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# Pore Pressure Diffusion and the Mechanism of Reservoir-Induced Seismicity

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*Abstract* – The study of reservoir-induced seismicity offers a controlled setting to understand the physics of the earthquake process. Data from detailed investigations at reservoirs in South Carolina suggested that the mechanism of transmission of stress to hypocentral locations is by a process of diffusion of pore pressure (Pp). These results were compared with available worldwide data. The 'seismic' hydraulic diffusivity, $\alpha_s$, was estimated from various seismological observations, and was found to be a good estimate of the material hydraulic diffusivity, $\alpha$. Application of these results to a dedicated experiment to understand RIS at Monticello Reservoir, S.C., suggested that the diffusing Pp front plays a dual role in the triggering of seismicity. The spatial and temporal pattern of RIS can be explained by the mechanical effect of diffusion of Pp with a characteristic hydraulic diffusivity within an order of magnitude of $5 \times 10^4$ cm$^2$/s, corresponding to permeability values in the millidarcy range. The triggering of seismicity is due to the combined mechanical effect of Pp in reducing the strength and, possibly, the chemical effect in reducing the coefficient of friction between the clays in the pre-existing fractures and the rocks that enclose these fractures.

**Key words:** Mechanism, reservoir, induced seismicity.

## *1. Introduction*

The incidence of reservoir-induced seismicity (RIS) is usually confined in both space and time, and it is now being routinely monitored on dense local networks. The seismic data so collected provide a controlled setting to study the mechanism of RIS in particular and the physics of the earthquake process in general. Until recently it was thought that RIS was triggered by the loading of a reservoir and/or by the effect of pore pressure (Pp) in lowering the strength of rocks at hypocentral depths. As case histories accumulated, Pp was considered to be the primary factor. It was assumed that the rocks are close to failure and small perturbations in the *in situ* stress field due to Pp changes trigger the observed RIS. It has been further assumed that the coefficient of friction $\mu$ is constant and lies between 0.6 and 0.85.

Pore water can play a two-fold role in the earthquake process, the first, a mechanical effect as pore pressure, and second, a chemical effect as stress-aided corrosion. There is evidence to suggest that pore water or pore pressure diffuses along pre-existing fractures, bedding planes, etc. (WITHERSPOON and GALE, 1977); or it can be associated with new crack propagation through stress corrosion (ANDERSON and GREW, 1977). We suggest that the mechanical effects of pore pressure control the spatial and temporal

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pattern of RIS, whereas the actual onset of seismicity may be influenced by the chemical effect of water in reducing the coefficient of friction in clays (filling pre-existing fractures). We shall be primarily concerned with the first effect, although we recognize that stress corrosion as it applies to hydration of clay minerals under increasing pore pressure is an important factor in RIS.

In the last ten years, we have monitored RIS at four locations in South Carolina, including, since its inception in October 1975 and December 1977, Lake Jocassee and Monticello Reservoir, respectively. The observation of a linear growth rate of epicentral area following impoundment at Lake Jocassee and other locations of RIS, suggested that the mechanism of transmission of the pressure front to hypocentral locations was by a process of pore pressure diffusion. We do not have any direct observation of a propagating pressure front. However, our conclusions are based on the inference that the onset and migration of RIS is associated with such a pressure front.

By monitoring the epicentral growth, and assuming it to be associated with the diffusion of pore pressure, we calculate a parameter, which we label 'seismic' hydraulic diffusivity, $\alpha_s$. We then suggest that by making some simple yet reasonable assumptions, $\alpha_s$ can be estimated from a wide variety of RIS data.

We have examined the available published literature on RIS and wherever possible estimated the 'seismic' hydraulic diffusivity $\alpha_s$. Considering the uncertainties in various published data the surprising result from observation of 22 case histories was the relatively narrow range of values estimated for $\alpha_s$. Over 30 estimates of $\alpha_s$ were within one order of magnitude of $5 \times 10^4$ cm$^2$/s. We show that these estimates of 'seismic' hydraulic diffusivity ($\alpha_s$) are within an order of magnitude of the material hydraulic diffusivity ($\alpha$). The parameter $\alpha$ is not commonly used in hydrogeological literature. However, it is directly related to a more commonly used hydraulic property, permeability, $k$ by the equation $\alpha = k/\mu\phi\beta$ where $\mu$, $\phi$ and $\beta$ are fluid viscosity, porosity of the rocks and compressibility of the fluid respectively. If $\mu$, $\phi$ and $\beta$ are known, then $k$ can be estimated from $\alpha$. Thus we have a simple, albeit an indirect method of estimating the *in situ* 'crustal' permeability. Even allowing for uncertainties in the estimation of $\alpha_s$, and $\phi$, we obtained a narrow range of 'crustal' permeabilities associated with RIS. These lie in the millidarcy range, in agreement with those inferred at this scale by BRACE (1984).

These ideas of diffusion of pore pressure, narrow range of hydraulic diffusivity, and inferred permeabilities were tested in an elaborate experiment at Monticello Reservoir, a site of ongoing RIS. Besides an in depth investigation of seismological, hydrological, and geological factors, two deep boreholes were drilled in the active hypocentral regions. Borehole investigations included the measurement of *in situ* stress, orientation and density of fractures, pore pressure, etc. (ZOBACK and HICKMAN, 1982; SEEBURGER and ZOBACK, 1982; MOOS and ZOBACK, 1983). A comparison with detailed fault plane solutions suggested that the seismicity was occurring on pre-existing fractures (TALWANI, 1981b). Thus for earthquakes to occur, the frictional resistance on a fracture has to be overcome. BYERLEE's (1978) laboratory data indicated that at pressures corresponding to shallow crustal depths, $\tau = 0.85\ \sigma_n$, where $\tau$ and $\sigma_n$ are the shearing and normal stress at which frictional resistance is overcome on a fracture. The significance of this result was that the coefficient of friction $\mu$, was constant (=0.85) and independent of rock type or surface conditions. For higher pressures $\mu$ was found to be 0.6. This result was labelled Byerlee's law by BRACE and KOHLSTEDT (1980). MEISSNER and STREHLAU (1982) noted that Byerlee's data were based on relatively short time scales, low temperatures and obtained with small samples, and cautioned against extrapolation to crustal dimensions. In his paper, BYERLEE (1978) did note that the coefficient of friction was drastically reduced in the presence of clays.

In our analyses of the *in situ* stress data at Monticello Reservoir, we were led to the conclusion that to explain the observed RIS there, BYERLEE's law (1978) for friction between rock and rock may not be applicable and that the seismicity is probably associated with the effect of pore pressure on clays filling the pre-existing fractures.

These results have possible applications in regions where the growth of aftershock zones or the migration of seismicity is relatable to fluid flow.

## 2. *Background about theories of RIS*

GOUGH and GOUGH (1970) suggested that the observed increase in seismicity at Lake Kariba following its impoundment could have been triggered by loading or due to the effect of increased fluid pressure. They preferred the former mechanism. However, in the last decade there is a considerable body of data that very persuasively suggests that changes in the pore pressure are the main cause of the observed seismicity. HUBBERT and RUBEY (1959) showed that an increase in subsurface pore pressure reduces the effective normal stress on fault planes. These results were used by SNOW (1972) who showed that filling of a reservoir built on an infinite halfspace was conducive to RIS in strike slip and normal fault environments and inhibitive in a thrust fault environment. In all cases, however, the effect of increase in pore pressure was to decrease the effective strength of rocks leading them to failure. However, if the logic of Snow's arguments was followed, we should not expect RIS in areas of thrust faulting except by unloading of the reservoir. The work of HUBBERT and RUBEY (1959) and SNOW (1972) has been incorporated by GUPTA and RASTOGI (1976) in their book, *Dams and Earthquakes*, and will not be pursued further here. Snow's arguments were incorporated in a review paper by SIMPSON (1976) and by WITHERS (1977) and WITHERS and NYLAND (1978), who used consolidation theory to calculate the time history of stress below a freshly impounded reservoir. Making many simplifying assumptions, they concluded, as did SNOW (1972), that RIS was most likely in strike slip and normal fault regimes. In another theoretical study, BELL and

NUR (1978) showed that anisotropy in rock permeability and relative changes in ground-water played an important role in RIS.

The role of pore pressure diffusion as a mechanism of stress transmission was recognized over a decade ago (e.g. NUR and BOOKER, 1972; SCHOLZ *et al.*, 1973) and was used to explain the onset time of seismicity following injection in wells at Matsushiro (OHTAKE, 1974), Rangeley (RALEIGH *et al.*, 1976) and Dale (FLETCHER and SYKES, 1977). HOWELLS (1974) suggested that it could explain RIS, and TALWANI (1976) used the concept to account for the observed time lag between lake level rise and onset of seismicity at Clark Hill reservoir. Subsequently, TALWANI and RASTOGI (1978) and TALWANI (1981a) evaluated the then available worldwide data and suggested that pore pressure diffusion is the preferred mechanism for RIS. This conclusion was based on the observation of a linear growth of epicentral area with time at Lake Jocassee (TALWANI *et al.*, 1976). If we assume that an increase in epicentral area is directly caused by diffusion of pore pressure to hypocentral locations (with little change in depth), then 'seismic' hydraulic diffusivity can be estimated from the epicentral growth rate. It can also be estimated from other seismological parameters such as the time lag between the filling (or draining) of a reservoir and the onset of seismicity, and the time lag between the start (or cessation) of injection in a well and the onset (or cessation) of seismicity (OHTAKE, 1974; FLETCHER and SYKES, 1977; TALWANI and RASTOGI, 1978; TALWANI, 1981a). This method of calculation of 'seismic' hydraulic diffusivity is subject to various uncertainties. These include uncertainties in the location of 'ground zero' from where the pore pressure front originates. Assuming the transmission of pore pressure along fractures, the 'ground zero' would be on a line where those fracture(s) intersect the bottom of the reservoir. With the available data from the case histories this is seldom determinable. Therefore we plot the data as available, and note that the uncertainty in 'ground zero' location is proportional to the dimensions of the reservoir. The hypocentral locations for the earlier cases of RIS (accurate to a few km) are not known with the same accuracy as those with modern microearthquake networks (accurate to within a km). Another uncertainty lies in the difficulty in choosing the time of water level rise (or fall), (the onset of the pore pressure front). In calculating $\alpha_s$ we further assume that the pore pressure follows a straight line path.

In our calculations of $\alpha_s$, we have estimated the uncertainties in hypocentral locations and in the calculation of the time lags. These are represented by error bars in the parameters $L$ and $t$ (see next section). If the pore pressure front does not follow a straight line path, but one twice as long, it will lead to an error of 0.4 of an order of magnitude in the estimate of $\alpha_s$.

Owing to the various uncertainties discussed above, and the simplifying assumptions, this method is obviously an approximation, however, its main justification lies in the coherent results obtained. We show below that the hydraulic diffusivity obtained from seismological parameters ($\alpha_s$) is a meaningful estimate of the material hydraulic diffusivity. This method of estimating $\alpha_s$ is now being increasingly used (e.g. KEITH *et al.*, 1982; ZOBACK and HICKMAN, 1982).

### 3. *Calculation of* $\alpha_s$

After the first felt event in October 1975, the seismicity at Lake Jocassee has been monitored continually since November 1975. After a few felt events that occurred between 8 and 10 November 1975, the seismicity increased, with a $M_L$ 3.2 event occurring on 25 November 1975. The epicentral region, which was initially in the vicinity of the dam, increased from November 1975 to February 1976, decreased in March and then increased again in April and May 1976, when it was the largest (Fig. 1). Subsequent activity in the next seven years has been confined, for the most part, to lie within the epicentral envelope defined by the seismicity in May 1976.

In Fig. 2, we note that the epicentral area growth for the period November 1975 to February 1976, is linear with time. The growth rate is about 40 km$^2$/100 days (or about $4.6 \times 10^4$ cm$^2$/s). The growth between March and May 1976 is again linear, the growth rate is now about $5.0 \times 10^4$ cm$^2$/s.

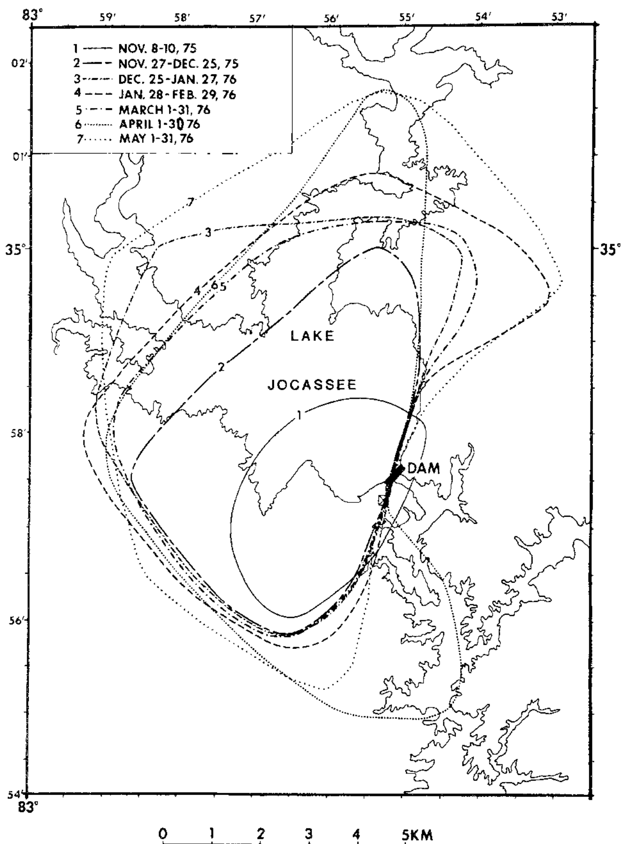


Figure 1

The epicentral growth of seismicity at Lake Jocassee in the first six months after its inception in November, 1975. Note the decrease in the epicentral area in March 1976 (curve 5) after the growth in the previous months.

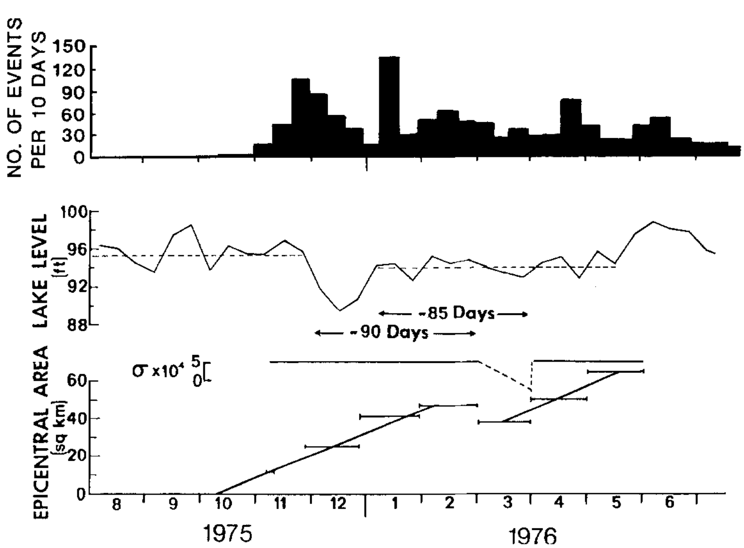


Figure 2
Epicentral growth at Lake Jocassee as a function of time compared with the lake level and the number of events per 10 day periods. The growth rate corresponds to a hydraulic diffusivity of $5 \times 10^4$ cm$^2$/s. The decrease in epicentral area in 3/76 is related to lake level decrease in 12/75 (top).

In Fig. 2 the epicentral growth was compared to the lake level (the ten day average of the 8 a.m. readings). Water level was suddenly lowered after the $M_L$ 3.2 event on 25 November 1975, and raised again in January 1976. There appears to be a corroborative decrease in growth of the epicentral area about three months after the decrease in the lake level. This decrease in March 1976 was located on the periphery of the epicentral envelope, about 5 to 6 km away from the dam.

The linear growth rate of the epicentral area (Fig. 2) suggested a relationship of the kind $L^2 \sim t$, or that $L^2/t = \alpha_s$, a constant we call 'seismic' hydraulic diffusivity. Such dependence is expected from the diffusion equation describing flow of some kind. Thus, the observed linear growth rate of the epicentral area lends support to the diffusion mechanism as the transmitter of pore pressure.

To check this conclusion, we can calculate the time it will take to change the pore pressure at a distance of 5 to 6 km due to a change in the lake level (e.g. 12/75) for a hydraulic diffusivity of $5 \times 10^4$ cm$^2$/s. The required time is 58 to 83 days. For a hydraulic diffusivity of $4.6 \times 10^4$ cm$^2$/s, the corresponding delay is 64 to 93 days. These time periods agree well with the observed delays (Fig. 2).

There are very few examples where epicentral growth as a function of time due to RIS is well documented. However, there are several examples where the hypocentral distance from the location of impoundment, $L$, is known, and the time lag between impoundment, and the onset of seismicity, $t$, is also known. In such cases, assuming that the earthquakes have been caused by pore pressure diffusion, $\alpha_s$, the 'seismic' hydraulic diffusivity can be easily estimated ($=L^2/t$). For Monticello and Clark Hill reservoirs in South Carolina this method gave $\alpha_s$ values of the order of $10^4$ cm$^2$(s.

Encouraged by these results, $\alpha_s$ was also estimated for cases of seismicity associated with the injection of fluids in boreholes and where the seismicity was known to have spread linearly along fault planes. The results of various methods used are presented below.

## 4. *Estimation of $\alpha_s$ from epicentral growth*

Besides Lake Jocassee, data on the initial growth of epicentral area of RIS were available at Koyna, India (GUHA *et al.*, 1974). Hsingfengkiang, China WANG *et al.*, 1976) and at Oroville, California (LESTER *et al.*, 1975).

The seismicity that followed the impoundment in 1962 of Shivaji Sagar Lake by Koyna Dam in the Peninsular shield of India is among the best known examples of RIS. Before impoundment no tremors had been reported for the region. Although seismicity was first felt in 1962, no epicentral data are available before 1964. The seismicity grew in space and magnitude until two significant events occurred in 1967. These were the magnitude 5 and 6.3 events that occurred on 13 September and 10 December respectively. The epicentral growth rate and hence $\alpha_s$ observed at Koyna were not uniform. For the period from 1962 to mid-July 1967, $\alpha_s$ was about $5 \times 10^3$ cm$^2$/s. However, following the 13 September event and before the 10 December main shock, the growth rate was higher, yielding $\alpha_s = 9 \times 10^4$ cm$^2$/s $\pm 2 \times 10^4$ cm$^2$/s including uncertainties in $L$ and $t$. The main event occurred on 10 December, and the pursuant growth rate for the first three months of 1968 was even higher, $\alpha_s \sim 2 \times 10^5$ cm$^2$/s. Thus it appears that the larger two events caused changes in the surrounding rocks that resulted in an increase in $\alpha_s$. Such an increase would be associated with the formation of new fractures.

Such a noticeable change was not observed at Hsingfengkiang, China. There, after impounding the reservoir in October 1959, small shocks occurred concurrently, and the epicentral area grew and on 19 March 1962 a magnitude 6.1 earthquake occurred (WANG *et al.*, 1976). The epicentral growth for this ~2.5 year period implied $\alpha_s \sim 5 \times 10^4$ cm$^2$/s $\pm 1 \times 10^4$ cm$^2$/s. The epicentral region continued to grow and for the remaining part of the year $\alpha_s$ was $\sim 6 \times 10^4$ cm$^2$/s $\pm 1 \times 10^4$ cm$^2$/s.

The results of aftershock studies immediately following the $M_L$ 5.7 Oroville, CA earthquake of 1 August 1975 were presented by LESTER *et al.* (1975). The rapid epicentral growth rate in the first week was followed by a slower growth rate in the following months, with the corresponding values of $\alpha_s$, $5.7 \times 10^{\times}$ and $3.8 \times 10^4$ cm$^2$/s, respectively. SAVAGE *et al.* (1976) also carried out an aftershock survey in the month following the main shock. $\alpha_s$ calculated from their data also shows a similar decrease, decreasing from $5.1 \times 10^5$ cm$^2$/s to $3.2 \times 10^4$ cm$^2$/s.

In summary, the 'seismic' hydraulic diffusivities calculated from an estimate of epicentral growth rate, $\alpha_s$, range from $5 \times 10^3$ to $6 \times 10^5$ cm²/s, with most values clustering around $5 \times 10^4$ cm²/s. The larger values ($\sim 10^5$ cm²/s) were obtained immediately after the main shock and are probably associated with increased fracturing following the main shock, and the lower values ($\sim 10^4$ cm²/s) are more representative of the hypocentral region.

## 5. *Estimation of $\alpha_s$ from time lag between change in lake level and onset of seismicity*

A careful perusal of various listed cases of RIS reveals that there is always a time lag between filling (or draining) of a reservoir and the onset of seismicity. In most cases, however, adequate epicentral data are not available. If we make the assumption that the diffusion of pore pressure is the operative mechanism of transmitting pore pressure changes to hypocentral depths, it is possible to estimate $\alpha_s$ from the observed time lag. WITHERSPOON and GALE (1977) have shown that water flow in fractured rocks occurs principally through joints, faults and other planar features. Here we assume that due to the generally observed anisotropy in rocks, diffusion is also restricted to the plane of the fault or fracture. With these assumptions the rate of diffusion of the pore pressure front which is associated with seismicity can be estimated directly. The hydraulic diffusivity is obtained from the distance $L$, between the source of the pressure front (the reservoir) and the location of seismicity, and from the delay $t$, between generating this front (filling or draining the reservoir) and the onset of seismicity, by the simple relation, $\alpha_s = L^2/t$. This method is simple and generates reasonable values, as illustrated by the following examples.

Figures 3 and 4 show the location of epicenters for the period 12/77–3/78 and cumulative number of earthquakes following the impoundment of Monticello Reservoir in December 1977. Fault plane solutions indicate that the seismicity is occurring in a thrust fault environment. In such an environment the filling of a reservoir is associated with two effects. The increased load of water inhibits seismicity, whereas the delayed effect of pore pressure favors it. Thus observed seismicity is primarily due to changes in pore pressure. We note that there is a lag of about three weeks between the start of filling and the start of seismicity. After completion of filling, in early February 1978, the increasing inhibiting effect of loading stopped, and there was a marked increase in the seismicity for about three weeks, after which a steady increase was noted. This three week lag was associated with the diffusion of pore pressure to hypocentral locations at depths of 1 to 2 km. By the method outlined above, the corresponding values of $\alpha_s$ are $5.5 \times 10^3$ cm²/s $\pm$ $0.4 \times 10^3$ cm²/s and $2.2 \times 10^4$ cm²/s $\pm$ $0.2 \times 10^4$ cm²/s.

Similar estimates were made for all the reservoirs where adequate data were available. (The details are being published elsewhere.) Forty-two such values were obtained and they ranged between $5 \times 10^3$ and $5 \times 10^5$ cm²/s (Fig. 5). This result is in good agreement with the values obtained from the growth rate of epicentral area.

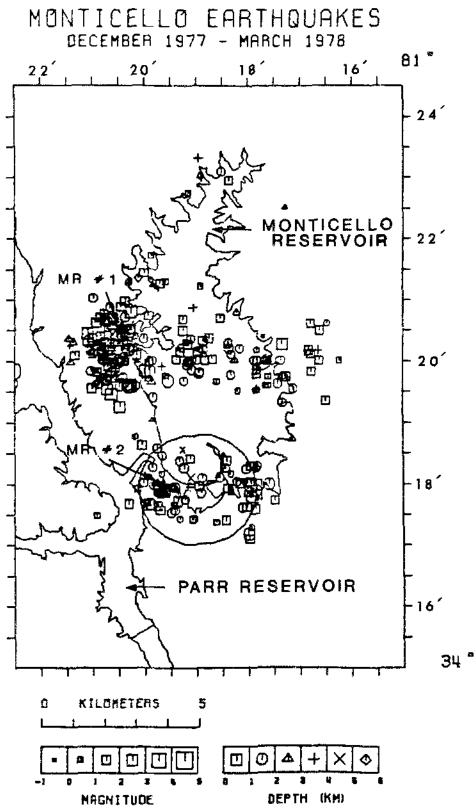


Figure 3
Spatial distribution of seismicity at Monticello Reservoir, from its inception in the last week of December 1977 through March 1978. MR nos 1 and 2 show the locations of the two 1 km deep wells.

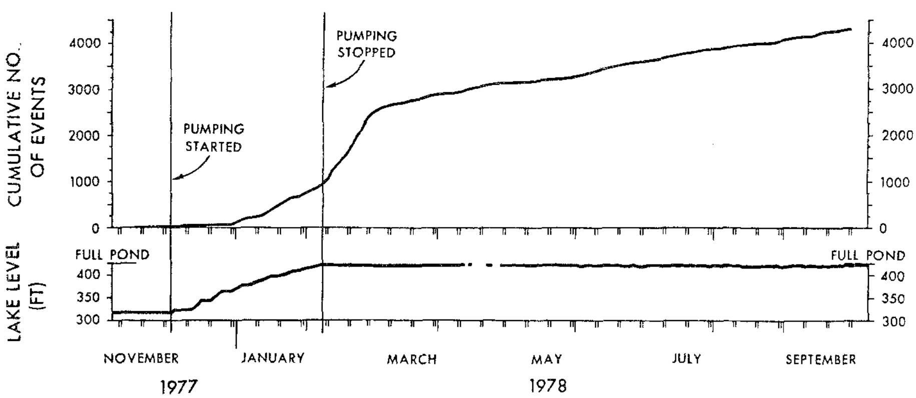


Figure 4
Cumulative seismicity at Monticello Reservoir (top) compared with the lake level. The reservoir was filled by pumping from the lower Parr Reservoir.

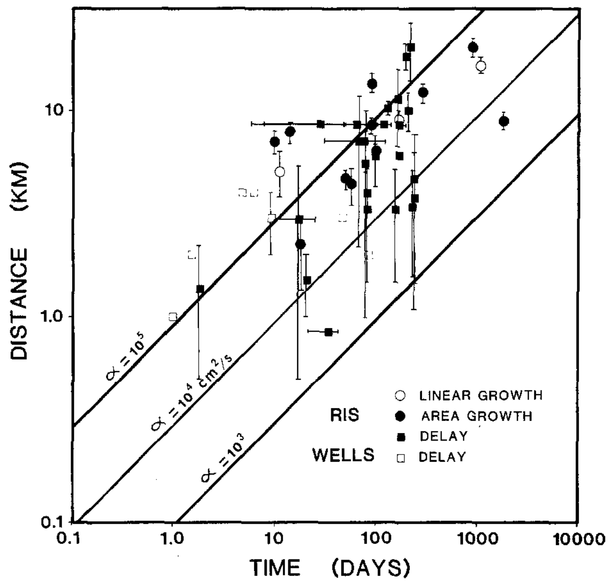


Figure 5
Hydraulic diffusivity associated with RIS calculated by different methods. Uncertainties in the calculation of time lags and hypocentral distances are represented by error bars.

### 6. *Estimate of $\alpha_s$ from linear growth of seismicity*

In some cases a more or less linear epicentral growth is observed, indicating earthquake migration along a fault. If this pattern is associated with the reservoir at one end, $\alpha_s$ can be estimated in a manner similar to that described in the previous section. Here if $L$ is the distance that the earthquake front has propagated, in time $t$, the 'seismic' hydraulic diffusivity, is as before, $\alpha_s = L^2/t$. Such an epicentral spreading was observed at Kariba, S. Rhodesia, now Zimbabwe (GUPTA and RASTOGI, 1976), and at Nurek, in Tadjikistan, Russia (SOBOLEVA and MAMADALIEV, 1976; SIMPSON and NEGMATULLAEV, 1981). The calculated values of $\alpha_s$ were $5.3 \times 10^4 \pm 0.8 \times 10^4$ and $2.9 \times 10^4 \pm 0.5 \times 10^4$ cm$^2$/s, respectively.

### 7. *$\alpha_s$ from time lag between fluid injection in deep wells and onset of seismicity*

There are many known cases of seismicity associated with the injection of fluids in deep wells (HEALY *et al.*, 1968; OHTAKE, 1974; FLETCHER and SYKES, 1977; RALEIGH *et al.*, 1976). In such cases again we make similar assumptions about the time lag and distance of seismicity from the well to hypocentral locations. In these cases, both the $t$ and $L$ are known very accurately.

For example, seismicity was found to be located about 4 km from the bottom of the 1800 m deep cased well at Matsushiro, Japan (OHTAKE, 1974). Spurts of seismicity were correlatable with sudden increases in fluid pressure and followed them by 9.3, 6.2 and 4.8 days. The seismicity was in the vicinity of the Matsushiro fault – the location of a large swarm of seismicity about 4 to 5 years earlier. The calculated values of $\alpha_s$ ($\sim 10^5$ cm$^2$/s) are high and possibly reflect the fractured state of the hypocentral region after the Matsushiro swarm.

Similar estimates of $\alpha_s$ were obtained from other cases of induced seismicity associated with the injection of fluids in a well (or between cessation of injection and associated cessation of seismicity) at Dale, N.Y. (FLETCHER and SYKES, 1977); Denver arsenal well (HEALY *et al.*, 1968); and the Rangeley, Colorado well (RALEIGH *et al.*, 1976). The calculated value of $\alpha_s$ in all cases ranged between $5 \times 10^3$ and $4 \times 10^5$ cm$^2$/s.

In summary, $\alpha_s$, the 'seismic' hydraulic diffusivity calculated from seismological data – for all cases of RIS where usable data were available, was found to be within one order of magnitude of $5 \times 10^4$ cm$^2$/s. We examine the significance of this result in the following section.

## 8. $\alpha_s$ *and the material diffusivity*

To a first approximation, for one-dimensional diffusion of pore pressure $p$,

$$(\partial^2 p/\partial z^2) = (1/\alpha)(\partial p/\partial t) \tag{1}$$

where $z$ is distance, $t$ the time and $\alpha$ is the diffusivity of the material in which diffusion occurs. This equation has a solution of the form

$$(p(z,t)/p_0) = 1 - \operatorname{erf}(z/2\sqrt{\alpha t}) \tag{2}$$

where $p_0$ is the pore pressure applied at $z = 0$, $t = 0$, and maintained for $t > 0$; and $p(z,t)$ is the pore pressure at distance $z$ after time $t$. The behavior of the solution (equation 2) is illustrated in Fig. 6 for $\alpha = 10^4$ cm$^2$/s. We label $M$ the ratio $p(z,t)/p_0$ on the left-hand side of equation (2). It is the ratio of the diffused pore pressure at depth $z$ after time $t$ to the applied pore pressure, $p_0$.

Now in the case of RIS, the activity is triggered by the perturbation in the ambient pore pressure at the hypocenter due to the diffusing pore pressure front. The seismicity occurs at some threshold, i.e., some value of $p(z,t)$; hence at a particular value of $M$. For example, if the applied pore pressure $p_0$ is 10 bars and the stress field conditions are such that seismicity is triggered at a depth of 2 km when the perturbing pore pressure is 5 bars, then, $p(2,t) = 5$ bars, $M = 0.5$ and from Fig. 6, $t = 50$ days.

However, *a priori*, we do not know the particular value of $M$ at which an earthquake is triggered, so that the diffusivity that will be estimated from seismic data,

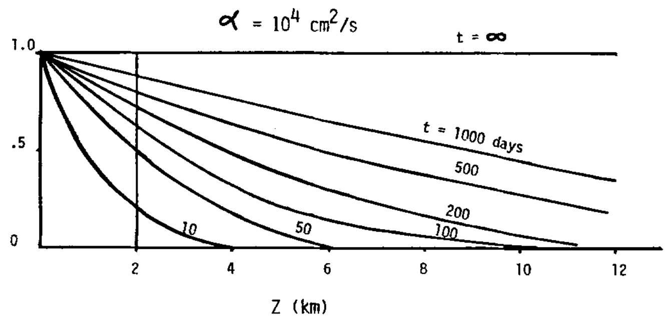


Figure 6
Variation of ratio $M$ for hydraulic diffusivity of $10^4$ cm²/s plotted as a function of time and distance.

$\alpha_s$ ($L = 2$ km, $t = 50$ days, $\alpha_s = 9.3 \times 10^3$ cm²/s) will not equal the material diffusivity, $\alpha$ ($= 10^4$ cm²/s). Suppose that the earthquake had been triggered at $M = 0.2$ or 0.61, at the same depth (2 km), the corresponding times are 10 and 100 days, and estimated values of $\alpha_s$ are $4.6 \times 10^4$ and $4.6 \times 10^3$ cm²/s respectively.

Clearly any *calculated* value of hydraulic activity from seismic data will depend on $M$. Recall, $M = p(z, t)/p_0$ is some fraction of the applied pore pressure perturbation $p_0$, and the value of $M$ at which seismicity is triggered is dependent on the ambient stress conditions. So the question arises, is the hydraulic diffusivity calculated from seismic data at all meaningful? To answer that, rewrite equation (2).

$$M = p(z, t)/p_0 = 1 - \operatorname{erf}(z/2\sqrt{\alpha t})$$
$$= 1 - \operatorname{erf}(z^2/4\alpha t)^{1/2}$$

Now, $z^2/t$ is the 'seismic' hydraulic diffusivity, $\alpha_s$, and $\alpha$ is the material diffusivity; then

$$M = 1 - \operatorname{erf}(\alpha_s/4\alpha)^{1/2} \qquad \text{or} \qquad \alpha_s/\alpha = 4[\operatorname{erf}^{-1}(1 - M)]^2 \tag{3}$$

Figure 7 shows the variation of $\alpha_s/\alpha$, the ratio of the 'seismic' hydraulic diffusivity to the 'true' material diffusivity, for different values of $M$. We note that for extreme values of $M$ (0.9 and 0.03), $\alpha_s$ is $0.1\alpha$ and $10\alpha$ respectively. Thus the estimated 'seismic' hydraulic diffusivity $\alpha_s$, is within an order of magnitude of the 'true' material diffusivity $\alpha$.

## 9. *Permeability and hydraulic diffusivity*

The hydraulic diffusivity, $\alpha$, is related to the permeability of a medium, $k$, by the relation (BODVARSSON, 1970; TALWANI, 1981a)

$$\alpha = k/\{\mu[\phi\beta_f + (1 - \phi)\beta_r]\} \tag{4}$$

where $\mu$, $\phi$, $\beta_f$ and $\beta_r$ are fluid viscosity, porosity of the rocks, and the compressibilities of fluid and rock respectively. If diffusion is in the fluid filled fractures, equation (4)

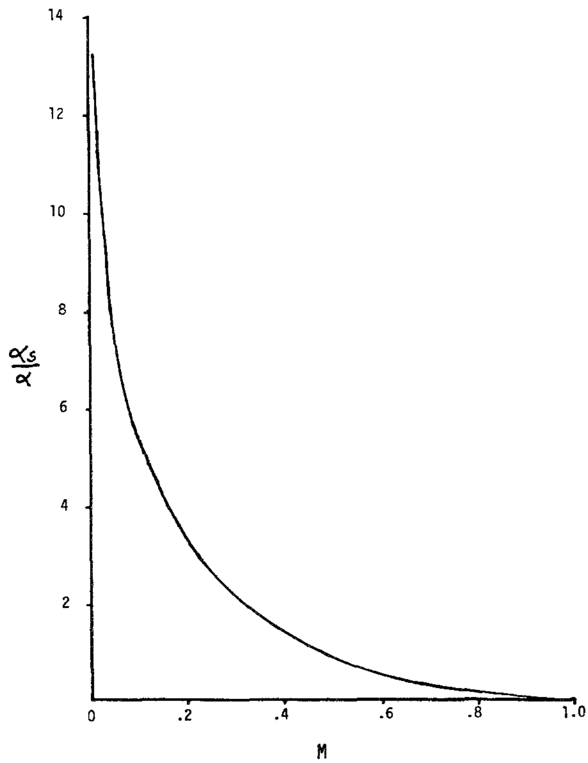


Figure 7
Relation between 'seismic' and 'material' diffusivity as a function of $M$.

can be simplified (see e.g., SCHOLZ *et al.*, 1973 or BRACE, 1980), to $\alpha = k/(\mu\phi\beta_f)$. If we assume typical values of these parameters,

$\mu$ = viscosity of water = $10^{-2}$ poise = $10^{-8}$ bar sec.

$\phi$ = porosity of fractured rock = $10^{-2}$ to $10^{-3}$, say $3 \times 10^{-3}$ (BRACE, 1980, 1984)

$\beta_f$ = effective compressibility of fluid = $3 \times 10^{-5}$ bar$^{-1}$

then, $\alpha \sim 10^4$ to $10^5$ cm$^2$/s implies a permeability value of $10^{-11}$ to $10^{-10}$ cm$^2$ or 1–10 millidarcies. If we allow for two orders of magnitude uncertainty in $\alpha_s$ due to uncertainties in parameters or our incomplete knowledge of the stress field conditions at hypocentral location, the range of $\alpha$ becomes $10^3$–$10^6$ cm$^2$/s (clearly bracketing all our data – Fig. 5). Further allowing for an order of magnitude uncertainty in our estimate of the porosity of the fractured rocks, these values imply a maximal range of $10^{-13}$ to $10^{-9}$ cm$^2$ (0.01 to 100 md). Our data (Fig. 5) however suggest a narrower range, $10^{-12}$ to $10^{-10}$ cm$^2$ (0.1 to 10 md) as being more representative of the true range of permeabilities in crustal rocks. These values are in general agreement with those given by BRACE (1984), who noted that for crustal phenomena, estimates made *in situ* on a kilometer scale may be the most meaningful. These values are of the same order as those for the more permeable intervals in boreholes, where the values range from $10^{-9}$ to $10^{-15}$ cm$^2$ (BRACE, 1984).

Thus, although the permeability of rocks encountered in nature varies from the nanodarcy to the darcy range ($10^{-17}$ to $10^{-8}$ cm$^2$) (BRACE, 1984), RIS is usually associated with rocks having a permeability in the millidarcy range (0.1 to 10 md or $10^{-12}$ to $10^{-10}$ cm$^2$).

## *10. Application to RIS at Monticello Reservoir*

As we have shown above, the process of pore pressure diffusion adequately accounts for the *time lag* between the impoundment of a reservoir (or lake level fluctuations) and the onset of seismicity. The working model (SNOW, 1972; SIMPSON, 1976; TALWANI, 1976; WITHERS and NYLAND, 1978) has been that the stress field is very close to critical and small increases in the pore pressure decrease the effective normal stress sufficiently to drive the rocks to failure. Different workers have assumed different levels of cohesive strength ranging from laboratory values to assumed zero cohesion for pre-existing fractures.

The first test of these models was provided by the experiment at Monticello Reservoir (ZOBACK and HICKMAN, 1982). *In situ* stresses, pore pressure and permeability values were obtained in two boreholes. In explaining the seismicity, Zoback and Hickman assumed the coefficient of friction $\mu$ to be 0.6 to 0.8, and cohesion to be zero. The value of $\mu$ was based on the work of BYERLEE (1978).

On examining cores collected in the two boreholes, three different groups of fractures were encountered. One set was subhorizontal open fractures – along exfoliation joints. The other set consisted of steeply dipping fractures filled with fluorides, pink zeolites and carbonates. In other cores, the material filling the fractures was mixed illite and smectites, including montmorillonite. These deposits were similar to others in the region resulting from hydrothermal action of hot water.

The presence of the expandable clays (montmorillonite, etc.) has also been noted in fault zones. CHU *et al.* (1981) note that the clay composition in the fault gouge between the depths of 300 and 1000 feet from the San Andreas fault zone is fairly uniform. The predominant components are montmorillonite (40%), kaolinite (40%) and 10% each of illite and chlorite (LIECHTI and ZOBACK, 1979). X-ray diffraction of a clay sample from Monticello Reservoir revealed the presence of montmorillonite clay. We do not know how widespread is the presence of montmorillonite and other expandable clays. WU *et al.* (1979) note that clay minerals have been encountered in fault zones to depths of 2 km, whereas WANG *et al.* (1978) suggest that clay rich fault gouge may exist down to 10 km on the San Andreas fault.

In old metamorphic terranes like the South Carolina Piedmont, where there is evidence of large scale hydrothermal activity, it is not unreasonable to expect that clays may be present in fractures at observed hypocentral depths.

There has been a series of excellent papers describing the frictional and other properties of materials in simulated or real fault gouge. See, for example, ENGELDER *et al.* (1975), SUMMERS and BYERLEE (1977), WANG and MAO (1979), CHU *et al.* (1981), MORROW *et al.* (1982), among others. Two U.S.G.S. Conference Reports on Experimental Studies on Rock Friction (EVERNDEN, 1977) and Analysis of Actual Fault Zones in Bedrock (EVERNDEN, 1979) also contain relevant data.

For rock on rock, BYERLEE (1978) showed for a wide range of rocks and at a wide range of stress conditions, that at normal stress ($\sigma_n$) above 2 kb shear stress ($\tau$) required to activate rock on rock sliding is given by $\tau = 0.5 + 0.6\ \sigma_n$, and for normal stress below 2 kb, $\tau = 0.85\ \sigma_n$. The coefficient of friction ($\mu$) being constant (0.6 to 0.85) for large varieties of materials led this relationship to be referred to as Byerlee's law (BRACE and KOHLSTEDT, 1980). However, for fault gouge and other clays, $\mu$ was found to be 0.4 to 0.2, and decreased to 0.2 and below when the clays were saturated (WANG and MAO, 1979). Increasing pore pressure may further reduce the coefficient of friction in montmorillonite (BIRD, 1984). Similar results were obtained by MORROW *et al.* (1982). This decrease in $\mu$ is due to the hydration of the clay minerals.

These observations suggested a possible explanation for the observed seismicity near Monticello Reservoir. From a detailed analysis of fault plane solutions occurring in various earthquake clusters near the Monticello Reservoir and a comparison with *in situ* fractures encountered in the boreholes, TALWANI (1981b) demonstrated that the seismicity is occurring along pre-existing fractures.

In Fig. 8, we have plotted in Mohr circle configuration the measured stresses at a depth of 312 m in deep well no. 2 and at a depth of 961 m in deep well no. 1.

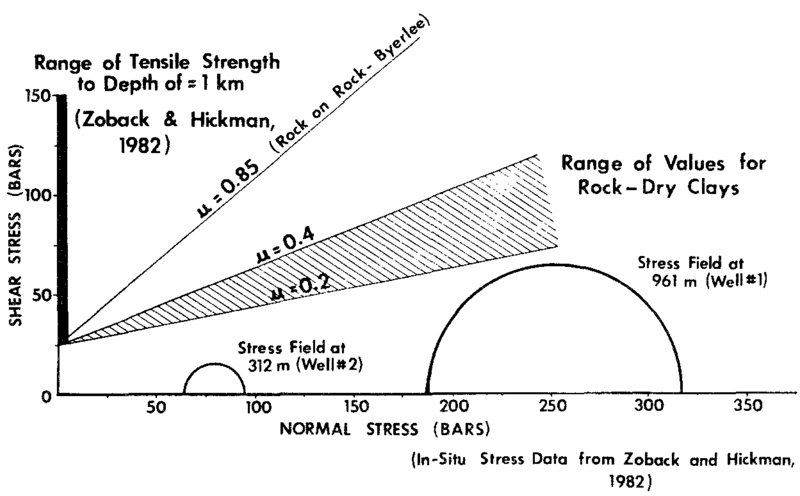


Figure 8

*In situ* stress data for Monticello Reservoir compared with failure envelopes for different coefficients of friction. Tensile strengths are those observed in well no. 2 and decrease to zero for pre-existing fractures.

The tensile strength measured by ZOBACK and HICKMAN (1982) ranges from 26 bars to 152 bars and has been indicated. Failure envelopes corresponding to $\mu = 0.85$ (Byerlee's value for these stress levels) and $\mu = 0.2$ and 0.4 ($\mu$ for dry clays) are also indicated for a tensile strength of 25 bars (corresponding to the lowest observed value). We note that the failure envelope lies over the stress values. Small changes in pore pressure (say by 10 bars) will shift the Mohr circles to the left, but not account for the observed seismicity, if we assume a coefficient of friction according to Byerlee's law.

We suggest that what in fact changes is the coefficient of friction. As the lake is impounded, pore pressure diffusion occurs, increasing pore pressure in the fractures. If the fractures were initially dry, then WANG and MAO's (1979) laboratory data suggest that there is a reduction in $\mu$ to 0.2–0.1 for pure clays. The experimental results of MORROW *et al.* (1982) confirm a similar trend at a variety of confining pressures with mixed clay fault gouge. This is accompanied by a reduction in the cohesive strength. If the rock was already saturated, there is still a reduction in the strength and $\mu$ with increasing pore pressure (BIRD, 1984). The failure envelope will now approach the Mohr circle and lead to earthquakes.

Following development by BRACE and KOHLSTEDT (1980), ZOBACK and HICKMAN (1982) examined their *in situ* stress data at Monticello deep wells 1 and 2 in an attempt to explain the observed RIS. By assuming a coefficient of friction ranging between 0.6 and 0.8, they concluded that the observed (thrust fault) seismicity should be concentrated at depths to 500 m only. However, well constrained hypocentral depths extend to at least 2 km. In view of the discussion in the previous paragraph, we suggest that the actual coefficient of friction lies in the range 0.2 to 0.4 instead of between 0.6 and 0.8 as predicted by Byerlee's law.

In Fig. 9 we compare the *in situ* stress data of ZOBACK and HICKMAN (1982) with the range of $S_{h\,\max}$ where seismicity would occur if the coefficient of friction lay in the range 0.2 to 0.4. These stress measurements were taken in July 1978 at Monticello deep well 1 and in January 1979 and August 1980 at Monticello deep well 2, some 7 to 32 months after the onset of induced seismicity at the reservoir. Earthquakes occurring in the vicinity of the wells would have already produced localized stress drops of up to 100 bars as estimated by ZOBACK and HICKMAN (1982). These data, therefore, may reflect the released stress state in these areas. We note that now the *in situ* stress data with the assumption of lower coefficients of friction are compatible with the observed hypocentral depths.

## *11. Conclusions*

It is now generally accepted that changes in pore pressure are the primary causes of observed RIS. In this paper we suggest that the mechanism by which changes in pore pressure are transmitted to hypocentral depths is by diffusion. Analyses of available RIS data further suggest that there is a characteristic range of 'seismic' hydraulic diffusivity ($10^3$–$10^5$ cm$^2$/s) and permeability (0.1–10 millidarcy)

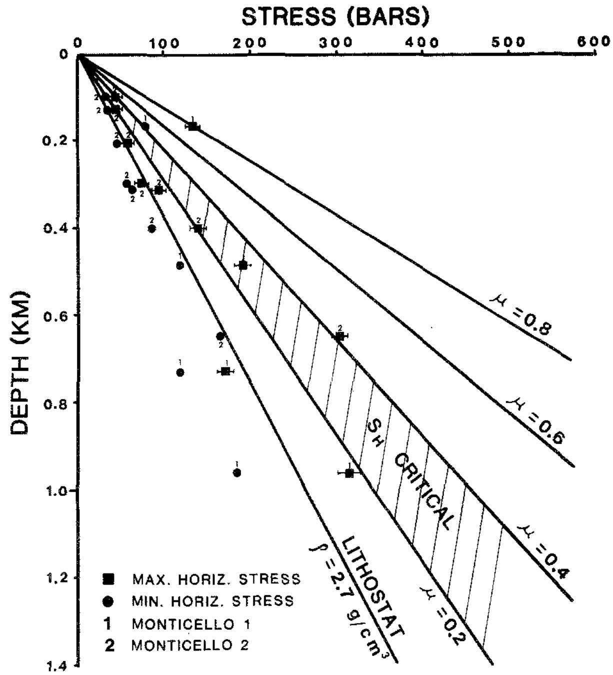


Figure 9
Hydraulic fracturing stress measurements at Monticello Reservoir deep wells 1 and 2 as a function of depth. The area labelled $S_H$ critical indicates the magnitude of maximum horizontal stress expected to result in reverse faulting along appropriately oriented fault planes assuming a coefficient of friction between 0.2 and 0.4. Stress data from ZOBACK and HICKMAN (1982).

values. Field data at Monticello Reservoir also indicate that the diffusing pore pressure plays a dual role in triggering seismicity, one by decreasing the coefficient of friction between the clays in the pre-existing fractures and the rocks that enclose these fractures, and two by decreasing the strength.

These results have a possible application in our understanding the physics controlling the growth of aftershock zones of tectonic earthquakes.

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