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Permeability and porosity of the Illinois UPH 3 drillhole granite and a comparison with other deep drillhole rocks

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Abstract. Permeability, porosity, and volumetric strain measurements were conducted on granite cores obtained at depths of 0.7 to 1.6 km from the Illinois UPH 3 drillhole at effective confining pressures from 5 to 100 MPa. Initial permeabilities were in the range of 10^{-17} to 10^{-19} m^2 and dropped rapidly with applied pressure to values between 10^{-20} and 10^{-24} m^2 at 100 MPa, typical of other deep granite core samples. These values are several decades lower than equivalent weathered surface granites at comparable effective confining pressures, where weathering products in cracks and pores inhibit crack closure with applied pressure. Permeabilities of the Illinois cores were inversely related to sample depth, suggesting that stress relief and thermal microfractures induced during core retrieval dominated the fluid flow. Thus these samples provide an upper bound on in situ matrix permeability values. A comparison of core permeability from UPH 3 and other deep drillholes shows that stress relief damage can often dominate laboratory permeability measurements. We conclude that it may be difficult to make meaningful estimates of in situ permeability based on either borehole samples (possible damage during retrieval) or surface-derived analogs (altered by weathering). Volumetric strain determined from porosity measurements was compared with differential strain analysis (DSA) data reported by other investigators on samples from the same depths in the drillhole. Our strain measurements (0.002 to 0.005 at 100 MPa) were nearly twice as large as the DSA values, probably because of the crack-enhancing effects of fluids present in our samples that are absent in the dry DSA cores, as well as other time-dependent deformation effects. This difference in observed strain magnitudes between the two measurement methods may be an important consideration if strain and/or porosity data from deep core samples are used in models of stress, fluid circulation, and excess fluid pressure generation in the midcrust.

Introduction

The physical properties of crustal rocks are strongly influenced by the presence of cracks and fractures. At depth, fractures are generated by tectonic processes or stress relief from the erosion of overburden, while at the same time the competing process of healing and sealing by circulating hydrothermal fluids counteracts fracture generation. Fractures are also introduced by sampling, such as drilling damage, blasting, or stress relief and thermal microfracturing of deep core samples. Many techniques have been described which attempt to distinguish these induced fractures from naturally occurring ones [e.g., *Morrow et al.*, 1994a; *Kowallis and Wang*, 1983]. Recent comparisons of the physical properties of deep borehole rocks with their surface equivalents [*Morrow and Lockner*, 1994] show that surface weathering and retrograde metamorphism may result in crack morphologies that do not reflect the crack properties of the rock at depth. Porosity, permeability, and other transport properties of deep rocks are found to be more sensitive to pressure than the surface-derived samples upon which many models of crustal processes are based, with implications for such diverse concerns as the circulation of fluids at depth, heat flow, and the strength of shear zones.

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The Illinois UPH 3 borehole revealed much information on the relation between in situ stress, permeability, fractures, and other characteristics of midcontinent basement granitic rocks [*Carlson and Wang*, 1986; *Haimson and Doe*, 1983; *Kowallis and Wang*, 1983; *Kowallis et al.*, 1987]. Downhole permeability studies showed that permeability decreased with depth more rapidly than expected, independent of downhole fracture density [*Haimson and Doe*, 1983]. In addition, by comparing downhole permeabilities to typical laboratory values, Haimson and Doe concluded that the top of the granite sequence (670 m depth), which was at the surface during the Precambrian, must have decreased in permeability because of secondary mineralization since burial by sediments. Although the permeability of these rocks was not studied in detail in the laboratory, the observations of permeability sensitivity and the comparison to surface granites are consistent with *Morrow and Lockner*'s [1994] results from other borehole studies. Consequently, the UPH 3 core samples have been revisited in an effort to learn more about the permeability and porosity characteristics of these rocks relative to other deeply buried granites. In addition, the Illinois samples provide our first opportunity to study the permeability and porosity behavior of large grain-sized samples (up to 15 mm) to compare with more typical fine-grained granites and ultrafine mafic rocks from many other drillholes. Finally, we wish to determine how volumetric strain calculated from porosity reduction (this study) compares to volumetric strain measured by differential strain analysis (DSA), on Illinois UPH 3 samples from the same depths [*Kowallis and Wang*, 1983; *Carlson and Wang*,

Table 1. Sample Parameters

depth* m	density gm/cm^3	porosity ϕ	P_p† MPa	P_c† MPa	α	γ†† 10^{-2} (MPa)$^{-1}$
751.3V	2.635	0.0052	7.37	19.38	20.7	3.3 - 6.8
751.3H	2.623	0.0064			26.9	3.1 - 6.7
1010.4V	2.621	0.0079	9.91	26.06	13.8	3.4 - 6.7
1010.4H	2.628	0.0127			42.5	4.7 - 6.5
1451.9V	2.629	0.0086	14.24	37.45	16.6	3.6 - 6.4
1451.9H	2.634	0.0072			12.5	4.4 - 6.9
1604.9V	2.703	0.0136	15.74	41.39	14.0	1.8 - 3.5

*including core orientation vertical (V) or horizontal (H).
†Estimated in situ pore (P_p) and confining (P_c) pressures based on depth and average sample density. $P_e = P_c - P_p$.
††Stress sensitivity of permeability (1) for P_e from 5 to 100 MPa.

1986]. In these earlier studies, differential strain analysis, in conjunction with borehole televiewer images, optical microcrack studies, and ultrasonic velocity measurements, was used to determine the principal crack strain directions, microcrack porosity, and the porosity distribution of crack closure pressures in the Illinois cores. Thus the strain comparisons on like samples provide a valuable check on different strain measurement methods in standard use.

Measurement Technique

The samples used in this study (Table 1) were coarse grained granites of uniform texture with grain sizes from 2 to 15 mm. Intergranular and intragranular cracks are visible in hand specimen. The three shallowest samples were similar in appearance and contained abundant orthoclase, the deepest sample was plagioclase-rich. Cylindrical cores were prepared from the original UPH 3 cores. These test specimens were 2.54 cm in diameter and in length and were made in both a vertical and horizontal direction (labeled V and H in Table 1). The 1604.9 m horizontal sample was too fractured for use and was not tested. This may partly reflect the fact that the UPH 3 core showed frequent disking below 1602 m.

Permeabilities and volumetric strains were determined at room temperature. First, fluid pressure (water) and confining pressure were increased simultaneously until fluid pressure reached the estimated in situ value based on sample depth. With pore pressure fixed, confining pressure was then further increased in a stepwise fashion to produce effective pressures ($P_e = P_c - P_p$) of 5 to 100 MPa. At each effective pressure step, steady state flow was established through the sample for 1 to 3 days by maintaining the inlet pore pressure of the sample at 1 MPa above the outlet pressure. Permeability was then calculated from flow rate data, sample geometry, and fluid viscosity according to Darcy's law. The steady state method was chosen rather than the pulse-decay technique because it is a direct measurement requiring no assumptions about fluid and rock compressibilities [see *Brace et al.*, 1968], and is suitable for low-permeability measurements if the temperature of the system is carefully controlled to around ±0.02°C. With this method, the lower-permeability limit of the system was 2 x 10^{-24} m^2. Accuracy of the measurements varied with permeability and was approximately ±5% for values above 10^{-20} m^2; ±10% for values between 10^{-20} and 10^{-22} m^2; and around ±20% below 10^{-22} m^2. After each confining pressure increase, the volume of water expelled from the sample due to crack closure was measured. This volume discharge, ranging between 1 x 10^{-3} and 1.5 x 10^{-2} cm^3, typically appeared as a large instantaneous response, followed by a more gradual time-dependent component as the sample adjusted to the higher pressure. At low effective pressures, equilibrium was generally reached after several minutes, with little time-dependent contribution. At higher effective pressures, the time-dependent volume change often lasted for hours or days. Accuracy of the volume discharge measurements was approximately ±5 x 10^{-5} cm^3. Porosity change was determined from these volume discharge measurements and initial porosity measurements of all samples. Accuracy of the initial porosity measurements was approximately ±0.05%.

Results

Permeability

Permeability values for effective pressures to 100 MPa (Figure 1) ranged over 7 orders of magnitude (10^{-16} to 10^{-23} m^2) and decreased rapidly with applied pressure, typical of other deep core samples [*Morrow and Lockner*, 1994]. The large grain size (2 to 15 mm) of the granite did not appear to be a factor; permeability values were comparable to other more fine-grained granites [*Brace et al.*, 1968; *Morrow et al.*, 1994a; *Morrow and Lockner*, 1994]. Anisotropy between the vertical and horizontal directions was minimal for the 751.3 m sample but varied by up to 2 orders of magnitude for the deeper samples, diverging as pressures increased. The 751.3 m vertical core was retested after an interval of several weeks to show the effect of stress cycling. Permeability values were slightly lower during the second set of measurements, indicating a permanent change in crack configuration typical of other stress-cycled granites [*Morrow et al.*, 1986].

The permeability values of the UPH 3 samples decreased more rapidly with applied pressure than most surface-derived granites such as Westerly or Barre granites [*Brace et al.*, 1968], reaching values of around 10^{-21} m^2 after only a few tens of MPa of applied pressure, compared to hundreds of MPa for typical surface samples. This pressure sensitivity of

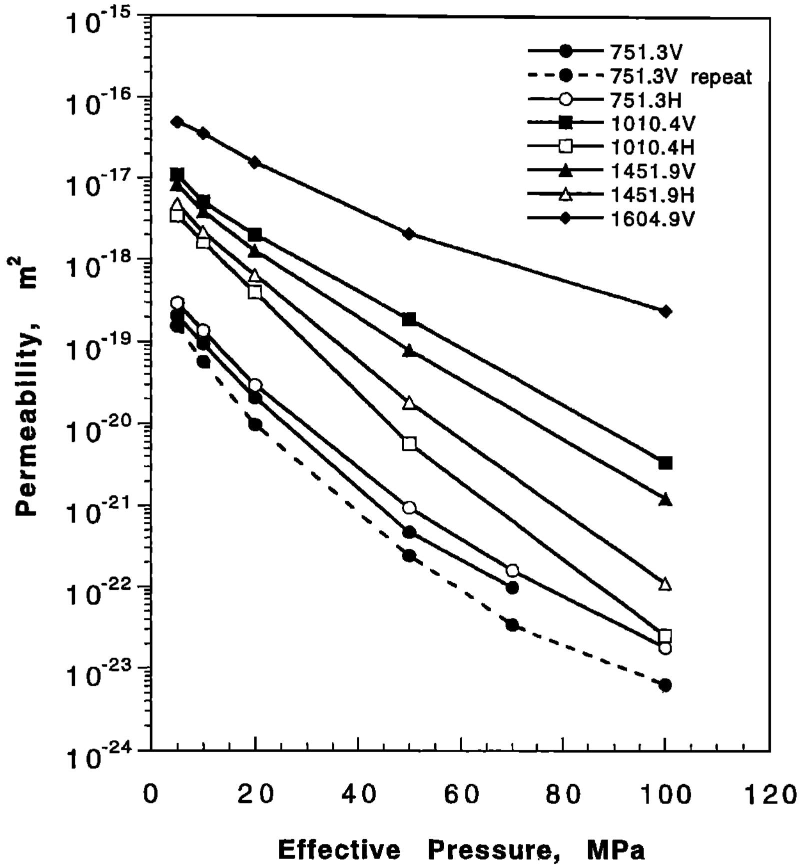

Figure 1. Permeability of Illinois UPH 3 drillhole samples as a function of effective pressure. Sample depth in meters for vertical (V) and horizontal (H) cores.

permeability is quantified by the parameter γ, defined as

$$\gamma = -\delta \log_{10} k / \delta P_e, \quad (1)$$

where k is permeability and P_e ($= P_c - P_p$) is effective pressure. The parameter γ was in the range of 0.018 to 0.069/MPa for the UPH 3 samples (Table 1), and except for the shallowest cores, was relatively independent of pressure above 10 MPa. These values are similar to those of other deep core samples, including both fracture-free and stress relief fractured samples of various rock types [*Morrow and Lockner*, 1994]. In contrast, γ is usually around 0.002 to 0.020/MPa for surface quarried granites [*Morrow and Lockner*, 1994]. Pressure sensitivity typically lessens with increasing pressure in surface-derived samples as the cracks become propped open and mismatched by secondary mineralization due to weathering, making them stiffer and more difficult to close with pressure [*Morrow and Lockner*, 1994]. Pressure sensitivity of permeability is an important parameter in many geophysical models incorporating transport properties of rock such as models of heat flow, excess fluid pressure generation, compaction, and electrical resistivity response.

Note that with one exception, the permeability curves in Figure 1 shift to higher values with increasing depth. This trend holds regardless of the applied pressure, suggesting that increased amounts of stress relief and thermal microcracking with depth of retrieval had a dominating effect on the overall permeability level in the cores, as has been observed in many other low-porosity rocks from deep drillholes [*Morrow et al.*, 1994a; *Morrow and Lockner*, 1994]. To further illustrate, we plot permeability at the estimated in situ effective pressures of the samples together with downhole permeabilities determined from packer tests at the drillhole [*Haimson and Doe*, 1983] in Figure 2. The laboratory values fall in the range 10^{-17} to 10^{-19} m^2 and show a poor overall correlation with the decreasing permeability/depth trend of the downhole data. Permeabilities of the shallowest (751.3 m) samples are well below the downhole values, probably because the downhole tests were conducted in a region of high fracture density [*Haimson and Doe*, 1983] whereas the core samples would be more representative of the matrix permeability, which is typically 2 to 3 orders of magnitude lower than in situ values [e.g., *Brace*, 1980]. Permeabilities of the 1010.4 and 1451.9 m samples fall within the range of the downhole data (but above the 751.3-m values), whereas the deepest core (1604.9 m) permeability is higher than the in situ trend, reflecting the many stress relief fractures clearly visible in hand specimen of this sample. This is in contrast to the particularly low fracture density in the drillhole below 1300 m [*Haimson and Doe*, 1983]. The overall trend of the laboratory data demonstrates that (at least for the three deepest samples), measured permeability is dominated by fractures introduced during coring and retrieval, giving an upper bound on the in situ matrix permeability values. This observation has important consequences for models of midcrustal processes based on laboratory permeability data where stress relief fractures clearly dominate the fluid flow.

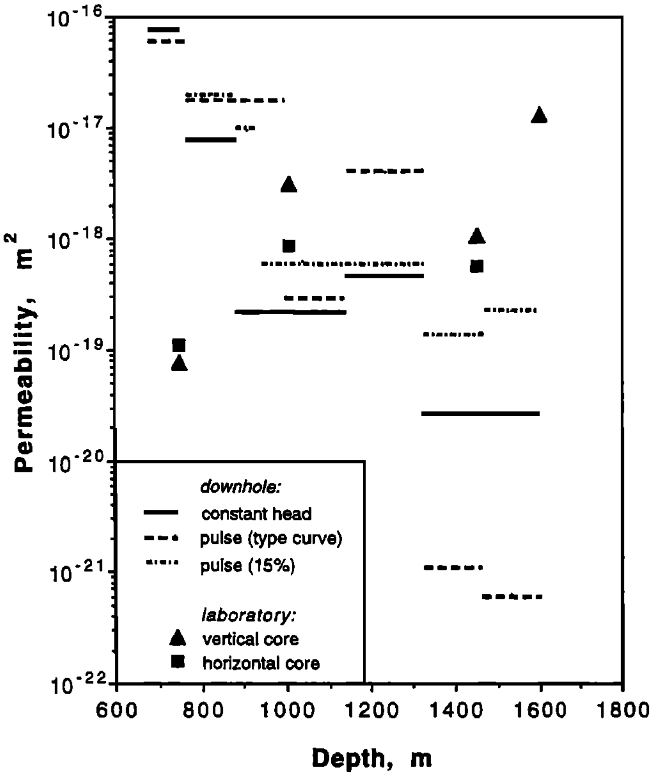

Figure 2. Permeability (from Figure 1) of core samples at estimated in situ effective pressures (closed symbols, this study) and downhole permeability determined from constant head and pulse flow tests for selected depth intervals (solid or dashed lines, modified from *Haimson and Doe* [1983]). See *Haimson and Doe* [1983] for a discussion of uncertainty errors for the three types of downhole flow tests.

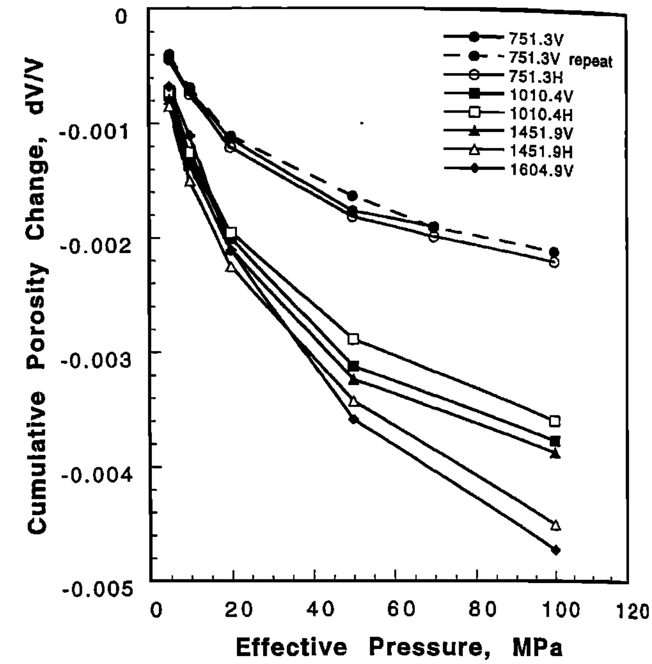

Figure 3. Cumulative porosity change of Illinois UPH 3 core samples as a function of effective pressure. Sample depths in meters.

Porosity

Cumulative porosity change as pressure is increased between permeability measurements is shown in Figure 3, with porosity values plotted at the higher pressure for each confining pressure step. Porosity measurements show little difference between vertical and horizontal core directions compared to the permeability anisotropy because the porosity change is a volumetric, rather than unidirectional measurement. Again, note the correlation with sample depth, suggesting that stress relief cracking has created additional porosity in the deeper rocks which then is reduced as the samples are repressurized in laboratory tests.

Crack characteristics of the samples are often quantified by comparing permeability with porosity change (Figure 4) through a power law of the form

$$k = k_0(\phi/\phi_0)^{\alpha} \qquad (2)$$

where k and k_0 are permeability values corresponding to porosities ϕ and ϕ_0, respectively, and α is defined as the porosity sensitivity exponent. This exponent is an important indicator of porosity/flow behavior. For the Illinois cores, α ranges from 12 to 42 (Table 1), a slightly higher range than the values of 3 to 25 reported by *David et al.* [1994] for surface granites and sandstones. It is not entirely clear what effect stress relief fracturing may have on porosity sensitivity in our Illinois samples, except to note that higher α values are also common among nonfractured rocks, particularly at very low porosities [*Zhu et al.*, 1995].

The higher α values of the Illinois samples and slight curvature of the permeability-porosity plot (Figure 4) suggest that the permeability-porosity relation may in fact be more complex than that described in (2). Recent laboratory studies on hot pressed calcite, quartz, and naturally lithified sandstone [*Lockner and Evans*, 1995; *Zhu et al.*, 1995; *Zhang et al.*, 1994a, b] indicate that there are distinct regimes within the permeability-porosity relationship; at high porosities the power law (2) applies with $\alpha \approx 3$, but at porosities below some crossover value ($\phi \approx$7-15%), permeability reduces rapidly until some limiting value of porosity, ϕ_r, (percolation threshold or disconnected porosity) is reached below which there is no fluid flow, following the relation

$$k = (\phi - \phi_r)^{\alpha}. \qquad (3)$$

The Illinois UPH 3 data (Figure 4) supports the interpretation (3), but because all of our porosity data lie within the narrow range of 0.2 to 1.0%, residual and crossover porosity values are difficult to determine.

Discussion

Comparison of Porosity Change and Differential Strain Analysis Data

The permeability and porosity data described above reveal much about the nature of the cracks within the core samples and their response to applied pressure. This type of information can also be gleaned from DSA of the core samples [*Siegfried and Simmons*, 1978; *Simmons et al.*, 1974]. Differential strain analysis can be used to determine the principal values and orientation of the strain tensor of a rock sample as well as volumetric crack porosity and the porosity distribution of crack closure pressures. Our porosity change

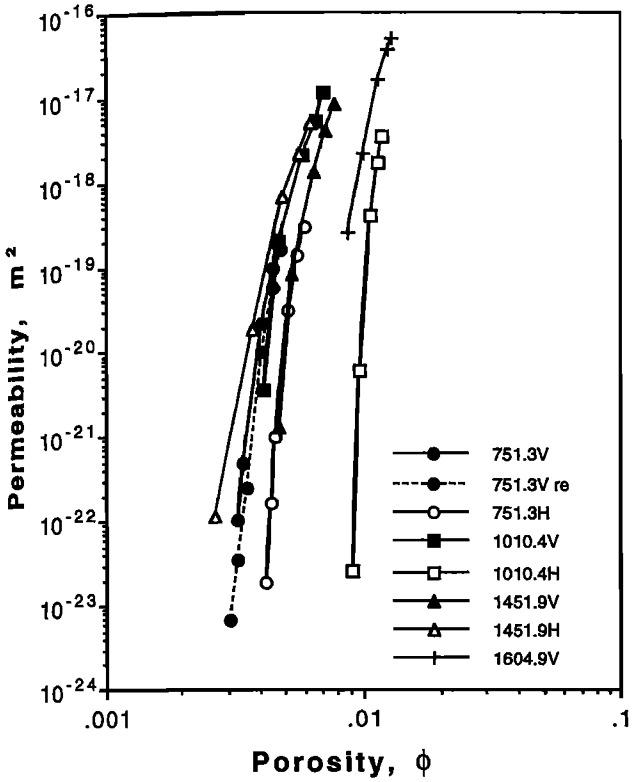

Figure 4. Permeability and porosity power law relationship, slope α is the porosity sensitivity exponent in (2).

data, Figure 3, obtained by measuring fluid discharge from the cores with applied pressure should be comparable to the scalar "volumetric crack strain" calculated for the Illinois UPH 3 core samples using DSA [*Carlson and Wang*, 1986; *Kowallis and Wang*, 1983]. Thus the porosity data provide an opportunity to compare the two methods of strain measurement on samples from the same depths in the drillhole.

With the DSA technique, strain gauges are secured on a rock cube and a cube of fused silica prepared in an identical manner (crack-free standard) in three orthogonal directions. Strain components are determined as a function of pressure. By subtracting the strain of the fused silica from that of the rock (termed "differential strain"), errors due to instrument drift and temperature changes are largely canceled out. Crack strain, due to the closure of all cracks below a given pressure P, is found as the zero-pressure intercept of a tangent to the differential strain curve at pressure P. The resulting crack strains are numerically differentiated to yield crack spectra.

A comparison of our volume measurements with data from *Kowallis and Wang* [1983] (Figure 5) shows considerably more volume strain in our samples relative to the DSA technique. There are three factors which may contribute to this difference. The first is related to the equilibrium time between pressure increases. In our experiments, variable periods of an hour to days (depending on the pressure range) were allowed for fluid equilibration between pressure increases because the decreasing permeability with applied pressure of the saturated samples makes it more difficult to expel fluids at higher pressures. Even so, it is difficult to determine when equilibration has been reached and the full amount of porosity change has been measured. The time constant for the DSA samples is considerably shorter because there are no pore liquids to expel. Accordingly, *Kowallis and Wang* [1983] find that 8-10 min between successive pressure increases is sufficient to allow heat generated by compressing the confining fluid to dissipate. However, if some residual time-dependent deformation persisted in either method beyond the chosen equilibration period, the total strain could be underestimated. Second, fluid-assisted cracking mechanisms may operate in the saturated samples (as well as downhole) that would be sluggish or fully absent in the dry DSA samples, thereby increasing the porosity of our cores more than the dry DSA cores. Finally, our samples were tested many years later than Kowallis and Wang's. It is possible that time-dependent strain relaxation acted over this period to increase the initial crack porosity of the rock. These three factors cannot be easily distinguished when comparing our strain results with the DSA data.

In spite of the strain differences observed in Figure 5, an important point to note is that the trend of both data sets with depth are strikingly similar. Volumetric crack porosity (Figure 6) at 100 MPa for our four sample depths shows a systematic excess strain compared to the DSA data. Crack porosity increases with depth, suggesting that most of the cracks are caused by stress relief microfracturing, consistent with permeability, petrographic, and borehole studies [*Kowallis et al.*, 1987; *Haimson and Doe*, 1983]. In addition, the crack distribution spectra ζ indicate that most of these cracks were formed at effective pressures of 20 MPa or less, comparable to the in situ effective pressures of the cores (Table 1).

We conclude that the two strain measurement techniques may not be completely comparable because of the crack-enhancing effects of pore fluids present in our samples, even if time-dependent factors were not an issue. However, both data sets provide further evidence that the physical properties of the Illinois UPH 3 core samples were dominated by microfractures.

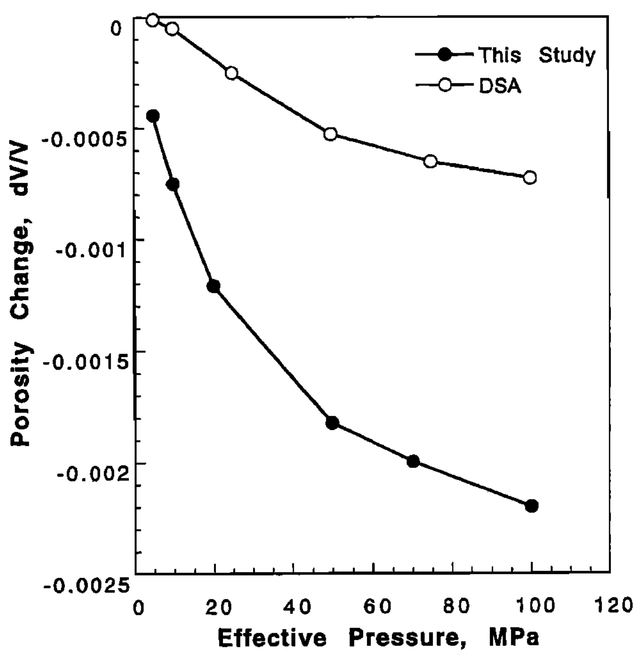

Figure 5. Comparison of porosity change (this study) and volumetric crack strain (from Differential Strain Analysis data, *Kowallis and Wang* [1983]) as a function of effective pressure for the 751.3H-m sample.

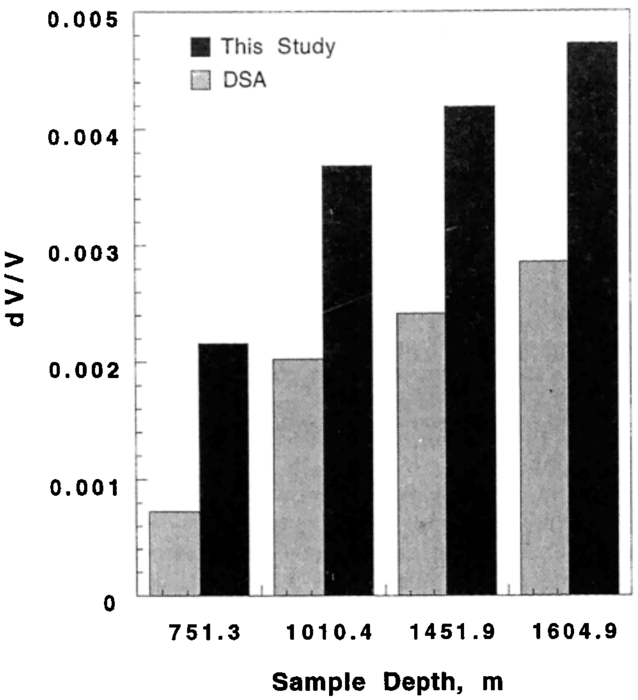

Figure 6. Comparison of volumetric crack porosity at 100 MPa, vertical and horizontal samples (this study) averaged. Differential Strain Analysis data from *Carlson and Wang* [1986].

Stress Relief Fractures

Both the permeability and porosity/DSA measurements discussed above show that stress relief cracking had a major effect on the Illinois UPH 3 core samples. One might ask, are all core samples affected by stress relief and thermal cracking, and if so, what usefulness do laboratory measurements have? A comparison with other deep core samples from our previous studies shows that these are difficult questions to answer. *Morrow and Byerlee* [1992] demonstrated that out of 67 granite, granodiorite, monzogranite and tonalite samples from the 3.5-km drillhole at Cajon Pass, California, about one fifth of the permeability measurements fell 1 to 3 orders of magnitude above the trend of decreasing permeability with depth which presumably represented in situ characteristics. We interpret the off-trend, high-permeability measurements as being due to stress relief and thermal cracking, often clearly visible in hand specimen as well as in thin section. All of these samples were from below the 2000-m level in the drillhole. Fine-grained metabasalt core samples from two depths in the 7-km KTB drillhole in Germany [*Morrow et al.*, 1994a] generally showed lower permeability at the lower depth. Although fracturing is evident at all depths of the KTB cores [*Kern et al.*, 1991; *Siegesmund et al.*, 1993], surface crack porosity values are exceedingly small. Granites and amphibolites from the 12 km Kola superdeep well in Russia [*Morrow et al.*, 1994a] were similar to the Illinois UPH 3 samples with permeability increasing with depth, whereas basalts from the Kola drillhole and also the Con Mine, Yellowknife district, Canada [*Morrow et al.*, 1994b], sampled from 1.3 to 3.9 km, showed both trends. In all, our permeability data from deep drillholes with a variety of mineralogy show mixed results, with half the samples having a decreasing trend of permeability with core retrieval depth (unfractured) and half showing a trend of increasing permeability with core depth due to stress relief and thermal cracking.

What factors cause some samples to be more prone to fracturing than others? As shown above, mineral assemblage does not seem to be a consistent indicator, even though quartz is highly susceptible to thermal and stress relief cracking [*Nur and Simmons*, 1970] and is clearly important in many cases. Nor is sample depth a reliable indicator, considering that UPH 3 is a relatively shallow drillhole compared to the others. In particular, fracture zones may locally relax the in situ stress field, making a depth and stress relief fracture correlation unreliable. *Carlson and Wang* [1986] found that cores recovered from the highly fractured 1155-1320-m interval had lower porosities than the sparsely fractured 930-1155-m interval. Corresponding in situ stresses in the fractured interval were relatively low compared to the less-fractured interval [*Haimson and Doe*, 1983]. Thus in situ fracture zones may account for some of the cases in which permeability decreased with core retrieval depth. The effect of fracture zones on microcrack porosity and in situ stress has been recognized in many other drillholes [*Meglis et al.*, 1991; *Martin and Chandler*, 1993; *Haimson and Lee*, 1984].

Grain size, thermal gradient, and stress state are other possibilities to consider, but each in turn fails to completely and independently explain our diverse permeability-depth results. For instance, the grain size of the UPH 3 samples is large (up to 15 mm), hence the grain boundaries are less tortuous and are longer compared to many other drillhole cores studied. However, permeability values of the coarse-grained UPH 751-m samples are comparable to the fine-grained, intact granodiorites from Cajon Pass, California [*Morrow and Lockner*, 1994], most of which are not extensively cracked. In contrast, the permeability of the deeper Illinois samples are similar to fine-grained granites from the Kola superdeep drillhole [*Morrow and Lockner*, 1994], which do show evidence of stress relief cracking.

Temperature gradients in the earth vary widely depending on tectonic setting, mineral conductivity, and radiogenic element content of the rocks. Does the large difference between in situ and surface temperatures combined with thermal expansion properties of constituent minerals cause cracking in the UPH 3 cores? *Carlson and Wang* [1986] report very low crack porosities for two samples from the 1179 m depth (near the intersection of three large fractures) compared to other nearby samples. If thermal cracking was the principal cause of fractures in the cores, then these two samples would not have unusually low porosities. A comparison with other drillholes also shows that temperature variation is probably not the principal cause of cracking. The geotherm for the UPH 3 drillhole was around 23°C/km [*Rahman and Roy*, 1981], within the typical range of 20-25°C/km for stable cratons. In contrast, the geotherm for the less fracture-prone granite rocks of the Cajon Pass drillhole was around 35°C/km [*Sass et al.*, 1992], high even considering its tectonic setting. Mafic rocks, with their lower mineral conductivities, often exhibit a higher geotherm than quartz-rich rocks, as for example with the KTB drillhole (30°C/km, *Jobmann and Clauser* [1994]). Here again, cracking was less pronounced than the UPH 3 cores, although mineral properties are a factor as well as temperature in this case.

Finally, are the UPH 3 samples under a greater deviatoric stress at depth than the less fracture-prone rocks from our previous studies? Crack mismatch resulting from the difference between the triaxial in situ stress state and the hydrostatic stress state of the laboratory measurements may explain why stress relief cracks do not always reclose completely under pressure. *Durham and Bonner* [1994] show that even slight crack offsets can cause substantial increases in hydraulic conductivity over wide pressure ranges. Does the stress difference between in situ and laboratory states also explain why the UPH 3 core samples are so highly fractured to begin with? A comparison of deviatoric and average effective stresses (Figure 7) for several deep drillholes on samples for which we have both laboratory permeability and downhole stress measurements at the same depth [*Zoback and Healy*, 1992; *Zoback et al.*, 1993; *Haimson and Doe*, 1983] shows that rocks from the UPH 3 hole were actually under less deviatoric stress than other drillhole samples where stress relief cracking was not as significant. The apparent coefficient of friction in the UPH 3 drillhole, determined from a Mohr's circle construction of shear and average normal stress is below 0.4, compared to 0.5 and above for Cajon Pass and KTB downhole measurements. Therefore high in situ deviatoric stresses alone can not explain the pervasiveness of the stress relief cracks in the Illinois core samples. It is more likely that a combination of the above mentioned factors, including others such as differences in elastic compliance of adjoining grains, mineral fabric orientation relative to the stress field, and the effects of hydrothermal alteration contribute to the somewhat unpredictable tendency for stress relief fracturing in the various drillholes. Clearly, this is a topic for further study and points out the importance of identifying induced fractures in core samples, particularly for the purpose of inferring in situ permeabilities and porosities from laboratory measurements.

Sensitivity Parameters γ and α

One of the purposes of this paper was to determine sensitivity parameters γ and α, compare them to values for typical surface-derived rocks, and then consider what effect these parameters have on models of midcrustal processes such as compaction and excess fluid pressure generation, key factors in the broader discussion of the strength of fault zones [e.g., *Walder and Nur*, 1984; *Rice*, 1992]. Because it is not clear what effect stress-relief fractures have on these permeability and porosity sensitivity coefficients, it would not be relevant to use the UPH 3 data in a modeling discussion. However, one of our most puzzling observations is that values of γ and α for stress relief fractured rocks tend to be in the same general range as drillhole cores that are less prone to cracking or entirely crack-free [*Morrow and Lockner*, 1994; *David et al.*, 1994]. For this reason, it may be worth a slight digression to consider what values of γ and α are more favorable to the generation of excess fluid pressure generation.

Walder and Nur [1984], in their model of fluid pressure response to porosity reduction, assume a power law permeability-porosity relationship similar to (3), with a porosity sensitivity coefficient of $\alpha = 2$ as a conservative estimate of compaction behavior. For this fixed α, they find that lithostatic fluid pressures would not be attained in a typical crustal layer unless the compaction rate of the layer was high (5×10^{-16}/s) and permeability was low (5×10^{-20} m^2). *David et al.* [1994], in an expansion of this model, show that while $\alpha = 2$ is a reasonable value for mechanical compaction of unconsolidated materials, much higher values of α are more typical of tectonic settings where mechanical and chemical processes are combined. In this case, lithostatic pore pressures are easily generated and maintained when $\alpha \geq 10$. Because high α values are not uncommon even in fracture-free rocks [*David et al.*, 1994], it implies that the physical characteristics of deep drillhole samples are more favorable to the generation of excess fluid pressures in the crust than values derived from typical surface samples, which generally have a lower α and higher permeability.

In *Rice*'s [1992] fault model, near-lithostatic fluid pressure is supplied by a continuous influx of fluid from the ductile root of the fault zone. The model requires that the excess fluid pressure in the fault zone be maintained by low-permeability material, where permeability decreases with effective normal stress as in (1). Thus higher fluid pressures are favored by more pressure-sensitive (high γ) rocks. *Rice* [1992] calculated fluid flux, fluid pressure, effective stress, and other parameters assuming $\gamma = 0.2$ MPa^{-1}, a high value compared to laboratory measurements on many different rock types. However, the model is applicable for a wide range of parameter values, particularly considering the uncertainty in the critical factor of fluid upflow from depth.

Conclusions

Permeability and Porosity

We have shown earlier [*Morrow and Lockner*, 1994] that because of weathering and other near-surface processes, surface-derived samples tend to give permeabilities that are higher than their deep in situ counterparts. We also found [*Lockner et al.*, 1991; *Morrow et al.*, 1994a] that low effective pressure determinations of deep borehole core samples gave high estimates of in situ permeabilities due to the introduction of stress relief and thermal cracks during the retrieval process. Permeability measurements of the Illinois UPH 3 granites are consistent with these previous findings; permeabilities were low compared to surface granites and dropped rapidly with applied pressure, resulting in high pressure sensitivity and porosity sensitivity coefficients γ and α. In spite of the low-permeability values of the Illinois UPH 3 granites, both permeability and porosity increased with sample depth at all pressures studied, indicating that stress relief and thermal crack damage induced during the core retrieval process dominated the physical properties of these granites even at pressures greater than the estimated in situ values.

We suggested in earlier studies [*Lockner et al.*, 1991; *Morrow et al.*, 1994a] that it may be possible to separate the effects of in situ and stress relief cracks. We postulated that at low effective pressures, stress relief cracks would dominate the matrix permeability measured in the laboratory, but that above in situ effective pressures, stress relief cracks would be closed and permeability and porosity would be controlled by the natural crack population. By measuring permeability over a wide range of pressures, a pressure region could be identified in

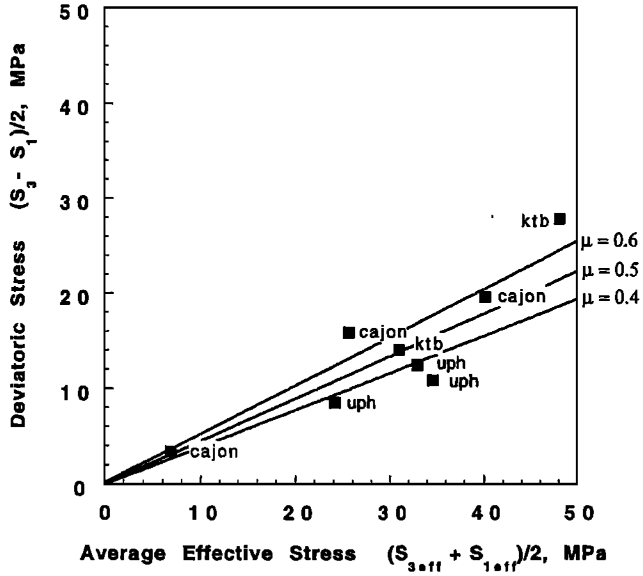

Figure 7. Deviatoric and average effective stress from downhole measurements at the Cajon Pass (California), KTB (Germany) and UPH 3 (Illinois) drillholes. In situ stress data [*Zoback and Healy*, 1992; *Zoback et al.*, 1993; *Haimson and Doe*, 1983] chosen at depths from which we also have laboratory permeability measurements on core samples. Lines show coefficient of friction envelopes for μ = 0.4, 0.5, and 0.6.

which in situ crack properties dominated. This method was successful for granites from the 11.4 to 12-km level of the Kola Superdeep well in Russia [*Lockner et al.*, 1991], but not for cores from the full range of depths and rock types in the Kola well, nor the KTB drillhole in Germany [*Morrow et al.*, 1994a]. Our present results indicate that this approach is also not suitable for the UPH 3 granites. Thus it may be that even if appropriate in situ pressure conditions could be reproduced in the laboratory, permeability measurements may still be dominated by drilling- and retrieval-induced cracks because microcrack growth is inherently an irreversible process. Reproducing the average in situ stress state does not guarantee generation of the grain-to-grain tractions necessary for microcrack closure. Therefore the measurements provide at best an upper bound on the in situ matrix permeability. This conclusion has important implications for the inference of in situ permeabilities from laboratory measurements.

It should be noted that even if laboratory measurements are treated as upper bounds they suggest that in situ matrix permeabilities for crystalline rocks are extremely low and may, in some cases, approach zero [*Wang and Simmons*, 1978]. This is consistent with recent observations of rapid permeability loss in samples maintained at elevated temperatures and fluid pressures [*Moore et al.*, 1994] and also supports the notion that fluid flow in the mid to lower crust will in general be dominated by flow in fractures. Dewatering and densification processes will raise pore pressures, lowering effective pressure, to allow pore fluids to escape at a rate that is controlled by the rate of fluid generation or the compaction rate. However, once the source of pore fluids is exhausted, matrix permeability is likely to drop rapidly to very low values. Low absolute permeability, together with high porosity and pressure sensitivity of permeability of deeply buried rocks, favors the development of high fluid pressures, as discussed by *Walder and Nur* [1984], *Rice* [1992], *Sleep and Blanpied* [1994], and *Lockner and Byerlee* [1995].

Volumetric Strain

Volumetric strain measurements conducted using differential strain analysis [*Kowallis and Wang*, 1983; *Carlson and Wang*, 1986] were consistently lower than porosity change data based on volume discharge due to crack closure (this study), probably because of the crack-enhancing effects of fluids present in our samples that are absent in the dry DSA cores, as well as possible time-dependent deformation effects. This observation has important consequences if volumetric strain/porosity measurements of deep core samples are used in models of stress, fluid circulation, and excess fluid pressure generation in the midcrust. Both measurement methods show a systematic trend of increased porosity with core retrieval depth, indicating that stress relief fractures dominate the porosity of the core samples, consistent with permeability and petrographic data.

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