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On the Nature of Reservoir-induced Seismicity

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Abstract—In most cases of reservoir-induced seismicity, seismicity follows the impoundment, large lake-level changes, or filling at a later time above the highest water level achieved until then. We classify this as initial seismicity. This "initial seismicity" is ascribable to the coupled poroelastic response of the reservoir to initial filling or water level changes. It is characterized by an increase in seismicity above preimpoundment levels, large event(s), general stabilization and (usually) a lack of seismicity beneath the deepest part of the reservoir, widespread seismicity on the periphery, migrating outwards in one or more directions. With time, there is a decrease in both the number and magnitudes of earthquakes, with the seismicity returning to preimpoundment levels. However, after several years some reservoirs continue to be active; whereas, there is no seismicity at others. Preliminary results of two-dimensional (similar to those by ROELOFFS, 1988) calculations suggest that, this "protracted seismicity" depends on the frequency and amplitude of lake-level changes, reservoir dimensions and hydromechanical properties of the substratum. Strength changes show delays with respect to lake-level changes. Longer period water level changes ($\sim$1 year) are more likely to cause deeper and larger earthquakes than short period water level changes. Earthquakes occur at reservoirs where the lake-level changes are comparable or a large fraction of the least depth of water. The seismicity is likely to be more widespread and deeper for a larger reservoir than for a smaller one. The induced seismicity is observed both beneath the deepest part of the reservoir and in the surrounding areas. The location of the seismicity is governed by the nature of faulting below and near the reservoir.

Key words: Mechanism of reservoir-induced seismicity, Koyna, Monticello Reservoir, Lake Mead.

Introduction

Since the identification of a causal association of seismicity with the impoundment of Lake Mead in the early 1940s (CARDER, 1945) reservoir-induced seismicity (RIS) has been observed at over seventy locations worldwide (SIMPSON, 1976, 1986; GUPTA, 1992). Following damaging reservoir-induced earthquakes in the 1960s at Koyna, India; Hsingfengkiang, China; Kariba, Zimbabwe and Kremasta, Greece, there was great improvement in seismic monitoring. Local networks were deployed in the vicinity of several reservoirs in the 1970s. These resulted in lower detection thresholds and improved locations of recorded seismicity. Complementary field studies led to the identification of factors that control the observed RIS. These

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factors include ambient stress field conditions, availability of fractures, hydromechanical properties of the underlying rocks, geology of the area, together with dimensions of the reservoir and the nature of lake-level fluctuations.

As case histories of RIS accumulated, the effect of reservoir loading on the existing stress field has been the subject of several studies (SNOW, 1972; BELL and NUR, 1978; TALWANI and ACREE, 1984; SIMPSON, 1976, 1986; SIMPSON *et al.*, 1988; ROELOFFS, 1988; RAJENDRAN and TALWANI, 1992). Except for ROELOFFS' (1988) study, all of them addressed seismicity associated with the initial impoundment of the reservoir. However, there are other cases in which protracted and "significant" RIS has been observed several years after initial impoundment. "Significant" here implies both a larger number and a higher magnitude of seismicity than at preimpoundment levels. The seismicity in such cases appears to be related to water level fluctuations. The ongoing seismicity at Koyna, India, three decades after impoundment; the observed seismicity at Lake Mead for over three decades after the observation of the initial seismicity and to a lesser extent, the current seismicity at Lake Jocassee and Monticello Reservoir South Carolina are examples of this protracted RIS.

Initial and Protracted Seismicity

We classify the seismic response of a reservoir into two temporal categories. The first, which is widely observed, is associated with the initial impoundment or large lake-level changes. This category also applies to seismicity associated with lake-level increases above the highest level attained thus far. We call this category of RIS, "initial seismicity." The second category of seismicity, which is observed in rare cases, occurs after the effect of initial filling has diminished. It persists for many years without a decrease in frequency and magnitude. We call it "protracted seismicity."

The initial seismicity results from the instantaneous effect of loading (or unloading) and the delayed effect of pore pressure diffusion. Following this initial activity, there is an increase in the frequency and magnitude of earthquakes. The largest associated event usually occurs after completion of the reservoir impoundment and the attainment of maximum water level. The delay between the start of filling and the larger events varies from months to years and is associated with the reservoir and local site characteristics. Spatially there is a general stabilization and (usually) an absence of seismicity beneath the deepest part of the reservoir and widespread seismicity on the periphery, migrating outwards in one or more directions. This period of increased seismicity is followed by a gradual decay in activity (over months to years) to preimpoundment levels, indicating the cessation of the coupled poroelastic response to the impoundment.

In the case of protracted seismicity, modeling suggests that the pore pressure increase that causes the seismicity, is related to the frequency and amplitude of lake-level changes (ROELOFFS, 1988). Peak changes in pore pressures occur directly beneath the lake and decrease away from it. Strength changes show delays with respect to lake levels. In this category, earthquakes are associated with reservoirs with large and/or rapid lake-level rises and longer periods (lower frequencies) of water level changes. Seismicity is observed both beneath the deepest part of the reservoir and in surrounding areas. The seismicity continues for decades and does not appear to die out.

The factors controlling the spatial and temporal patterns of these two categories of RIS are different. In this paper I address the nature of these two categories of RIS.

Poroelastic Response to Reservoir Impoundment

BELL and NUR (1978) defined the change in strength ΔS by the following equation

$$\Delta S = \mu_f(\Delta\sigma_n - \Delta p) - \Delta\tau \tag{1}$$

where $\Delta\tau$ and $\Delta\sigma_n$ are changes in shear stress on the fault in the direction of slip and compressive normal stress across the fault respectively. μ_f and Δp are the coefficient of friction and change in pore pressure respectively. Failure occurs when ΔS decreases below a threshold level. From equation (1) we note that a decrease in ΔS can be brought about by a decrease in $\Delta\sigma_n$ (unloading) or an increase in pore pressure. The temporal effect of impoundment can be divided into two parts, instantaneous and delayed. (We use the poroelastic approach of RICE and CLEARY (1976) in which both the solid and fluid phases are assumed to be compressible.)

The instantaneous effect is due to the elastic and undrained response to loading. The delayed effect is due to the drained response and pore pressure changes by diffusion. The net result is a *coupled* response of the different responses mentioned above. These responses are shown schematically in Figure 1. The various effects have been reviewed by RAJENDRAN and TALWANI (1992). In order to present a comprehensive review here, and compare with field observations, the next section has been extracted from that paper and illustrated with Figure 1.

Elastic response: The elastic response of the subsurface to loading causes changes in normal and shear stresses on the fault plane. Under assumptions of isotropic conditions, $\Delta\sigma_n$ (Figure 1b) mimics the reservoir loading curve (Figure 1a), the change being instantaneous. In general, increased normal stress tends to stabilize (increase ΔS) the region, especially under the reservoir. An example of this is provided by comparing the RIS observed at Lake Mead with elevation changes found by releveling. Intense seismicity was observed following the impoundment of

Lake Mead behind the Hoover Dam in the late 1930s and early 1940s. CARDER and SMALL (1948) found that the epicenters were located near the periphery and were not associated with the region of maximum crustal load due to the lake.

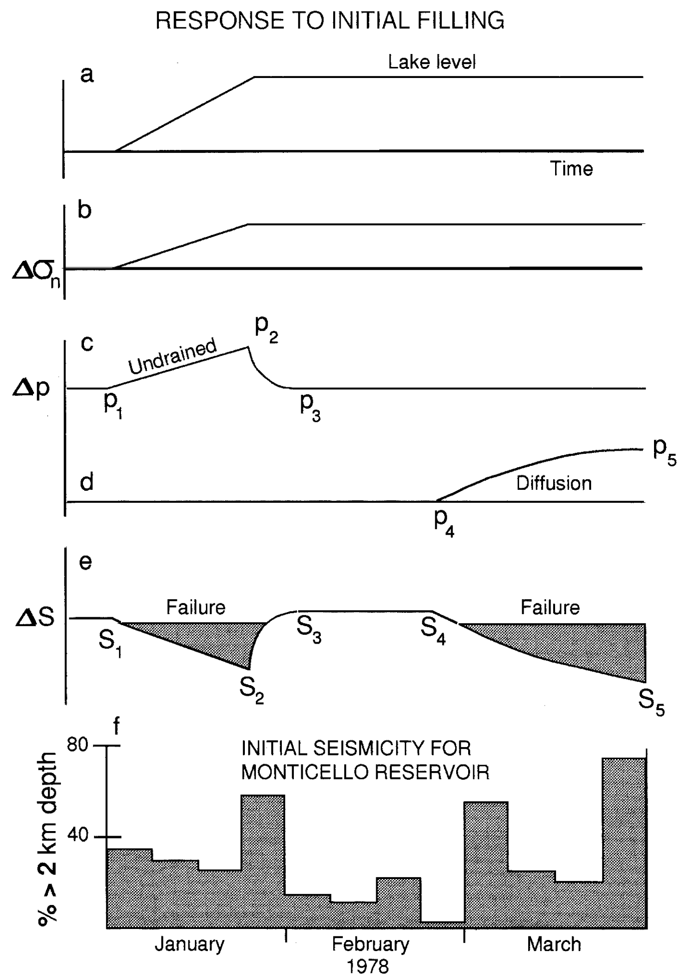

Figure 1

Schematic figure to illustrate the processes observed in initial seismicity. (a) shows the filling curve of the reservoir, it is associated with an increase in $\Delta\sigma_n$ (b) due to the load. The undrained response in a clogged pore causes an increase in the pore pressure (p_1 to p_2) (c) and a corresponding decrease in strength (S_1 to S_2) (e). When the pore is unclogged, the increased pore pressure dissipates (p_2 to p_3) and the strength increases (S_2 to S_3). When the pore pressure front due to reservoir loading arrives, there is an increase in pore pressure (p_4 to p_5, (d)) and a corresponding decrease in strength (S_4 to S_5, (e)). When the strength decreases below a critical threshold (marked FAILURE) seismicity occurs (shaded pattern). Panel (f) shows the percentage of "deep" events associated with the initial filling of Monticello Reservoir.

Undrained response: We borrow the nomenclature from soil mechanics, according to which undrained conditions prevail if the rock sample is subjected to a change in confining pressure and pore fluid is prevented from escaping or entering. In the case of reservoir impoundment, there will be an instantaneous increase in pore pressure in the substratum due to the additional load at the surface. If no fluid is allowed to flow, for example in the case of clay filled fractures, there will be an increase in pore pressure, Δp_u. This pore pressure increase, due to undrained response, Δp_u, will persist until it dissipates into the surrounding fractures. The undrained response is given by

$$\Delta p_u = B\sigma_{\kappa\kappa}/3 \tag{2}$$

where B is the Skempton's constant and $\sigma_{\kappa\kappa}$ is the mean stress. If clogged fractures are present, Δp_u can increase with the loading (p_1 to p_2 in Figure 1c) and be sustained until flow occurs. In such a case, there is a corresponding decrease in the strength, (S_1 to S_2), which can lead to failure, when the strength decreases below a threshold value (labeled FAILURE in Figure 1e).

Drained response occurs when the pore fluid is enabled to enter or leave and the pore pressure decreases to the original value. In the case of a clogged fracture, the drained response occurs when the fluid leaves it and Δp_u decreases to zero (p_2 to p_3 in Figure 1c). The drained response is delayed with respect to the initial impoundment and the delay depends on the hydromechanical properties, chemical composition of fluids (for stress corrosion), nature of clays, etc. The drained response results in a decrease in pore pressure and an increase in ΔS (S_2 to S_3 in Figure 1e).

Pore pressure diffusion from the surface to the substratum also causes an increase in pore pressure. Pressure flow is governed by the diffusion equation (JAEGER and COOK, 1969). In one dimension it is

$$\delta^2 p/\delta z^2 = 1/C(\delta p/\delta t) \tag{3}$$

where p is the pore pressure at depth z, t is time and C is the coefficient of diffusivity

$$C = k/\eta\beta \tag{4}$$

where k is the permeability of the rock, η is the viscosity of the pore fluid and β is the bulk compressibility of fluid-filled rocks. The pore pressure increase following impoundment is delayed, the lag depending on hydraulic diffusivity C (and hence permeability, k) and the distance. Equation (3) has a solution of the form

$$p(z, t)/p(0, 0) = 1 - \mathrm{erf}\,[z^2/4ct]^{1/2}. \tag{5}$$

The pore pressure increase due to diffusion (p_4 to p_5 in Figure 1d) may occur after the increase in $\Delta p_u(p_1$ to $p_2)$ has already dissipated. This pore pressure increase is associated with a decrease in strength (S_4 to S_5 in Figure 1e) and earthquakes occur when the strength decreases below a threshold value (labeled FAILURE in Figure 1e).

Coupled response: Actually what we observe is the coupled response of the different responses mentioned above. For isotropic fluid-saturated porous elastic medium, RICE and CLEARY (1976) calculated the coupled response

$$\sigma_{ij} = 2G\varepsilon_{ij} + \frac{v}{1+v}\,\sigma_{\kappa\kappa}\,\delta_{ij} - \frac{3(v_u - v)}{B(1+v)(1+v_u)}\,p\,\delta_{ij} \tag{6}$$

where v and v_u are drained and undrained Poisson's ratios and G is the shear modulus. Using RICE and CLEARY's (1976) results, ROELOFFS (1988) modified equation (5) to include the term due to the undrained response. For a unit step increase in pore pressure at the surface, $p(0, t) = H(t)$, she calculated the pore pressure at a depth z after time t, $p(z, t)$. For a one-dimensional case she found

$$p(z, t) = (1 - \alpha)\ \mathrm{erfc}\ [z^2/4ct]^{1/2} + \alpha(H(t)) \tag{7}$$

where erfc is the complementary error function, $H(t)$ is Heaviside unit step function and $\alpha = B(1 + v_u)/3(1 - v_u)$. Thus the coupled response may be dominated by the undrained response immediately on impoundment and be primarily due to diffusion later. At any depth, after enough time has elapsed, $p(z, t)$ approaches the load applied at the surface, there are slight changes in the pore pressure and the RIS decays to preimpoundment levels.

The above arguments are given for isotropic conditions. However, the presence of fractures on which RIS is usually observed, clearly suggests that anisotropic conditions prevail. In such a case, pore pressure increase can cause earthquakes on vertical fractures in a normal faulting environment and on horizontal fractures in a reverse faulting environment (CHEN and NUR, 1992).

An Example of Initial Seismicity

Figure 2 shows the filling curve at Monticello Reservoir, South Carolina superimposed on monthly seismicity. Impoundment ($\sim$32 m) occurred between December 3, 1977 and February 8, 1978. Thereafter the lake level was kept within 1.5 m of the mean water level. "Initial seismicity" persisted for a few years and then began to decay, approaching preimpoundment levels.

The spatial pattern of the earthquakes provides further insight into the nature of initial seismicity. Most of the seismicity was shallow ($z < 3$ km). We should anticipate loading to increase $\Delta\sigma_n$ and thus ΔS, the greatest increase occurring beneath the deepest part of the reservoir. The increased pore pressure (which tends to destabilize) thus has a greater effect on the periphery of the reservoir. In 1978, following impoundment, the initial seismicity surrounded the deepest part of the reservoir (Figure 3). The "deeper" seismicity ($z \geq 2.0$ km) associated with impoundment provides further insight into the relative roles played by undrained response and diffusion. For $C \sim 10^4$ cm^2/s the time for pore pressure to diffuse to depths, of,

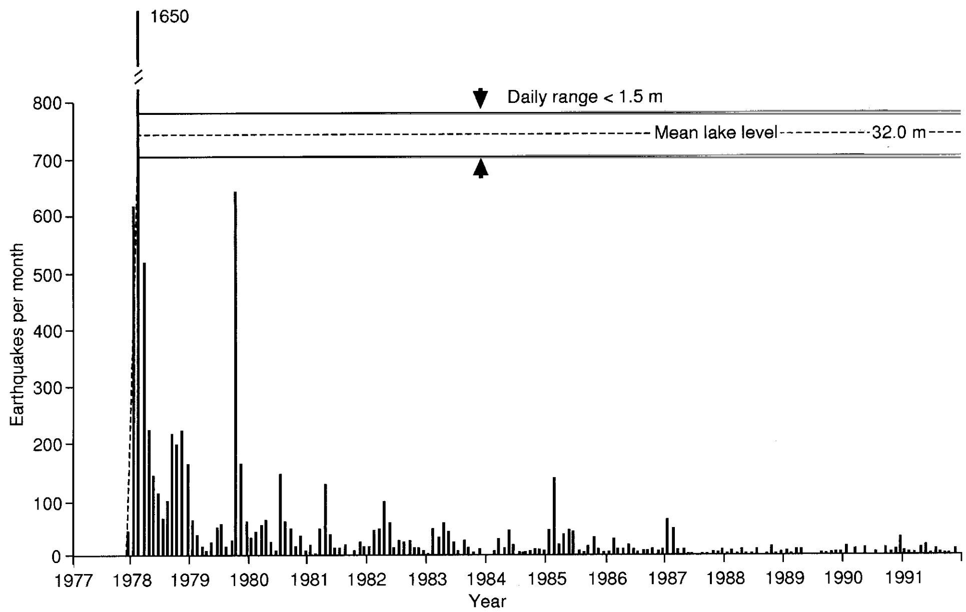

Figure 2
Lake level at Monticello Reservoir (dashed) compared with monthly seismicity for the period 1977 to 1991.

say 2 and 2.5 km is ~46 and 72 days. In January 1978, over 30% of the seismicity was deeper than 2 km (Figure 1f). We attribute this seismicity to increased pore pressure due to the undrained response of the reservoir. There was a decrease in the "deeper" seismicity in February, 1978 (<10%). We attribute this to leaking of the increased pore pressure (drained response, corresponding to p_2 to p_3 in Figure 1c). The decrease in pore pressure causes strengthening, (increase in ΔS) and thereby a decrease in the "deeper" seismicity (Figure 1f). The increased "deeper" activity (>40%) in March, 1978 is attributed to increased pore pressure due to diffusion.

Thus Figure 1f shows the different aspects of initial seismicity. The seismicity in January 1978 was due to both the undrained response (deeper earthquakes) and to diffusion (shallower earthquakes). The initial undrained response dissipated by February, 1978 and the ensuing seismicity was primarily due to diffusion. The initial seismicity was located outside the deepest part of the reservoir.

The burst of seismicity in 1985 was located beneath deeper parts of the reservoir. The nature of this protracted seismicity is different from the initial seismicity and is described in the next section.

Initial seismicity is the most widely observed category of RIS. Most case histories where adequate data are available, can be explained by the coupled response to initial loading, loading at a later time above the highest water level achieved, thus far and due to rapid water level changes. Examples of these include the observed seismicity at Lake Mead (CARDER, 1945); Nurek (SIMPSON and NEGMATULLAEV, 1981); Manic-3 (LEBLANC and ANGLIN, 1978); Kariba (SIMPSON *et al.*, 1988); Hsinfengkiang (SHEN *et al.*, 1974); Lake Jocassee (TALWANI *et al.*, 1976). In all of these cases there was an increase in seismicity above preimpoundment levels; large event(s) followed filling and there was a decay in seismicity to preimpoundment levels. Where accurate data are available, the initial seismicity appears to occur away from the deepest part of the reservoir and migrates outwards.

Protracted Seismicity

Unlike the case of initial seismicity, at some reservoirs seismicity continues for several years, even decades after impoundment. Figure 4 shows the seismicity and lake levels at Koyna for the period 1961–1995. The impoundment of the reservoir

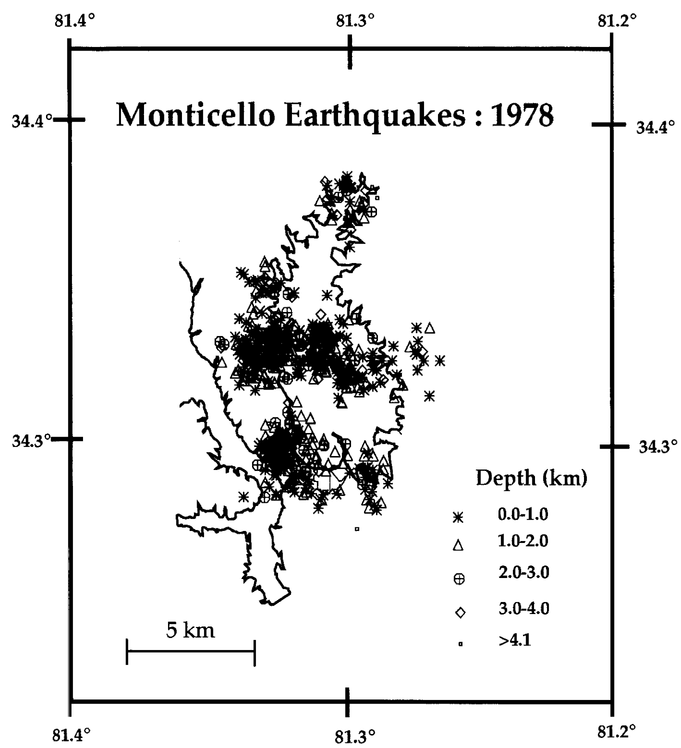

Figure 3
Seismicity observed near Monticello Reservoir in 1978. Note that most of the earthquakes lie in two bands—in the middle of the reservoir where the water is relatively shallow and on the southwestern and southern banks of the reservoir. There is a general absence of seismicity below the deepest (and southernmost) part of the reservoir.

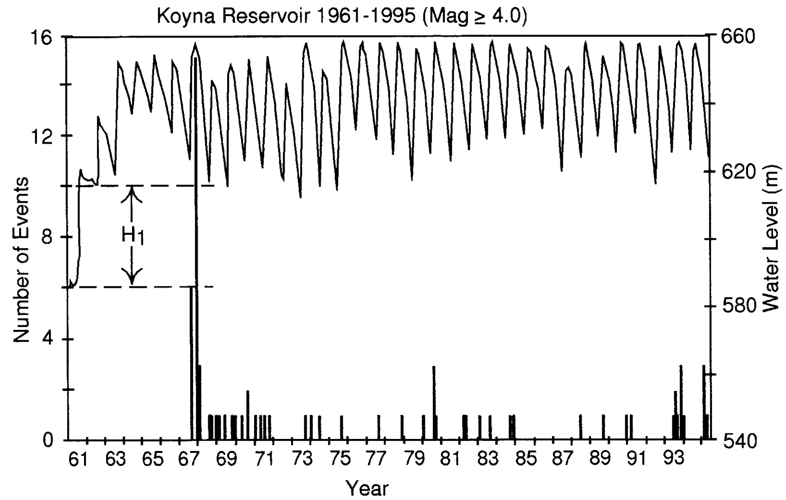

Figure 4
Monthly seismicity ($M \geq 4.0$) at Koyna Reservoir compared with lake levels for the period 1961 to 1995. (Figure taken from TALWANI *et al.*, 1996.)

occurred between 1962 and 1964. Earthquakes continue to occur there; the latest $M > 5.0$ event occurring in February, 1994.

This seismicity, occurring long after the strength changes associated with the coupled response to initial filling have stabilized, is relatable to the large water level changes. These are shown schematically in Figure 5. The daily/weekly/annual change H_2 is usually considerably less than the least depth of water, H_1 for the case of initial seismicity. This was the case at Monticello Reservoir where $H_1 \sim 31.8$ m and $H_2 \sim 1.5$ m (Figure 2). When the change in water level (weekly/monthly/annual) H_2 is comparable or a large fraction of the least depth of water, H_1, the seismicity is governed by the frequency lake-level changes. For Koyna H_2 is 20 to 40 m (with largest value ~ 47 m) and $H_1 \sim 30$ m.

This effect is illustrated by observations from Lake Mead impounded by the Hoover Dam. Figure 6 shows the lake level and the time of occurrence of larger earthquakes ($4.0 \leq M \leq 5.0$). (The revised magnitudes and lake levels are taken from a report by ANDERSON and O'CONNELL (1993).) Two large events (M 5.0 and 4.4) occurred in May and June 1939 following initial impoundment ($H_1 > 100$ m). Ten other events with $4.0 \leq M \leq 4.9$ took place in the period between 1942 and 1963. These events happened following large annual changes in the lake levels (15 m $\leq H_2 \leq 30$ m) (Figure 6). Four events with $M \geq 4.0$ occurred between 1963 and 1965. These were related to elastic unloading (decrease in $\Delta\sigma_n$ and hence ΔS). The average magnitude of the larger events between 1939 and 1963 was 4.3. Lake-level changes (H_2) were less than 10 m following the construction in 1965 of Glen Canyon Dam located upstream of Lake Mead. No earthquake with $M > 3.7$ has occurred since 1965. The mean magnitude of the 13 largest events in the period 1965–1992 is 3.3. These observations illustrate that the amplitude of lake-level changes (H_2) plays an important role in protracted seismicity.

ROELOFFS (1988) computed the coupled effect of pore pressure changes due to a cyclic load. She also incorporated the effect of the location and type of faulting. She found that the effect of an oscillating reservoir depended on where it was located with respect to the fault and the nature of the faulting. The oscillating reservoir maintained a stabilizing effect if it was located on the hanging wall of a steeply dipping reverse fault, directly above a shallowly dipping thrust fault or if there was a shallow vertical strike-slip fault or normal fault at the reservoirs edge (Figure 7a). However, destabilization (earthquakes) occurred if the reservoir was

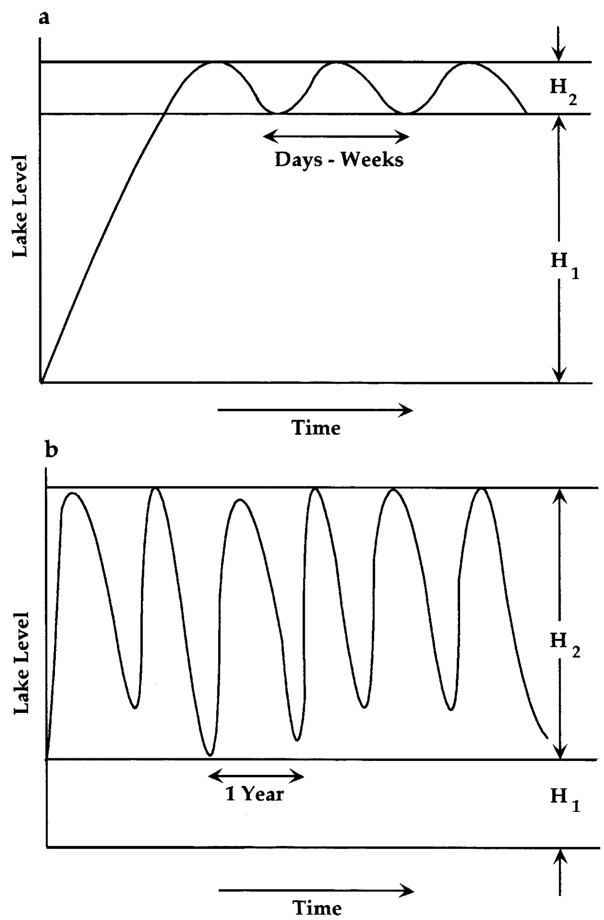

Figure 5
Comparison of the amplitude of cyclic lake level changes (H_2) with the least water level (H_1). For initial seismicity $H_2 \ll H_1$ and for *protracted* seismicity H_2 is comparable to H_1.

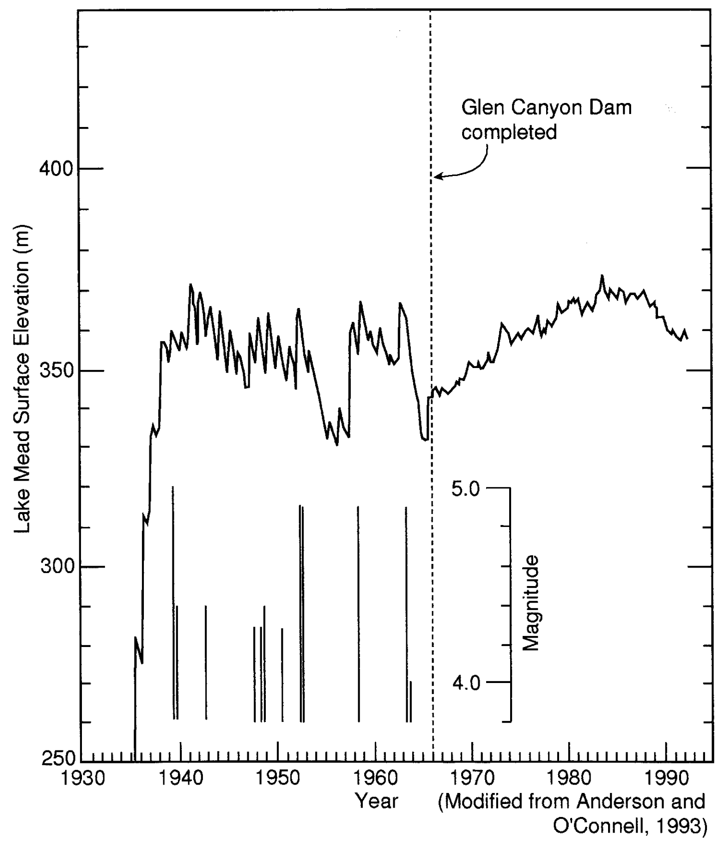

Figure 6
Larger earthquakes ($4.0 \leq M \leq 5.0$) at Lake Mead compared with lake levels for the period 1935–1992. Note a decrease in larger events after construction of Glen Canyon Dam in 1965. (Modified from ANDERSON and O'CONNELL, 1993.)

located on the foot wall of a steeply dipping reverse fault or on the hanging wall of a shallowly dipping thrust. Seismicity also occurred below the reservoir if there was a vertical strike-slip fault or a normal fault located there (Figure 7b).

Her calculations for a simple 2-D reservoir suggested that the changes in stress and pore pressure fields produced by reservoir loads are governed by a dimensionless frequency Ω, given by

$$\Omega = \omega L^2/2C, \tag{8}$$

where ω is loading frequency (1/year for Koyna and 1/day for Monticello Reservoir) and L is the width of the reservoir. The depth, z^*, below which pore pressure changes were negligible is given by

$$z^* = \Pi(2C/\omega)^{1/2}. \tag{9}$$

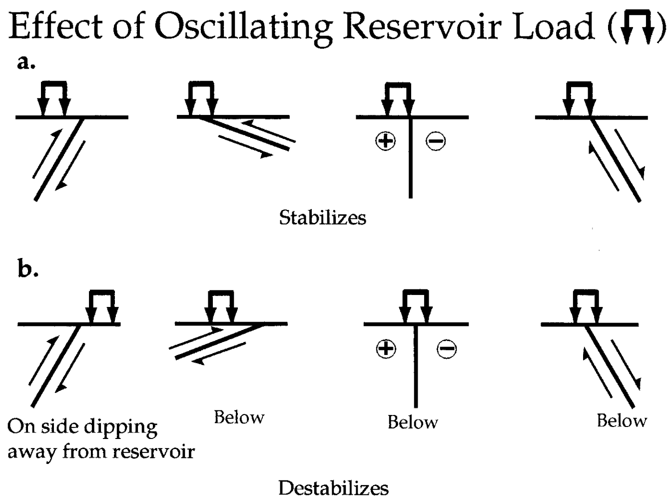

Figure 7
Schematic figure to show the effect of an oscillating reservoir load. It results in stabilization or seismicity, depending on the location and orientation of faults *vis à vis* the reservoir and on the stress field, based on ROELOFFS (1988).

Table 1 illustrates that the deepest effects occur in regions with higher diffusivity and longer periods. The model would predict pore pressure changes to a depth of about 25 km below Koyna for an assumed diffusivity value of 1 m^2/s and an annual cycle of lake-level changes. A detailed analysis provided good estimates of the depth extent of recent seismicity (1993–95) (TALWANI *et al.*, 1996). Current seismicity lies to the south of the reservoir and between depths of 5 and 16 km.

However, as ROELOFFS (1988) demonstrated, the location of maximum destabilization also depended on the nature of faulting. For a strike-slip and a reverse fault below the reservoir, we calculated the maximum change in strength over the entire cycle of loading (TALWANI *et al.*, 1992). The results are shown in Figures 8 and 9.

Table 1

z^* for various periods and diffusivities

Diffusivity m^2/s	Peroid (days) 1	30	360
	z^* km	z^* km	z^* km
0.1	0.4	2.3	7.8
1.0	1.3	7.2	24.8
10.0	4.1	22.6	78.4

z^* Depth below the surface, below which the coupled pore pressure changes due to lake-level changes are negligible.

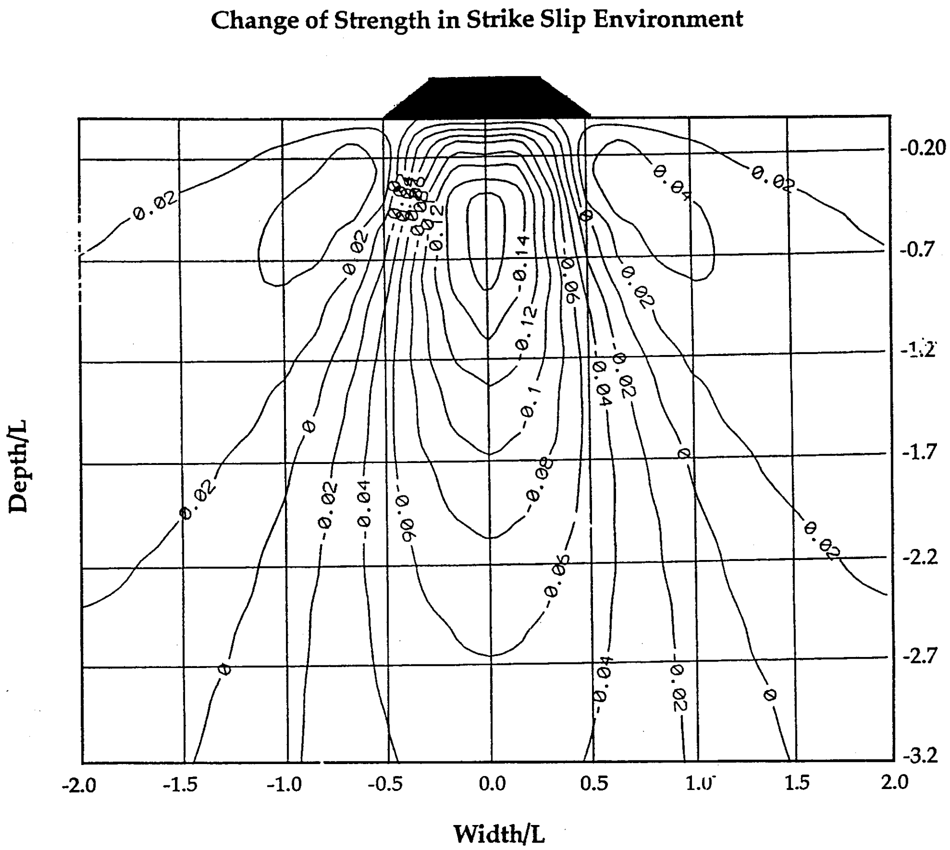

Figure 8

Maximum change in strength (in bars) over the entire cycle of lake-level change. A vertical fault is assumed directly below the reservoir (shaded). The 2-D reservoir and fault extend into the plane of the paper. The cyclic load varies from $+1$ bar to -1 bar over the 360 days cycle. The depth and width are normalized with respect to the width of the reservoir.

To calculate the strengths we modified Equation (7) and assumed parameters given in Table 2.

For the 2-D model we calculated the changes in strength (in bars) corresponding to a cyclic load (p_0 varying from $+1$ bar to -1 bar) ($+0.1$ MPa to -0.1 MPa). The depth and horizontal distances are normalized with respect to the width of the reservoir. For a cyclic load corresponding to p_0 of say ± 5 bars (0.5 MPa) (corresponding to water level changes of ± 50 m) the changes in strength will be multiplied by 5. Negative values of changes in strength correspond to weakening. The change in strength occurs over the entire cycle. The maximum change in strength over the entire cycle is plotted. In the case of the strike-slip fault, the largest changes occur under the reservoir (Figure 8), and for a reverse fault dipping 60° to the left weakening occurs on the left bank of the reservoir (Figure 9). To obtain actual depths and distances in Figures 8 and 9 multiply by the width of the reservoir. For a strike-slip fault the largest changes in pore pressure occur at depths between about 0.5 to 0.7 times the width of the reservoir, whereas for the reverse fault dipping at 60° and a daily cycle in water level changes (such as that observed at Monticello Reservoir), the largest changes in pore pressure occur in the middle and on the side of the fault dipping away from the reservoir. The location of seismicity in 1985 was significantly different from the initial seismicity observed in 1978 (Figure 3), and did in fact occur beneath and to the west of the reservoir. Geological data and focal mechanisms confirm the presence of steep westward dipping faults on the western edge of Monticello Reservoir.

For reservoirs lying in a compressive stress regime we would anticipate reverse faulting on shallow, dipping faults or strike-slip faulting on steeply dipping faults. For regions where reverse faulting dominates, the weakening would be similar to that observed at Monticello Reservoir. It would be more widespread and deeper if

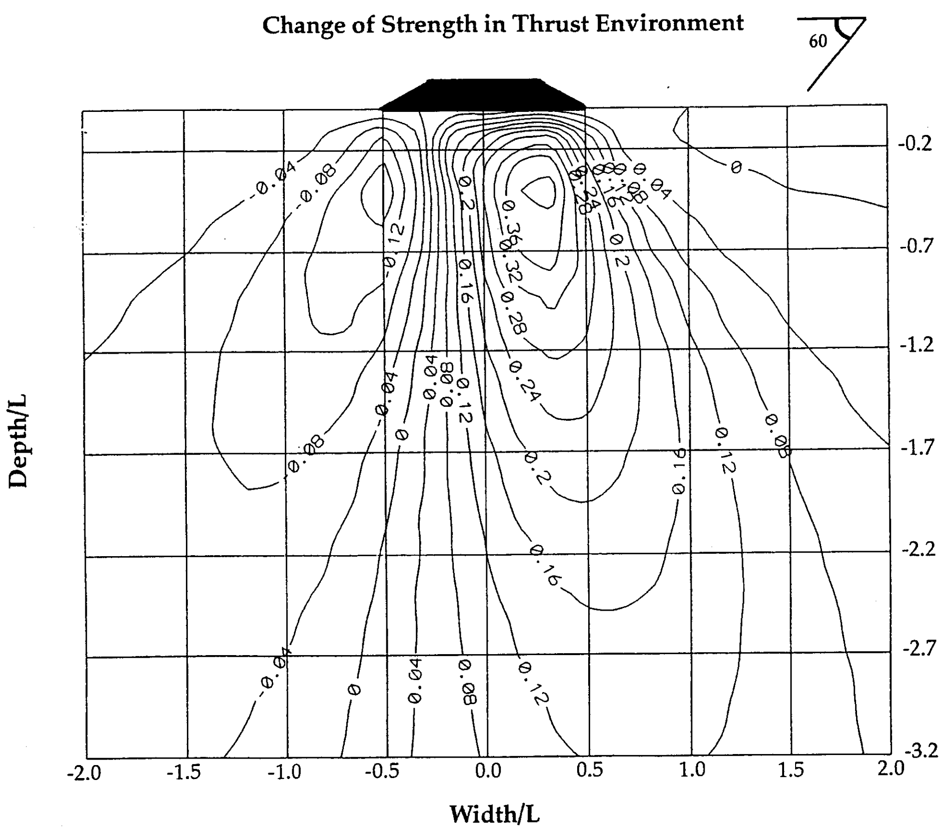

Figure 9
Maximum change in strength (in bars) over an entire cycle of lake-level change. A reverse fault below the reservoir dips 60° to the left. Weakening occurs below and to the left of the reservoir.

Table 2

Model parameters

	Strike-slip fault (Koyna)	Reverse fault (Monticello)
B	0.7	0.7
C	1 m^2/s	1 m^2/s
v_u	0.3	0.3
v	0.25	0.25
Period ($1/\omega$)	360 days	1 day

it is associated with longer periods of water level cycles, compared to Monticello Reservoir (1 day) (Table 1). The temporal changes in pore pressure and strength to the left, below and to the right of the reservoir at depths equal to half the width of the reservoir, over half a cycle are shown in Figure 10. Weakening occurs when the

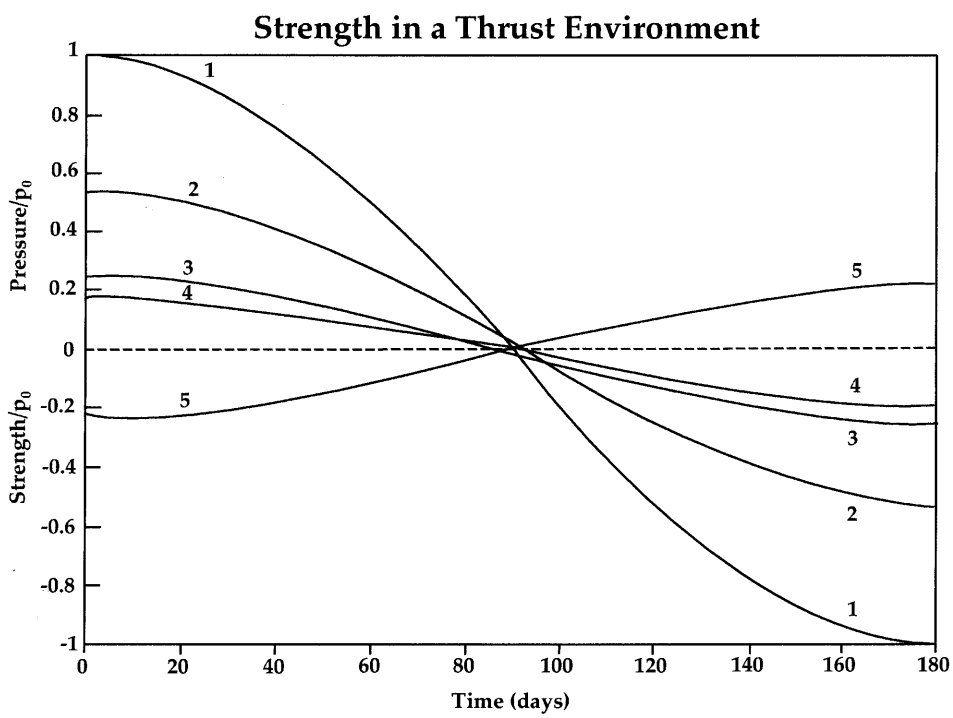

Figure 10

Temporal changes associated with 1/2 cycle of reservoir water level change. The water level change (curve 1) (corresponding to $p_0 = +1$ to -1) results in a change in pore pressure (curve 2). Pore pressure changes on the left bank, center and right bank (curves 5, 4 and 3) are calculated at depths equal to half the width of the reservoir.

ratio of strength to pore pressure becomes negative, and strengthening takes place when this ratio is positive. The fault is assumed to lie underneath the reservoir and dip 60° to the left. The change in pore pressure (curve 2) follows the lake level (load, curve 1), with a lag. At full reservoir there is weakening on the left side of the reservoir (decrease in ΔS, curve 5) and a strengthening on the right side of the reservoir (curve 3). The change in strength below the middle of the reservoir (curve 4) is similar to that on the right side. At minimum water level (day 180), there is a maximum weakening on the right bank of the reservoir and strengthening below the left bank. The weakening below the right bank at minimum load is comparable to the weakening below the left bank at maximum load. Thus, depending on the *in situ* conditions prevailing below the reservoir, we can have seismicity on the left side or on the right side. Also note that the changes in strength below different parts of the reservoir are out of phase with the water level curve. The delays depend on the hydraulic diffusivity, Skempton's constant, geometry of the reservoir, frequency of water level changes and fault geometry.

Discussion

We divide the temporal pattern of RIS into two categories. The first is associated with initial impoundment, the raising of water level above the highest water level achieved until then. The poroelastic response of the reservoir is a coupled response. Initially and occurring simultaneously with the impoundment is the undrained response. This occurs because of an increase in pore pressure in closed pores (by fault gouge and clay). As the increased pore pressure diffuses to the surrounding regions, there is a decrease in pore pressure (drained response). With the arrival of a diffusive pore pressure front, the pore pressure increases and causes seismicity. In reality all three effects occur together and the coupled response of the reservoir depends on which effect dominates. The time for an increase in pore pressure due to diffusion depends on the depth of the reservoir, geometry, availability of faults/fractures, etc. For shallow reservoirs the coupled response may take a few weeks to a few months (e.g., at Monticello Reservoir (TALWANI and ACREE, 1987), whereas for large and deep reservoirs it may take years (e.g., Hsingfengkiang SHEN *et al.*, 1974), Nurek (SIMPSON and NEGAMATULLAEV, 1981), etc. In both cases however we classify the temporal pattern of seismicity as initial seismicity. The initial seismicity is characterized by a general lack of seismicity beneath the deepest part of the reservoir and activity on the periphery of the reservoir. The seismicity increases after the impoundment is completed (or highest water level is achieved) and the largest earthquake usually occurs after that. Then there is a decay in seismicity (over 5–10 years) to preimpoundment levels.

SIMPSON *et al.* (1988) suggested that the temporal distribution of induced seismicity following the filling of large reservoirs exhibits two types of response;

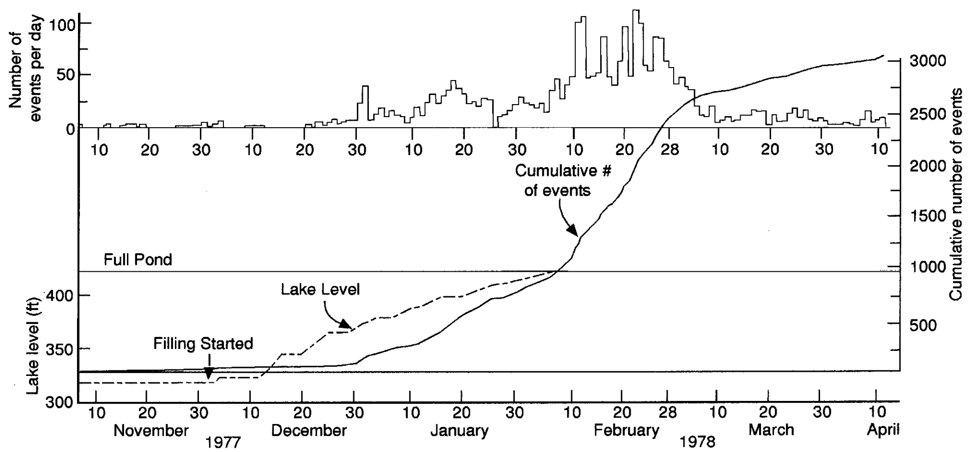

Figure 11
Seismicity associated with the filling of the Monticello Reservoir. (From TALWANI and ACREE, 1984.)

instantaneous and delayed. They suggested that the seismicity that began almost immediately following initial impoundment was due to the instantaneous elastic response and the undrained response, resulting in an increase in pore pressure. The delayed seismicity e.g., at Koyna, was attributed to an increase in pore pressure due to the diffusion of pore pressure to hypocentral depths. However, in our classification, both types of RIS alluded to by SIMPSON *et al.* (1988), are integral components of initial seismicity.

This is illustrated by thoroughly examining the observed seismicity at Monticello Reservoir in 1978. Filling occurred between December 3, 1977 and February 8, 1978 (Figure 11) (TALWANI and ACREE, 1984, 1987). The curve showing the cumulative seismicity is the most instructive. There are three distinct breaks in the slope of the curve. The first occurs about three weeks after the beginning of reservoir filling. It is associated with the start of RIS. The increased seismicity continues until February 8, 1978 (the date the reservoir is filled) when there is a second break in the cumulative seismicity curve. This third leg of the curve is associated with an increased rate of seismicity, which persists until about March 7, 1978. A lower rate of seismicity, last leg, follows thereafter. The seismicity between about the last week of December and February 8, 1978 is a result of three effects. These are the increasing stabilizing effect due to the reservoir load (especially below the deepest part of the reservoir), the destabilizing effect due to an instantaneous increase in pore pressure (undrained effect) and a delayed increase in pore pressure due to diffusion. The persistent three week lag between the bends in the filling curve and the cumulative seismicity curve attests to the role of diffusion in the observed seismicity. In the second phase of seismicity, between about February 8, 1978 and March 7, 1978, there is no longer an increase in the stabilizing effect (the reservoir has already been filled) although the effect of diffusion associated with increasing lake levels during the last three weeks of filling dominates, and results in the increased seismicity. After February 8, 1978, there are few (or no) changes in the lake levels and the seismicity is predominantly due to diffusion and possible undrained effects.

In summary, the seismicity following impoundment at Monticello Reservoir displays a coupled poroelastic response that includes pore pressure diffusion and an undrained response, and not ONLY an undrained, instantaneous response that SIMPSON *et al.* (1988) suggest. The difference arises because TALWANI and ACREE (1984) assume that diffusion and the undrained response begins when filling starts (December 3, 1977) and therefore by the time the reservoir is filled, February 8, 1978 (Figure 11), pore presure has diffused to hypocentral depths. SIMPSON *et al.* (1988) on the other hand, start the clock after the reservoir is filled (February, 1978) and thus the seismicity observed in February 1978 is classified as "rapid" or "instantaneous," and is *not* attributed to diffusion.

It is clear that "instantaneous" elastic and diffusion related responses are functions of the reservoir size and depths of earthquakes. The coupled, poroelastic, response at a large reservoir would take considerably longer to be fully manifested than for a small shallow reservoir like Monticello Reservoir. The coupled poroelastic response due to impounding at Monticello Reservoir thus appears to be instantaneous when compared with the response of larger reservoirs, although both cases display the same coupled poroelastic response.

SIMPSON *et al.* (1988) classify the seismicity at Koyna as "delayed response," and note that the major burst of activity did not occur until a number of annual filling cycles had passed. They attribute the seismicity as being dominated by diffusion of pore pressure. Our analysis suggests that the ongoing seismicity at Koyna is a case of protracted seismicity. The seismicity is the coupled poroelastic response to annual lake-level fluctuations.

For reservoirs where the lake-level changes are large, typically a large fraction of the least water depth, and have a longer period ($\sim$1 year), seismicity continues long after the initial coupled response due to impoundment is over. This protracted seismicity is rare; the two best cases being the seismicity observed at Lake Mead, U.S.A. and Koyna Reservoir, India. Protracted seismicity depends on the frequency of lake-level changes, the reservoir dimensions, hydraulic diffusivity and the geometry of faults *vis à vis* the reservoir. The longer the period of water level changes, the deeper and more pronounced are the effects, often over 10 km, for an annual cycle, and 100 s of m to a few km for daily or weekly cycles.

Thus in cases where the lake level depends on the annual rainfall, it is difficult to control the annual cycle of seismicity. However, for pumped storage dams, where the lake level is controlled by pumping water from the lower reservoir at times of low power demand, the seismicity is restricted to shallow depths and low magnitudes.

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