delta\rho = 1.7 \times 10^3\,\text{kg}\,\text{m}^{-3}$, $F = 1$ when $\dot{\phi} = 4 \times 10^{-16}\,\text{s}^{-1}$. In other words, this result suggests that were porosity reduction $\dot{\phi}$ to proceed throughout the 10-km-thick section at a rate lower than $4 \times 10^{-16}\,\text{s}^{-1}$ or about 1% million years, pore pressure would remain hydrostatic. In contrast, when $\dot{\phi}$ is greater than $4 \times 10^{-16}\,\text{s}^{-1}$, high pore pressure must develop, leading to pulsed, episodic releases.

Is the time required for pore pressure buildup to lithostatic fast enough to be geologically important? As mentioned above, the meager evidence which does exist indicates that porosity and permeability reduction can be geologically rapid if temperatures are sufficiently high. For example, Smith and Evans (1984) and Brantley et al. (1990) found that crack healing in quartz requires several hours at 600°C and several days at 400°C. Even at crustal depths of 5–12 km with temperatures between 100°C and 250°C, crack lifetimes increase to only hundreds of years. Furthermore, Sprunt and Nur (1977) found that porosity can be reduced by 30–40% in sandstones subject to moderate shear stress at 250°C in 2 weeks. The corresponding period of porosity reduction at 150°C, using (for lack of any more direct data) the Smith and Evans (1984) activation energy of 50–100 kJ mol^{-1}, would be 10^2–10^3 years.

The minimum time for pore pressure to reach lithostatic can be estimated for the extreme case of a hydraulically sealed rock system. Equation (1) for this case shows that the rate of pore pressure increase with time is simply proportional to the rate of relative porosity change with time:

$$\left.\frac{\Delta p}{\Delta t}\right|_{\text{no flow}} \cong \frac{\dot{\phi}}{\phi} \tag{3}$$

and the time τ_1 for pressure buildup to rupture, Δp, is

$$\tau_1 \cong \beta_f \phi \frac{\Delta P_p}{\dot{\phi}} \tag{4}$$

Taking $\Delta P_p = 200$ bar, $\beta_f = 5 \times 10^{-5}\,\text{bar}^{-1}$, $\phi = 5 \times 10^{-3}$, and $\dot{\phi} = 3 \times 10^{-16}\,\text{s}^{-1}$, we obtain $\Delta t \cong 10^{12}$ s, or 4×10^4 years. This buildup time can be much shorter, of course, if porosity reduction rate $\dot{\phi}$ is larger than $3 \times 10^{-16}\,\text{s}^{-1}$, when the system porosity ϕ is smaller than 5×10^{-3}, or when the pore pressure fluctuation is significantly less than 200 bar. For example for $\Delta p = 50$ bar, $\phi = 2 \times 10^{-3}$, and $\dot{\phi} = 10^{-15}\,\text{s}^{-1}$, we find $\Delta t = 10^{10}$ s, or only 1500 years. we will assume therefore that typical value of time τ_1 required for P_p to increase say from hydrostatic to lithostatic pressure is on the order of $\tau_1 = 10^3$–10^5 years.

The porosity reduction during one cycle of pressure rise and decay for $\phi = 10^{-15}\,\text{s}^{-1}$ is around 3×10^{-4} (or 0.03%) for $\tau_1 = 10^4$ years, and 3×10^{-3} (0.3%) for $\tau_1 = 10^5$ years. This reduction during one episode of pore pressure buildup and hydraulic fracturing is therefore only a fraction of the total porosity. Consequently, the porosity reduction process in a low-permeability crust can involve a large number of episodes, as illustrated schematically in Figure 3.

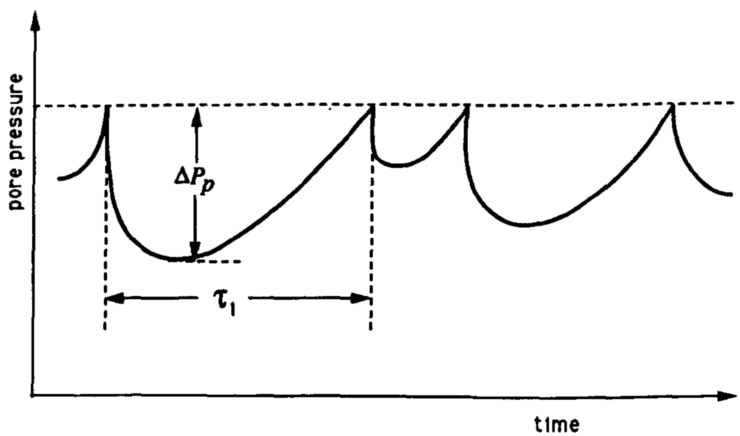

Figure 3. Idealized variations of pore pressure with time in a crust undergoing episodic pulses. Typical values for the duration of each episode are $\tau = 10^2$–10^4 years, and the pore pressure fluctuations $\Delta P_p = 100$–1000 atm. The time required to drain this kind of crust will be on the order of 10^5–10^7 years, involving tens to hundreds of episodes.

4. Geological Duration

The number of hydraulic pulses, involving P_p buildup, hydrofracturing, and release, can be estimated in two ways. Assuming that pulses will continue as long as porosity remains, we can define the duration $\tau_2 = \dot{\phi}^{-1}$. Thus for $\dot{\phi} = 10^{-16}\,\mathrm{s}^{-1}$, $\tau_2 = 300 \times 10^6$ years; and for $\dot{\phi} = 10^{-15}\,\mathrm{s}^{-1}$, $\tau_2 = 30 \times 10^6$ years. These estimates are probably unrealistically high, because porosity may lose its connectivity at some finite value, thus reducing the amount available for the reduction process considered here. But even if only 1/10 of the total porosity is involved, durations of $\tau_2 = 30 \times 10^6$ years (for $\dot{\phi} = 10^{-16}\,\mathrm{s}^{-1}$) to $\tau_2 = 3 \times 10^6$ years (for $\dot{\phi} = 10^{-15}\,\mathrm{s}^{-1}$) are expected.

Clearly, τ_2 is much longer than τ_1, the duration of a single pore pressure cycle. The number n of fluid release episodes is then simply

$$n = \frac{\tau_2}{\tau_1} \tag{5}$$

Taking $\tau_1 = 10^4$–10^6 years, and $\tau_2 = 10^7$–10^8 years, we find the range for $n = 10$–10^4 cycles.

The number of episodes can be estimated also from bounds on the magnitude of the pressure rise, ΔP_p during each cycle. Assuming that the pore fluid mass is conserved during the P_p buildup phase in each cycle (and is reduced only during the expulsion phase), we have

$$\frac{\Delta V_f}{V_f} = \beta_f \cdot \Delta p \tag{6}$$

where V_f is the pore fluid volume, ΔV_f is the change of V_f due to pore pressure Δp, β_f is the fluid's compressibility, and Δp is the induced P_p change during the buildup in one cycle. Neglecting the changes in fluid and grain density during the pore pressure cycle, we can write

$$\frac{\Delta\phi}{\phi} = \frac{\Delta V_f}{V_f} \tag{7}$$

where $\Delta\phi$ is the porosity change during a cycle. Combining eqs. (6) and (7) yields

$$\frac{\Delta\phi}{\phi} \approx \beta_f \cdot \Delta p \tag{8}$$

taking $\beta_f = 5 \times 10^{-5}\,\mathrm{bar}^{-1}$ for water, eq. (8) indicates that pressure rises of $\Delta p = 100$, 300, and 1000 bar require 0.5%, 1.5% and 5% reduction of the initial porosity, respectively. Accordingly, if these ΔP_p values represent realistic P_p fluctuations, somewhere between 20 and 200 episodes are needed before porosity is eliminated and the episodes of fluid expulsion cease. This value is in reasonable agreement with the estimate n of eq. (5). It is also of the same order of magnitude as the number of crack seal layers observed in situ in some cases (e.g., Ramsay, 1980).

5. Water Replenished by Subduction

The duration τ_2 of crustal dewatering could be significantly prolonged if water were added to the crust, most likely from underlying subducting oceanic slabs (Figure 4) rich in water and/or hydrous minerals (i.e., McGeary et al., 1985).

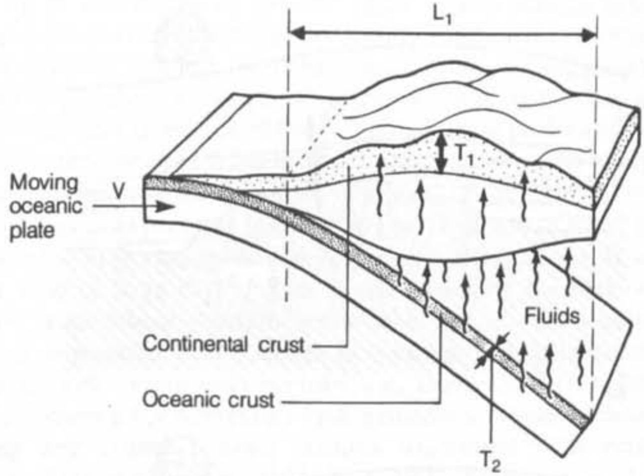

Figure 4. The addition of fluids from subducted lithosphere can prolong the lifespan of the restless, 'burping' overlying crust to be on the order of the duration of subduction – 10^8 years or more.

Table 2. The main time constants for the time-dependent hydraulic behavior of the crust

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

The amount of water added to the crust from a slab can be roughly estimated from the pore fluid fluxes involved. The volume V_c of fluid per unit width in the continental crust overlying subduction (Figure 4) is roughly $V_c = T_1 L_1 \phi_1$, where T_1 is the thickness of the fluid-bearing crust, L_1 is the length of the crust involved, and ϕ_1 is the average porosity over the thickness T_1. Taking $T_1 = 10$ km, porosity $\phi_1 = 1\%$, $L_1 = 300$ km, the volume of water per unit width in this crust is 30 km^2.

The volume V_0 of fluid subducted per unit width with the oceanic crust during the duration of subduction τ_3 is $V_0 = V\tau_3 T_2 \phi_2$, where V is subduction velocity, T_0 is the thickness of the porous oceanic crust, and ϕ_2 is its average porosity. Taking $T_2 = 5$ km, $T_2 = 1\%$, $\tau_3 = 10^8$ years, and $V = 5\,\mathrm{cm\,y^{-1}}$, the volume of free water per unit width of subducted slab is on the order of 250 km^2, about 10 times V_c. This suggests that waters released from slabs at depth can add several times the initial pore fluid volume to the continental crust. Consequently, crustal 'burping' may last as long as typical subduction, so that $\tau_2 = \tau_3 = 10^8$ years, as summarized in Table 2.

6. Discussion and Conclusion

We have attempted to show that the question of the presence of free water deep in the crust is inseparable from the state of porosity, permeability and pore pressure in the crust, and especially their time dependence. If we *assume* that porosity and permeability

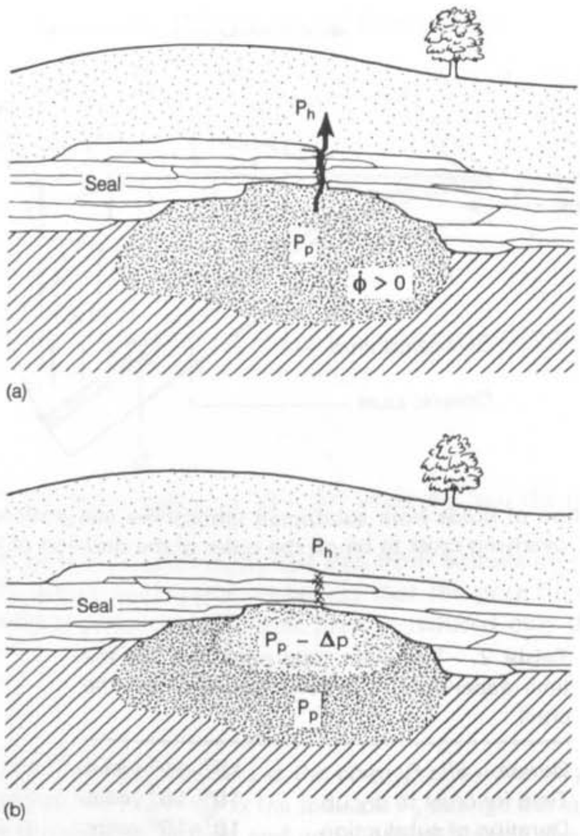

Figure 5. Schematic depiction of the process of natural hydrofracturing, local pore pressure reduction, hydrofracture sealing, and pore pressure buildup again.

are time-invariant, it follows that crustal pore pressure at depth must be generally hydrostatic. This conclusion is in conflict with geologic and laboratory evidence for rapid healing and sealing of flow paths, even at moderate temperatures.

It is therefore necessary to consider crustal hydrologic behavior which is time-varying (Figures 1 and 2). Our analysis suggests that crustal porosity, permeability, and water content in fact must decrease with time, as pores close inelastically at depth in the crust. When crustal permeabilities are high and the decrease of porosity with time is sufficiently small, the dewatering of the crust will be gradual, and pore pressure will remain near hydrostatic. However, as is more likely, if permeability is low and its rate of decrease with porosity or time is large, the pore pressure of the trapped fluid will rise rapidly. When the least principal stress is overcome, spontaneous hydrofracturing will occur. This will be accompanied by the pulsed release of water and the loss of some porosity, followed by a drop in pore pressure and sealing of the system (Figure 5). After a buildup period, another hydrofracturing episode will occur, and so on, as shown in Figure 3. Because inelastic pore deformation is enhanced or even driven by deviatoric stress, the episodic release process is most probable in tectonically active areas. Tens to hundreds of P_p buildup and release episodes, each with 10^3-10^5 years' duration, can occur before the crust dries up by dewatering in 10^6–10^7 years. Where oceanic lithosphere is being subducted, the process may last as long as subduction itself, on the order of 10^8 years.

Episodic pore-pressure buildup and release in the crust was suggested by Norton and Knight (1977) in conjunction with igneous instrusions, and by Cathles and Smith (1983) and Vrolijk (1987) for the formation of hydrothermal ore deposits and mineralized veins formed by periodic precipitation from brines (Ramsay, 1980). The analysis described in this chapter provides some of the quantitative aspects of the essential coupling between rock deformation and pore pressure involved in the formation of these geologic features.

Large earthquakes may also involve pore pressure fluctuations. It is often assumed that the repeat time of large earthquakes is controlled by the rate of tectonic strain accumulating on faults whose strength is constant. But it is now quite apparent that some sort of time-dependent fault strength is required, if only to reconcile laboratory rock failure results with in situ fault rupture (e.g., Dieterich, 1978). The increase of P_p towards lithostatic during a pore pressure cycle provides a simple mechanism for episodic fault weakening and failure. Unlike changes associated with elastic or dilatant deformation (e.g., Nur and Booker, 1971; Nur, 1972; Sibson et al., 1975, 1988; Rice, 1975, 1979; Rudnicki, 1977; McCaig, 1988; Byerlee, 1990), the pore pressure buildup due to the inelastic processes considered here is fairly insensitive to the details of the stress field accumulation around impending failure zones. Instead, we envision a robust process of sealing and trapping of the pore fluid in and around the fault zone, leading to the gradual buildup of P_p until the fault ruptures.

High pore pressure may be the cause of deep crustal seismic reflectors. Eaton (1980) and Jones and Nur (1984) on the basis of laboratory results (Nur and Simmons, 1969; Todd and Simmons, 1972) and field observation (Berry and Mair, 1977) suggested that pressurized crustal water trapped under permeability barriers could show up as seismic low-velocity zones. This possibility is especially intriguing in view of the growing evidence that these reflectors may be correlated with subhorizontal crustal detachment zones, which are mechanically very difficult to explain without high pore pressure being responsible for their low strength. Finally, Oliver (1986) has suggested that deep crustal fluids may migrate horizontally in the crust over very large distances away from subduction zones. The proposed process of high pore pressure development provides a mechanism which can drive fluids horizontally, sustain this drive, and account for the low-permeability barrier that is required to prevent the escape of pressurized brines upward.

Acknowledgments

The ideas and speculation presented in this chapter are the result of collaborations and discussions with many students and colleagues for over a decade or two. Many of these ideas here can be traced back in one form or another to work by Eve Sprunt, Terry Jones, Lee Bell, Ken Winkler, Sue McGeary, Mark Zoback, Bill Brace, Gene Simmons, Brian Evans, Jim Byerlee, Mike Etheridge, D. Norton, Rick Sibson, and others. We are grateful also for many important suggestions made by the two (anonymous) reviewers of this chapter. The funding for the research reported here was provided by the Division of Mathematical and Physical Sciences of the U.S. Department of Energy.

References

Bell, M.L. and Nur, A. (1978). Strength changes due to reservoir-induced pore pressure and stresses and application to Lake Oroville. *J. Geophys. Res.* **83**(B9), 4469–4483.

Berry, M.J. and Mair, J.A. (1977). The nature of the earth's crust in Canada. In *The Earth's Crust: Its Nature and Physical Properties* (ed. J.G. Heacock), pp. 319–348, Geophysical Monograph Series. American Geophysical Union, Washington, D.C.

Brantley, S.L., Evans, B., Hickman, S.H., and Crerar, D.A. (1990). Healing of microcracks in quartz: Implications for fluid flow. *Geology* **18**, 136–139.

Brace, W.F. (1980). Permeability of crystalline and argillaceous rocks. *Int. J. Rock Mech. Min. Sci. Geomech. Abstr.* **17**, 241–251.

Brace, W.F. and Walsh, J.B. (1984). The effect of pressure on porosity and the transport properties of rock. *J. Geophys. Res.* **89**(B11), 9425–9431.

Bredehoeft, J.D. and Hanshaw, B.B. (1968). On the maintenance of anomalous fluid pressures, I. Thick sedimentary sequences. *Geol. Soc. Am. Bull.* **79**, 1097–1106.

Bylerlee, J. (1990). Friction, overpressure and fault normal compression. *Geophys. Res. Lett.* **17**, 2109–2112.

Cathles, L.M. and Smith, A.T. (1983). Thermal constraints on the formation of Mississippi Valley-type lead-zinc deposits and their implications for episodic dewatering and deposit genesis. *Econ. Geol.* **78**, 983–1002.

Dieterich, J.H. (1978). Time-dependent friction and the mechanics of stick–slip. *Pure Appl. Geophys.* **116**, 790–806.

Eaton, G.P. (1980). Geophysical and geological characteristics of the crust of the Basin and range Province. In *Continental Tectonics* (eds. B.C. Burchfield, J.E. Oliver and L.T. Silver). National Academy of Science, Washington, D.C.

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

Feng, R. and McEvilly, T.V. (1983). Interpretation of seismic reflection profiling data for the structure of the San Andreas Fault zone. *Seismol. Soc. Am. Bull.* **73**(6), 1701–1720.

Fyfe, W.S., Price, N.J., and Thompson, A.B. (1978). *Fluids in the Earth's Crust*. Elsevier, Amsterdam.

Hanshaw, B.B. and Bredehoeft, J.D. (1968). On the Maintenance of anomalous fluid pressure, II. Source layer at depth. *Geol. Soc. Am. Bull.* **79**, 1107–1122.

Herrmann, R.B. and Mitchell, B.J. (1975). Statistical analysis and interpretation of surface-wave anelastic attenuation data for the stable interior of North America. *Seismol. Soc. Am. Bull.* **65**(5), 1115–1128.

Jones, T.D. and Nur, A. (1984). The nature of seismic reflections from deep crustal fault zones. *J. Geophys. Res.* **89**(B5), 3153–3171.

Kozlovsky, Ye. A. (ed.) (1986). *The Superdeep Well of the Kola Penisula*, Springer-Verlag, Berlin, 1–55.

McCaig, A.M. (1988). Deep fluid circulation in fault zones. *Geology* **16**, 867–870.

McGeary, S., Nur, A., and Ben-Avraham, Z. (1985). Spatial gaps in arc volcanism: the effect of collision or subduction of oceanic plateaus, *Tectonophysics* **119**, 195–221.

Nekut, A., Connerney, J.E.P., and Kuckes, A.F. (1977). Deep crustal electrical conductivity: Evidence for water in the lower crust. *Geophys. Res. Lett.* **4**, 239–242.

Norton, D. and Knight, J. (1977). Transport phenomena in hydrothermal systems: Cooling plutons. *Am. J. Sci.* **277**, 937–981.

Norton, D. and Taylor, H.P. Jr. (1979). Quantitative simulation of the hydrothermal systems of crystallizing magmas on the basis of transport theory and oxygen isotope data: An analysis of the Skaergaard intrusion. *J. Petrol.* **20**, 421–486.

Nur, A. (1972). Dilatancy, pore fluids, and premonitory variations of t_s/t_p travel times. *Seismol. Soc. Am. Bull.* **62**(5), 1217–1222.

Nur, A. and Booker, J.R. (1971). Aftershocks caused by pore fluid flow? *Science* **175**, 885–877.

Nur, A. and Simmons, G. (1969). The effect of saturation on velocity in low porosity rocks. *Earth Planet. Sci. Lett.* **7**, 183–193.

Oliver, J. (1986). Fluids expelled tectonically from orogenic belts: Their role in hydrocarbon migration and other geologic phenomena. *Geology* **14**, 99–102.

Padovani, E.R., Shirey, S.B., and Simmons, G. (1982). Characteristics of microcracks in amphibolite and granulite facies grade rocks from southeastern Pennsylvania. *J. Geophys. Res.* **87**, 8605–8630.

Raleigh, C.B. and Evernden, J. (1982). Case for low deviatoric stress in the lithosphere. In *Mechanical Behavior of Crustal Rocks* (eds. N.L. Carter, M. Freidman, J.M. Logan, and D.W. Stearns), pp. 173–186, Geophysical Monograph Series. American Geophysical Union, Washington, D.C.

Ramsay, J.G. (1980). The crack-seal mechanism of rock deformation. *Nature* **284**, 135–139.

Rice, J.R. (1975). On the stability of dilatant hardening for saturated rock masses. *J. Geophys. Res.* **80**(11), 1531.

Rice, J.R. (1979). Earthquake precursory effects due to pore fluid stabilization of a weakening fault zone. *J. Geophys. Res.* **84**(B5), 2177.

Richter, D. and Simmons, G. (1977). Microcracks in crustal igneous rocks: Microscopy. In *The Earth's Crust: Its Nature and Physical Properties* (ed. J.G. Heacock), pp. 149–180, Geophysical Monograph Series. American Geophysical Union, Washington, D.C.

Rudnicki, J.W. (1977). The inception of faulting in a rock mass with a weakened zone. *J. Geophys. Res.* **82**(5), 844.

Shankland, T.J. and Ander, M.E. (1983). Electrical conductivity, temperature, and fluids in the lower crust. *J. Geophys. Res.* **88**, 9475–9484.

Shemesh, A., Ron, H., Erel, Y., Kolodny, Y. and Nur. A. (1992). Isotopic composition of vein calsite and its fluid inclusions: Implication to paleohydrological systems, tectonic events and vein formation processes. *Chem. Geol.* **94**(4).

Sibson, R.H., Moore, J.M., and Rankin, A.H. (1975). Seismic pumping – a hydrothermal fluid transport mechanism. *J. Geol. Soc. London* **131**, 653–659.

Sibson, R.H., Roberts, F., and Paulsen, K.H. (1988). High-angle reverse faults, fluid-pressure cycling, and mesothermal gold-quartz deposits. *Geology* **16**, 551–555.

Smith, D.L., and Evans, B. (1984). Diffusional crack healing in quartz. *J. Geophys. Res.* **89**, 4125–4135.

Sprunt, E.S. and Nur, A. (1976). Reduction of porosity by pressure solution: experimental verification. *Geology* **4**, 463–466.

Sprunt, E.S. and Nur, A. (1977). Destruction of porosity through pressure solution. *Geophysics* **42**, 726–741.

Sprunt, E.S. and Nur, A. (1979). Microcracking and healing in granites: New evidence from cathodoluminescence. *Science* **205**, 495–497.

Taylor, H.P. Jr. (1977). Water/rock interaction and the origin of H_2O in granitic batholiths. *J. Geol. Soc. London* **133**, 509–588.

Thompson, B.G., Nekut, A., and Kuckes, A.F. (1983). A deep crustal electromagnetic sounding in the Georgia Piedmont. *J. Geophys. Res.* **88**, 9461–9473.

Todd, T. and Simmons, G. (1972). Effect of pore pressure on the velocity of compressional waves in low-porosity rocks. *J. Geophys. Res.* **77**, 3731–3743.

Vrolijk, P. (1987). Tectonically driven fluid flow in the Kodiak accretionary complex, Alaska. *Geology* **15**, 466–469.

Walder, J.S. (1984). Coupling between fluid flow and deformation in porous crustal rocks. Ph.D. thesis, Stanford University.

Walder, J. and Nur, A. (1984). Porosity reduction and crustal pore pressure development. *J. Geophys. Res.* **89**(B13), 11539–11548.

Winkler, K.W. and Nur, A. (1982). Seismic attenuation: effects of pore fluids and frictional sliding. *Geophysics* **47**(1), 1–15.