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C. H. Scholz
Columbia University

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TWO TYPES OF RESERVOIR-INDUCED SEISMICITY

BY D.W. SIMPSON, W.S. LEITH,[1] AND C.H. SCHOLZ

ABSTRACT

The temporal distribution of induced seismicity following the filling of large reservoirs shows two types of response. At some reservoirs, seismicity begins almost immediately following the first filling of the reservoir. At others, pronounced increases in seismicity are not observed until a number of seasonal filling cycles have passed. These differences in response may correspond to two fundamental mechanisms by which a reservoir can modify the strength of the crust—one related to rapid increases in elastic stress due to the load of the reservoir and the other to the more gradual diffusion of water from the reservoir to hypocentral depths. Decreased strength can arise from changes in either elastic stress (decreased normal stress or increased shear stress) or from decreased effective normal stress due to increased pore pressure. Pore pressure at hypocentral depths can rise rapidly, from a coupled elastic response due to compaction of pore space, or more slowly, with the diffusion of water from the surface.

INTRODUCTION

Increased earthquake activity has been associated with the filling of a number of large reservoirs (Simpson, 1976, 1986; Gupta and Rastogi, 1976). In some cases, the direct correlation of pronounced increases in seismicity with the first filling of the reservoir makes the causal relationship obvious; however, there are many cases in which there remains doubt as to whether the reservoir was directly responsible for increased seismicity. The most conclusive cases for induced seismicity are those relatively rare instances where there are data available from detailed monitoring of the reservoir region prior to impounding and where there is a substantial increase in seismicity on first filling of the reservoir (e.g., Nurek reservoir; Simpson and Negmatullaev, 1981). Even in the absence of detailed monitoring using sensitive instruments, the onset of felt earthquakes with the initial impounding is sometimes sufficient to establish a correlation between filling and seismicity (e.g., Hoover, Koyna, Kariba, Hsinfengkiang). In other cases there may be considerable delay between the initial filling and the start of detectable seismicity (e.g., Oroville, Aswan).

Two processes of stress modification have been suggested as the dominant mechanisms responsible for earthquake triggering by large reservoirs (Snow, 1972; Bell and Nur, 1978; Simpson, 1986; Roeloffs, 1988)—(1) the direct effect of loading, through increased elastic shear stress; and (2) the effect of increased pore pressure, through decreased effective normal stress. Increased pore pressure at depth can either be due to the volumetric strain component of the elastic field producing a decrease in pore volume or result from diffusion of pressure from the reservoir at the surface. There are substantial differences in the temporal and spatial characteristics of the response of the crust to these processes and it should be possible to identify the dominant mechanism in some cases, through a comparison of changes in seismicity with water level in the reservoir.

A comparison of well-documented case histories of seismicity at reservoirs suggests that two primary types of induced seismicity are observed: (a) a *rapid response*,

[1] Present address of W. S. Leith: USGS, Reston, Virginia, 22092

in which the seismicity follows immediately on first loading of the reservoir, consists primarily of low magnitude swarm-like activity, is confined to the immediate reservoir area, and is closely correlated with changes in water level in the reservoir; and (b) a *delayed response*, in which the seismicity follows with a significant delay after first filling, is often associated with large magnitude earthquakes, may extend significantly beyond the confines of the reservoir, and may not show an immediate correlation with major changes in reservoir level. Often, the reservoir water level may go through a number of apparently similar annual cycles in water level between first impounding and the onset of significant seismicity.

The first type we recognize as dominated by an *elastic* response to the load of the reservoir, both through an increase in elastic stress and/or an increase in pore pressure induced by elastic compression of pore space (i.e., not through transfer of water from the reservoir). The second type is dominated by the *diffusion* of pressure from the reservoir to hypocentral depths. Both responses are possible at any given reservoir. In general, induced seismicity will result from a complex interaction between the elastic stress increase, the cumulative effects of increased pore pressure from elastic and diffusion mechanisms and the preexisting stress regime (Bell and Nur, 1978; Simpson, 1986; Roeloffs, 1988). We propose here that through a comparison of the temporal variations in water level and seismicity at different reservoir sites, it is possible in some cases to identify the dominant mechanism responsible for increased pore pressure and thus to place additional constraints on the hydraulic properties of the crust.

Temporal Characteristics of Seismicity at Reservoirs

The temporal distribution of seismicity at reservoirs depends on two time-dependent phenomena—the response of the crust to the loading process and the way in which the load varies with time. In a simplified way, the response in seismicity can be thought of as the convolution of the response to a Heaviside function (crustal response to a step increase in water level) with a time-varying forcing function (water level changes in the reservoir). The crustal response to loading contains fundamental information on the elastic and hydraulic properties of the crust. We wish to isolate the nature of this response from that imposed by the influence of the water changes in the reservoir. The way in which the reservoir level changes with time is partially controlled by the size and geometry of the reservoir, and the watershed and climate conditions—small reservoirs in narrow canyons on fast flowing rivers fill much more rapidly than large, broad reservoirs in relatively flat topography. In addition, the control of water flow through a dam for irrigation or power generation can strongly influence the rate of change in water level. In the descriptions of water level changes that follow, general phrases such as rapid or slow will be used to describe the filling rates, but it must be remembered that these are relative both between different filling cycles of a particular reservoir and between reservoir sites. For reference in comparing the following figures (Figures 2 to 7) of seismicity at specific reservoir sites, Figure 1 shows the water levels at a number of reservoirs, plotted with the same time and height scales.

Examples of Two Types of Induced Seismicity

Rapid response

A rapid response is indicated by either an immediate increase in seismicity on first filling of a reservoir or an abrupt change in seismicity following a rapid change in water level. The identification of a rapid response on first filling may depend on

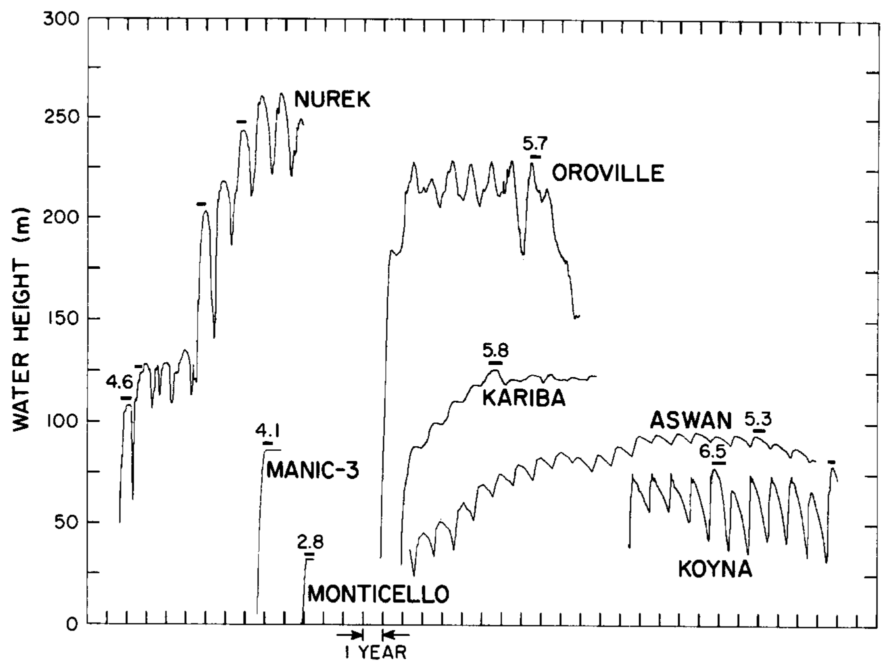

FIG. 1. Water levels and times of main earthquake activity at selected reservoirs showing induced seismicity (from Simpson, 1986). The water heights are relative to the base of the dam, so that direct comparison of water depths can be made. Time scales for all plots are the same but the origins have been adjusted for clarity in plotting. Numbers indicate magnitudes of largest events; bars are times of significant increases in seismicity. Details of the seismicity at each site are shown in Figures 2 to 7.

the presence of a sensitive monitoring network near the reservoir. This type of seismicity may be more common than the reported number of cases would indicate, because so few reservoirs have sufficient instrumentation. Monticello and Manic-3 are examples of relatively small reservoirs which filled rapidly and where instrumentation was sufficient to identify a rapid response to first filling. Nurek and Kariba are two large reservoirs showing rapid response which contrast significantly with Monticello and Manic-3 in size and tectonic setting. They both show a rapid response on first filling and Nurek also shows a clear rapid response to later changes in water level.

Monticello and Manic-3. Monticello (South Carolina; Talwani and Acree, 1985; Zoback and Hickman, 1982) and Manic-3 (Quebec; LeBlanc and Anglin, 1978) are the clearest examples of reservoirs which show rapid response of seismicity to filling (Figure 2). Manic-3 has a dam height of 108 m and a reservoir volume of 10.4 km^3; Monticello has a dam height of 52 m and reservoir volume of 0.5 km^3. Both have similar filling histories since they are pumped-storage facilities in which water is pumped back into the reservoir during off-peak hours, maintaining an almost constant water level in the reservoir after the initial rapid filling. At both sites, swarm-like seismicity began within weeks after impoundment, reached a peak near the time of first filling, and gradually decayed as the water level remained constant. At Manic-3, the largest earthquake was of magnitude 4.1 and occurred about one month before the water level had reached maximum level. At Monticello the largest earthquake was of magnitude 2.8 and the highest level of activity immediately followed the attainment of maximum water level. The activity in both cases is confined to the immediate reservoir area. Focal mechanisms in both cases show

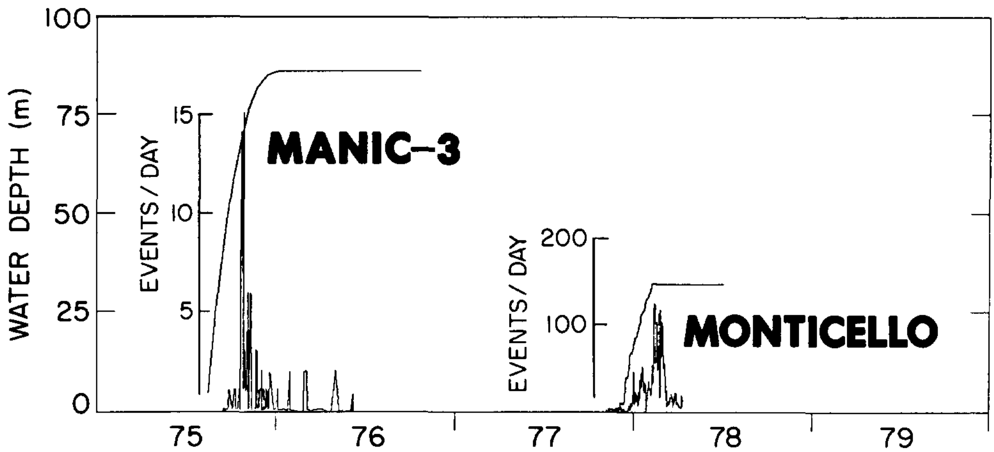

FIG. 2. Water level and number of earthquakes per day at Manic-3 and Monticello reservoirs. Water level and seismicity data were digitized from figures in Leblanc and Anglin (1978) for Manic-3 and Zoback and Hickman (1982) for Monticello.

reverse faulting and focal depths are extremely shallow, with most events less than 2 km deep.

Nurek. The 315 m high Nurek dam on the Vakhsh River in Tadjikistan, USSR, is the highest dam in the world, and impounds a reservoir volume of 10.4 km^3. The reservoir filled in two main stages—the first stage to over 100 m in 1972 and the second stage to over 200 m in 1976 (Figure 3). Annual variations in water level can exceed 50 m, with the rapid filling in late summer corresponding to runoff from snow melt in the high mountains of the Pamir and Tien Shan and the main outflow controlled by irrigation in the spring. The reservoir is located in the Tadjik Depression, a relatively young fold and thrust belt forming as part of the deformation related to the collision of India and Asia. Unlike most other reservoirs where induced seismicity has occurred, seismic activity near Nurek was monitored for 15 years prior to impounding. Thus, although the reservoir is in a region of relatively high natural seismicity, the influence of the reservoir can be clearly documented as a pronounced increased in seismicity in the immediate vicinity of the reservoir (Simpson and Negmatullaev, 1981). The largest induced earthquakes at Nurek were two events of magnitude 4½ which immediately followed the first filling to 100 m in 1972. A major swarm of activity was also related to the start of the second stage of filling, to 200 m, in 1976. The earthquakes at Nurek are shallow, confined to the sediments of the Tadjik Depression at depths of less than 8 km and occur mainly beneath or close to the main body of the reservoir. Although the predominant mode of tectonic deformation in the region is thrusting, various focal mechanisms (mainly thrust and strike-slip) are observed at Nurek, indicating a complex interaction of the reservoir with the local geological structure (Keith *et al.*, 1982).

The relationship between the level of seismicity and changes in water level is seen more clearly at Nurek than at most other reservoirs. Simpson and Negmatullaev (1981) show that the highest rates of seismicity at Nurek follow large negative values in the second derivative of water level—either abrupt decreases in the water level or sudden decreases in the rate of filling, even though the water level may continue to rise. They also show that these variations in filling rate are more likely to trigger activity if the water level is higher than it has been previously.

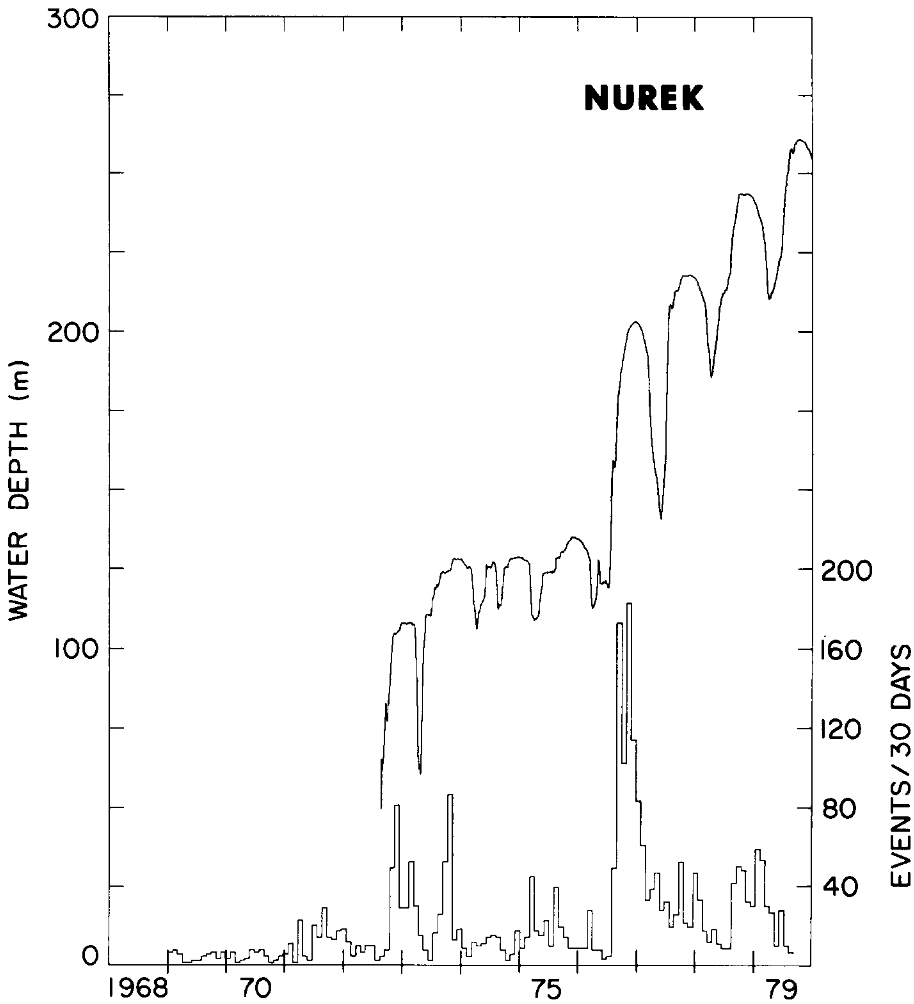

FIG. 3. Water level and number of earthquakes per 30 days at Nurek reservoir. Data from Simpson and Negmatullaev (1981).

Kariba. Kariba dam lies on the Zambezi River between Zimbabwe and Zambia. The concrete arch dam is 128 m high and impounds a reservoir with a maximum capacity of 160 km^3. Low level seismicity began shortly after the reservoir was impounded in 1958. The water level gradually increased in annual stages until approached maximum level in late 1963. Two earthquakes of magnitude near 6 occurred in September, 1963 shortly after the reservoir had first reached maximum depth (Figure 4). There are not sufficient local stations to determine the depths or focal mechanisms with great accuracy, but limited local data (Archer and Allen, 1969; Gough and Gough, 1970a) and analysis of teleseismic data (Sykes, 1967; Pavlin and Langston, 1983) indicate shallow focal depths (<10 km) and normal faulting mechanisms.

Gough and Gough (1970a,b) present a detailed analysis of the Kariba earthquakes up to mid-1968, and in a later paper (Gough and Gough, 1976) they compare the rate of activity for 1959 to 1968 with that from 1969 to 1974. They note that the correlation between the water level and the rate of seismicity, which was strong in

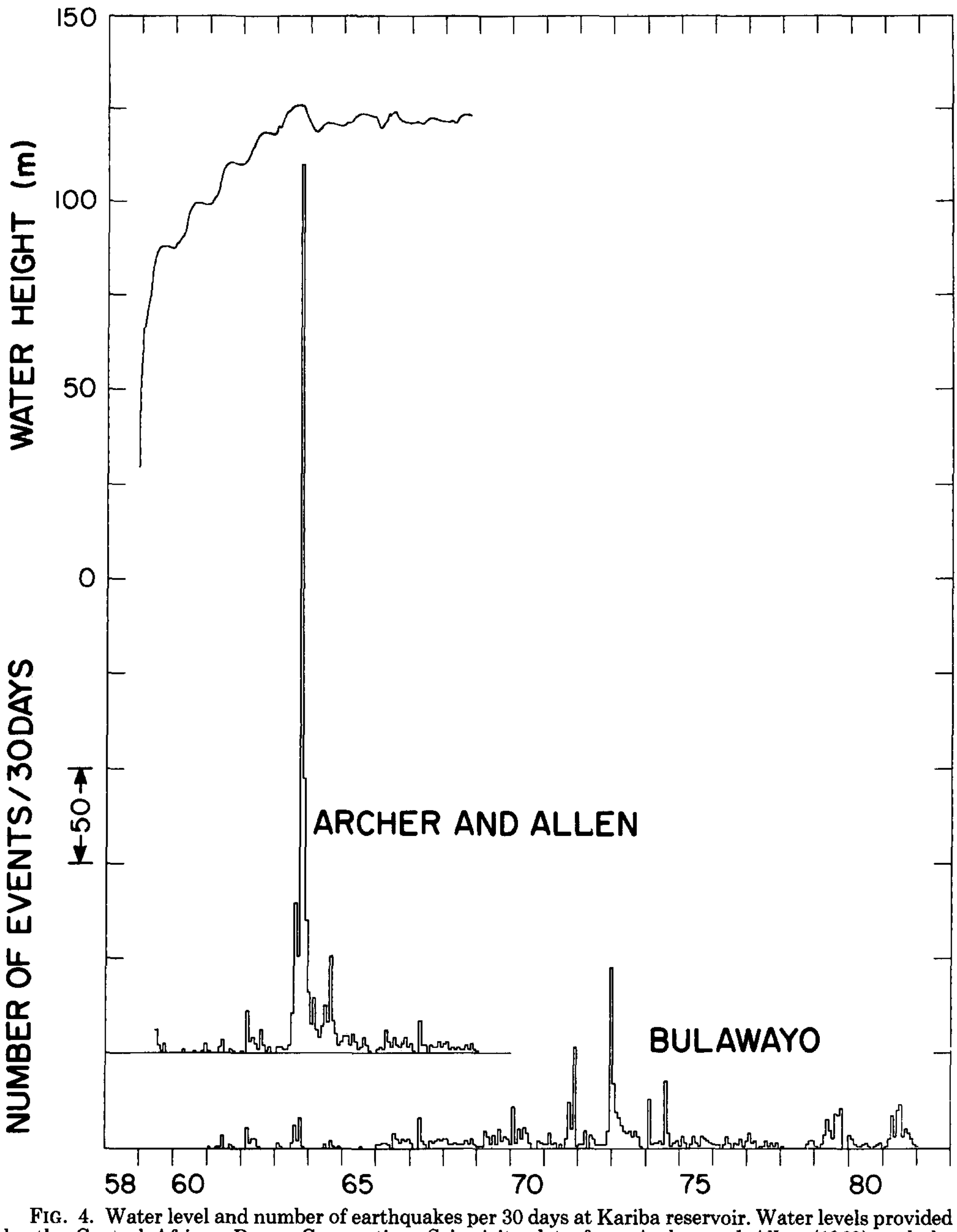

FIG. 4. Water level and number of earthquakes per 30 days at Kariba reservoir. Water levels provided by the Central African Power Corporation. Seismicity data from Archer and Allen (1969) and the Bulawayo catalog as reported in the International Seismological Center (ISC) earthquake catalog.

the early part of the sequence, became less obvious after 1966. Up to 1966, the activity at Kariba was directly beneath the reservoir, near the dam, and can be classified as primarily of the rapid response type, closely following changes in water level. After 1966, the activity migrated away from the dam (Simpson, 1975), and may be of the delayed response type, with a less obvious temporal correlation with water level.

Others. Some other examples of reservoirs where there is a clear indication of an abrupt initiation of seismicity soon after initial filling include Kremasta (Drakopoulos, 1974) and Talbingo (Timmel and Simpson, 1972).

Delayed response

Cases of delayed response contrast with rapid response in that the dominant seismicity occurs relatively late in the life of the reservoir (e.g., Aswan, Koyna and Oroville, Figure 1). The reservoir has often gone through a number of apparently similar annual cycles in water level change without any increased seismicity before the dominant seismicity occurs.

Koyna. Koyna dam impounds Shivaji Sagar Lake on the Koyna River in the Deccan Traps of western India. The dam is 103 m high and the reservoir has a capacity of 2.8 km^3. The historical record indicates a low level of natural background seismicity. Shortly after filling of the reservoir began in 1962, small earthquakes were felt in the area and the number and intensity increased after 1963 (Gupta and Rastogi, 1976). The water level in Koyna reservoir is controlled by the monsoon rains and shows a rapid filling in late summer with gradual withdrawal throughout the rest of the year. Annual changes in water level are 15 to 50 m. Because of the relatively small volume of the reservoir, the first impounding in 1962 quickly brought the water level to near its maximum, but the major episode of seismicity did not occur until after 4 years of similar cycles of annual changes in water level (Figure 5).

The largest of the Koyna earthquakes occurred in late 1967, with an event of magnitude 5.5 on 13 September 1967 and the main Koyna earthquake of magnitude 6.2 on 10 December 1967. The December 10 event is the largest known reservoir-induced earthquake and caused considerable damage to the dam and the town of Koynanagar, killing 200 and injuring over 1500.

Various fault plane solutions have been presented for the December 1967 Koyna earthquake (see Langston, 1976, for summary) most of which indicate almost purely strike-slip faulting. The best determinations of the focal depths for the mainshock and aftershocks give depths of less than 15 km with the most likely depth of the mainshock less than 5 km (Langston, 1976; Rastogi and Talwani, 1980). The reported locations concentrate in two main trends (Rastogi and Talwani, 1980). The most active trend, which includes the mainshock, is immediately downstream from the dam and covers a zone more than 10 km wide and 30 km long. Unfortunately, data from the local seismograph network do not provide the resolution necessary to determine whether this activity is concentrated on a single fault or a number of related structures. Within the location accuracy available, it is possible that most of these events are on a single fault extending south from the reservoir.

The seismicity at Koyna shows a tendency to correlate with periods of high water level and rapid rises in water level (Gupta, 1983), with the seismicity increasing each year following the rainy season. The December 1967 earthquake occurred at the time of the highest water level reached since the impoundment of the reservoir in 1962. Following the 1967 earthquake, the water level was kept well below the 1967 maximum until 1973. During this five year period, the earthquake activity remained relatively low. In August 1973 the reservoir level was allowed to increase beyond the 1967 maximum. Although the level exceeded that in 1967 by only 1 m, the high water in 1973 was associated with the highest level of seismicity observed during the period 1968 to 1973, and included an earthquake of magnitude 5.2 on 17 October 1973 (Figure 5). Simpson and Negmatullaev (1981) also note that seismicity at Nurek is more likely when the water level exceeds its previous maximum. These

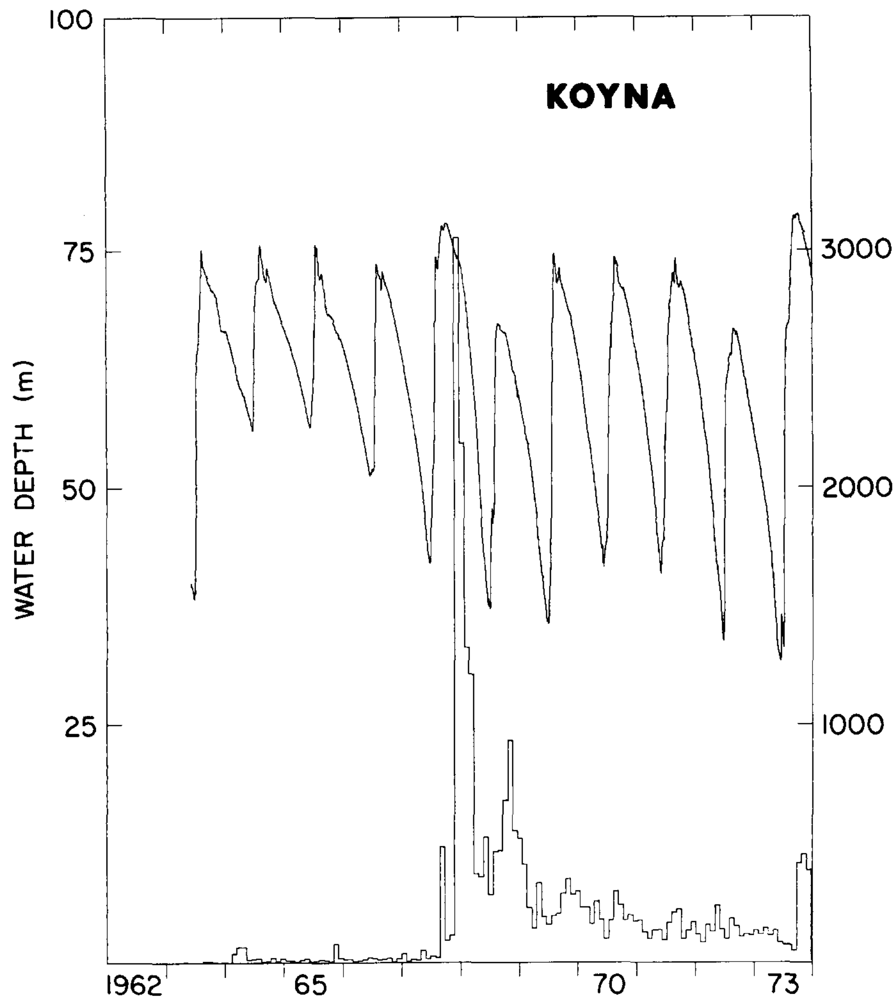

FIG. 5. Water level and number of earthquakes per 30 days at Koyna reservoir. Data provided by the Central Water and Power Research Station.

natural examples of memory of previous stress history are similar to effects observed in laboratory experiments on acoustic emission, where the necessity for the previous stress maximum to be exceeded before acoustic emission begins has been called the Kaiser effect (Kurita and Fujii, 1979).

Oroville. Oroville dam is located in the foothills of the Sierra Nevada, California. The earth fill dam is 235 m high and impounds a reservoir with maximum capacity of 4.3 km^3. A magnitude 5.7 earthquake occurred 12 km south (downstream) of the dam on 1 August 1975 (Morrison *et al.*, 1976). Aftershocks extended toward the dam and are on a westward-dipping normal fault (the Cleveland Hills fault) which crosses the reservoir upstream of the dam (Lahr *et al.*, 1976; Savage *et al.*, 1976). Focal depths extend to 15 km, with most of the activity above 10 km.

The Oroville earthquake of August 1975 occurred more than 7 years after the dam was impounded in 1968 (Figure 6). From 1969 (when the reservoir level first reached maximum) to 1974, the seasonal variations in water level were less than 25 m. During a drought in 1974 to 75, the water level dropped almost 50 m, and the

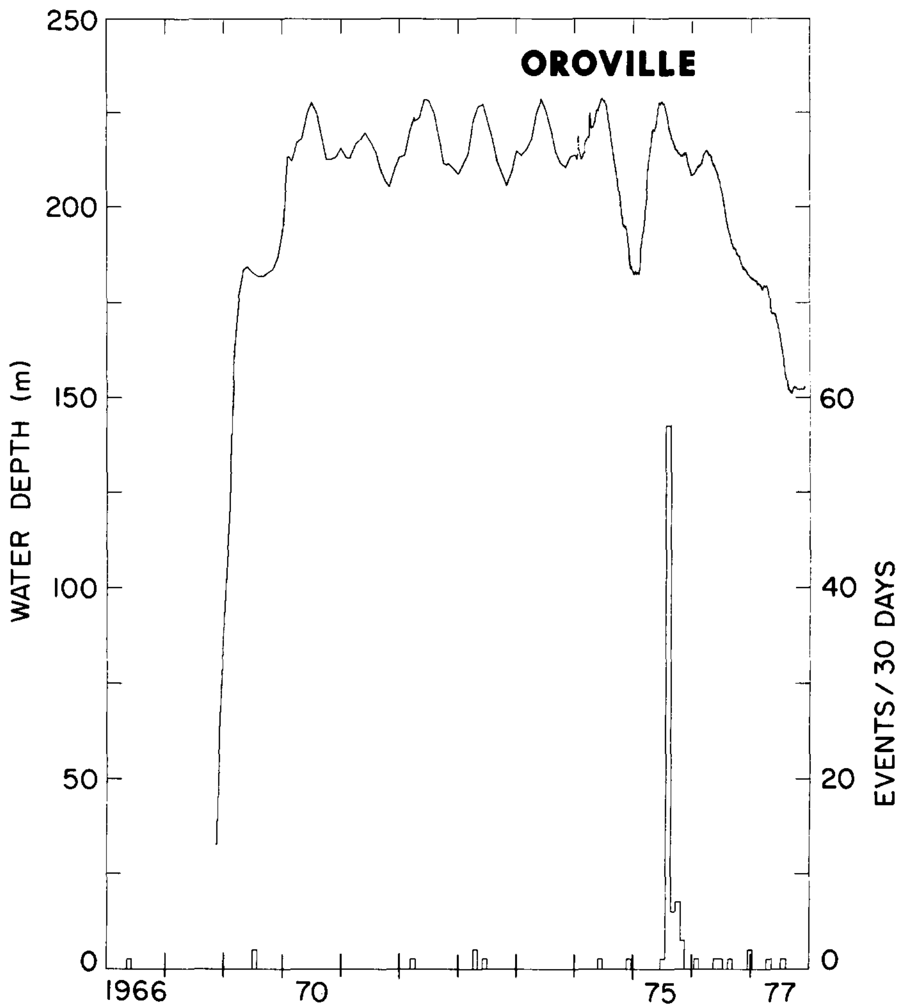

FIG. 6. Water level and number of earthquakes per 30 days at Oroville reservoir. Water levels provided by the California Division of Mines and Geology. Seismicity from the ISC earthquake catalog.

possible connection of the 1975 earthquake with this rapid change in water level has been discussed by Toppozada and Morrison (1982).

Aswan. The Aswan High Dam, on the River Nile in Egypt, impounds Lake Nasser, one of the largest reservoirs in the world. The dam is 110 m high and the impounded reservoir has a maximum volume of 160 km^3. Although the reservoir began to fill in 1964, it was not until 1975 that flooding reached the area above the epicenter of a magnitude 5.3 earthquake which occurred in November 1981 (Kebeasy *et al.*, 1987). The 1981 earthquake occurred just following the seasonal maximum in water level (Figure 7). Yearly peaks in water level, 1981 to 84, are also followed by increases in seismicity. The focal depths for earthquakes at Aswan are considerably deeper than for other cases of induced seismicity, with the main concentration of earthquakes at depths of 15 to 30 km and minor activity from 0 to 10 km.

Mixed response

Although at some sites it is possible to categorize the induced seismicity as belonging to one of these end member categories (e.g., Manic-3 and Monticello

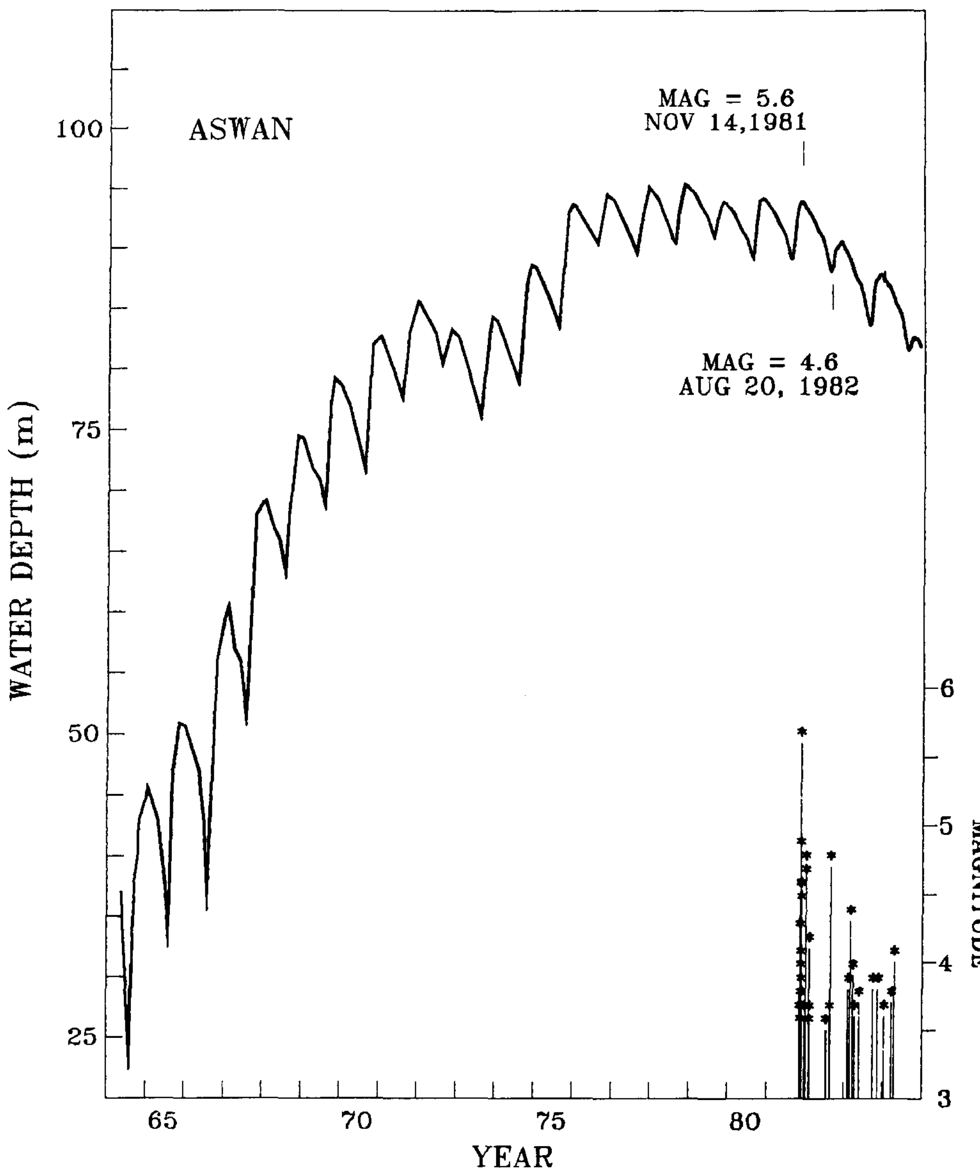

FIG. 7. Water level and magnitudes of earthquakes greater than magnitude 3 at Aswan reservoir. Data from Kebeasy *et al.* (1987).

appear to be cases of purely rapid response), both types of response may coexist at any one site. At Koyna and Lake Mead (Carder, 1945) low-magnitude seismicity was noticed soon after filling of the reservoir started, but the major burst of activity did not occur until after a number of annual cycles had passed. Some cases showing primarily a delayed response to initial filling also show a strong component of rapid response to shorter-term variations in water level. At Koyna (Gupta, 1983), Oroville (Toppozada and Morrison, 1982), and Aswan (Kebeasy *et al.*, 1987), all identified above as cases of delayed response, the timing of the highest levels of seismicity appears to be triggered by short-term changes, usually associated with the seasonal maximum in water level.

Other characteristics of rapid and delayed response

Where sufficient data exist to allow resolution of the spatial distribution of seismicity, cases of rapid or delayed response show a number of other common characteristics, in addition to the temporal similarities. Those sites showing rapid response tend to have seismicity that is shallow (≦10 km), low-magnitude, and concentrated directly beneath or near the edges of the reservoir (e.g., Nurek, Monticello, Manic-3). The seismicity at these sites is often spread over an active volume rather than being concentrated along well-defined fault planes.

In contrast, the cases of delayed response often have seismicity that is of larger magnitude, deeper (≧10 km), and sometimes extends to distances of 10 or more km from the reservoir. In some cases (Koyna, Aswan, and especially Oroville), there is evidence for the association of the activity with a well-defined fault zone passing through the reservoir.

The comparison of Nurek and Oroville is especially revealing since both reservoirs are large and of similar size (Figure 1), yet they contrast strongly in their response in seismicity. At Nurek, most of the seismicity appears to be on minor faults close to the edge of the reservoir. Larger faults exist in the Nurek area, but do not intersect the reservoir and have not shown increased activity (Keith *et al.*, 1982). In contrast, the main seismicity at Oroville was centered at a distance of more than 10 km from the reservoir, but on a fault that passed directly beneath the reservoir (Lahr *et al.*, 1976).

Discussion

Reservoir loading and changes in crustal strength. The major bursts in seismicity at the reservoirs described above fall into two obvious groups: those occurring soon after initial filling or after rapid changes in water level (e.g., at Manic-3, Monticello and Nurek); and those where the main seismicity occurs years after the first filling (e.g., at Oroville, Koyna, and Aswan).

The filling of a large reservoir can change the strength of the crust in a number of ways (Bell and Nur, 1978; Simpson, 1986; Figure 8)—through an increase in elastic stress due to the surface load, a concomitant instantaneous change in pore pressure proportional to the volumetric strain component, and increased pore pressure by diffusion from the surface. In cases where the water table prior to impoundment is low, flow of water from the reservoir into unsaturated pore space can also be important, as it also contributes to the total load applied to the crust (e.g., Aswan; Kebeasy *et al.*, 1987).

While the elastic and pore pressure effects are coupled in a complete formulation of the poro-elastic deformation process (Rice and Cleary, 1976), it is useful to consider observations of seismicity at reservoirs in terms of the two effects independently. In practice, such separation of the two effects may arise because of under-saturation and/or incompressible rock matrix (Figure 8) or for geometrical reasons—elastic effects may dominate in the highly stressed region directly beneath the reservoir; diffusion from the surface may be more important along fault zones that extend out of the stressed region away from the reservoir.

The first two effects—the increased elastic stress and the pore pressure increase that is coupled to it—are a result of the elastic deformation of the crust in response to the surface load, and they follow the variation of the surface load with little or no time delay. The latter two effects—diffusion of pore pressure and flow of water out of the reservoir—can lag considerably behind the surface load. We suggest that

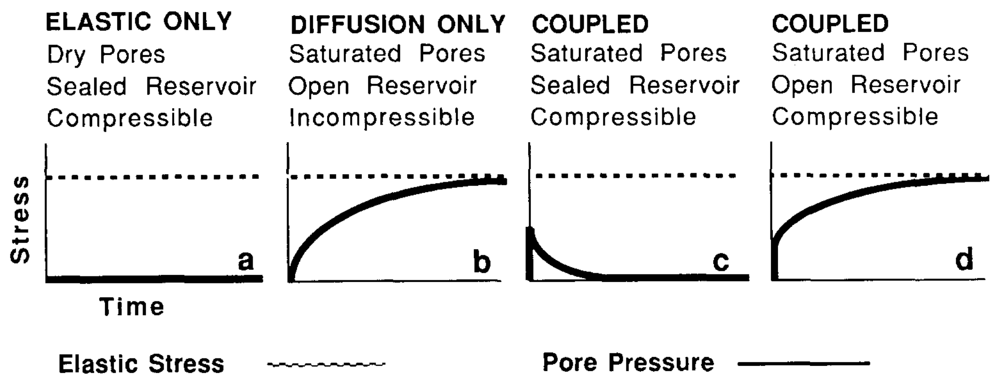

FIG. 8. Changes in elastic stress and pore pressure at a point beneath a reservoir produced by the sudden addition of a reservoir load. (a) The elastic effect in dry, compressible rock with no access of water from the reservoir to the rock beneath. No change in pore pressure. (b) The diffusion of water from the reservoir into a saturated, incompressible rock (no decrease in pore space). (c) The coupled elastic and pore pressure response of a saturated, compressible rock in which there is no access from the reservoir to the rock beneath. Initial compression causes a decrease in pore space and coupled rise in pore pressure, which decays in the steady state to zero. (d) The coupled response of a saturated, compressible rock in which water can flow from the reservoir into the rock beneath. The initial compression-induced rise in pore pressure is the same as in (c). Pore pressure continues to rise with the diffusion of pressure from the surface. Change in effective stress is the difference between the elastic stress and pore pressure. Change in strength will depend on the orientation of failure surfaces with respect to the preexisting tectonic stress (cf. Bell and Nur, 1978).

the rapid response in seismicity following the filling of some reservoirs is primarily controlled by strength changes of the first type related to elastic deformation and coupled pore pressure changes, whereas a delayed response is related to diffusion or flow from the reservoir.

If the rapid response in seismicity were due to an entirely elastic process, a complete and instantaneous correlation between water level and changes in seismicity would be expected. That this does not agree with observation indicates that in a coupled fluid-solid system, even those cases where elastic deformation dominates will be accompanied by a component of time-dependent fluid flow. In the same manner, it is unlikely that flow from a reservoir will act as a purely diffusion controlled process since some elastic deformation will accompany both the additional load of the reservoir and the increased pore pressure away from the reservoir.

While diffusion may play a role in both the rapid and delayed response, there is a fundamental difference in the distances over which pressure diffuses in the two cases. For the delayed response, in which we propose that simple diffusion dominates, the appropriate scale length for diffusion is the hypocentral distance from the reservoir to the earthquake activity. In cases of rapid response, the scale lengths are much shorter, with any diffusive component reflecting a local redistribution of pore pressure within the focal zone itself.

Changes in pore pressure and the hydraulic properties of the shallow crust. A number of studies have suggested that a diffusion-controlled process may be responsible for the $T{\sim}L^2$ relationship observed in time-dependent earthquake phenomena. Scholz *et al.* (1973) and Anderson and Whitcomb (1973) relate phenomena reported precursory to earthquakes to diffusion, and Nur and Booker (1972) suggest aftershocks may be controlled by pore pressure changes. Bell and Nur (1978), Keith *et al.* (1982), and Talwani and Acree (1985) have all attempted to estimate the hydraulic diffusivity of the crust from temporal associations of seismicity with water levels at reservoirs (either from delays between filling and seismicity as described here or from the spatio-temporal migration of induced seismicity). All point out, however, the difficulty in converting from a simple measure of squared distance and time to an actual estimate of hydraulic diffusivity. Without an independent measure of the magnitude of the pore pressure change necessary to trigger failure, it is possible only to place bounds on the crustal diffusivity (Keith *et al.*, 1982). The summary by Talwani and Acree (1985) suggests a value for what they term "seismic diffusivity" (slope of the L^2-T plot) of 10^4 to 10^5 cm^2/sec based on data from observations of induced seismicity. They argue that the "seismic diffusivity" determined in this way should be within one order of magnitude of the true hydraulic diffusivity. A "diffusivity" of the same magnitude is reported by Scholz *et al.* (1973) for precursory phenomena. For the cases of delayed response discussed here, which we interpret to be primarily diffusion-controlled, delays of 4 to 8 years are observed between first filling and major seismicity at hypocentral distances on the order of 10 to 30 km. This would give L^2/T values of 4 by 10^3 to 7 by 10^4 cm^2/sec, similar to those found by Talwani and Acree.

For cases of rapid response, it is much more difficult to identify which particular water level change associates with a particular change in seismicity. Most obvious are those where annual peaks or rapid changes in water level are quickly followed by bursts in seismicity (e.g., Simpson and Negmatullaev, 1981). However, even if it were possible to identify the time delay in these cases, it is still difficult to determine a precise value for diffusivity since the length scale is unknown—the appropriate distances now correspond to those within the focal region itself and not to the hypocentral distance from the reservoir to the earthquakes. This problem is avoided in those cases where a real growth of the focal zone is observed (Talwani and Acree, 1985) or there is migration of seismicity (e.g., Keith *et al.*, 1982).

Talwani and Acree (1985) and Talwani (1981) do not make the distinction we recognize between rapid and delayed response and they assume, in effect, that all cases of induced seismicity are the result only of diffusion. Many of the cases considered by Talwani (1981) of delays between filling and onset of seismicity are of the delayed response type and the time and length scales used are appropriate for a dominantly diffusion-controlled process. The case of Monticello reservoir (Figure 2) considered by Talwani and Acree (1985) presents an example of the difficulty in associating specific water level changes with changes in seismicity. They use the three week delay between the start of filling and the start of seismicity (see their Figure 4) to obtain an L^2/T value of 5.5 by 10^3 cm^2/sec for the diffusion to hypocentral depths of 1 to 2 km. We suggest that the pronounced increase in seismicity, which immediately follows the end of filling of the reservoir (Figure 2), is related to increased pore pressure at hypocentral depths, resulting from undrained elastic compression during the loading stage (rapid response). This mechanism does not require any diffusion from the reservoir but is related solely to processes within the hypocentral zone itself. If the time used in this case were the very short delay (<1 day?) between the end of filling and start of increased seismicity, the L^2/T value obtained would be two orders of magnitude higher than that given by Talwani and Acree (1985). If it is assumed that pure diffusion *from the surface* is the process involved, then this and other cases of rapid response, with delays of days and earthquake depths of up to 10 km, would imply L^2/T values on the order of 10^6 to 10^7 cm^2/sec corresponding to unreasonably high values for hydraulic diffusivity. The use of shorter length scales, more appropriate for processes occurring within the focal zone, lead to lower values of diffusivity. Assuming length scales within the focal zone of one tenth of the focal depths again leads to diffusivities of 10^4 to 10^5 cm^2/sec.

Values of hydraulic diffusivities determined from laboratory measurements for whole rocks, even at moderate effective stress, are orders of magnitude lower than the 10^4 to 10^5 cm^2/sec suggested by induced seismicity and show considerable variation between rock types (Rice and Cleary, 1976). The determination of such high diffusivities from observations of induced seismicity and other earthquake phenomena implies that fractures, and not whole rock properties, are the primary controlling factors in fluid flow throughout the crust. If hydraulic diffusivities derived from seismic observations vary by less than an order of magnitude, as suggested by Talwani and Acree (1985), then flow through fractures is also relatively independent of rock type and depth, at least in the shallow crust.

Conclusions

Two types of seismic response to the filling of large reservoirs result from two different mechanisms by which the addition of a reservoir at the surface can change the strength of the crust. Changes in seismicity which follow rapidly after the first filling of a reservoir, or follow substantial later changes in water level, are related to changes in elastic stress or changes in pore pressure coupled to the elastic stress. Such changes do not rely on the diffusion of water from the reservoir, and the length and time scales involved relate to the hypocentral region and not to the distance from the reservoir. Since the stress increase from the elastic load drops off rapidly with distance, seismicity in these cases is concentrated in the immediate vicinity of the reservoir and earthquake sizes tend to be small since the stress gradients are high.

In cases where there is a long delay between the filling of the reservoir and the start of increased seismicity, diffusion of pore pressure from the reservoir to hypocentral depths may play the dominant role. Faults intersecting the reservoir are usually present in these cases and may serve as conduits for flow away from the reservoir. This allows the influence of the reservoir to extend to greater distances, affecting longer fault segments and producing lower stress gradients, leading to larger magnitude earthquakes.

Since the development of a pore pressure system is a time-dependent process, temporal variations in seismicity near reservoirs provide an opportunity to determine hydraulic properties of the crust, such as diffusivity (Talwani and Acree, 1985). The delays between rapid changes in water level (or first filling) at a large reservoir and related increases in seismicity provide one measure of diffusion time. Before these delays can be interpreted in terms of hydraulic properties, care must be taken to identify the mechanism by which the pore pressure is developed. If it is by a local process of elastic compaction within the zone of influence of the reservoir, length scales appropriate for the focal zone must be used. If the pore pressure responsible for crustal weakening develops by diffusion from the reservoir at the surface, the appropriate length scales are the hypocentral distances from the reservoir. When cases of rapid and delayed response are identified as corresponding to these two mechanisms, respectively, both imply diffusivities on the order of 10^4 to 10^5 cm^2/sec. These diffusivities are higher than most estimates for whole rock, suggesting that fractures play a dominant role in controlling fluid flow throughout the crust (Witherspoon and Gale, 1977; Rice and Cleary, 1976).

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REFERENCES

Anderson, D. L. and J. H. Whitcomb (1973). Time dependent seismology, *J. Geophys. Res.* **80,** 1497–1503.

Archer, C. B. and N. J. Allen (1969). *A Catalogue of Earthquakes in the Lake Kariba Area, 1959–1968,* Meteorol. Serv. 35 pp.

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

Carder, D. S. (1945). Seismic investigations in the Boulder Dam area, 1940–1945, and the influence of reservoir loading on earthquake activity, *Bull. Seismol. Soc. Am.* **35,** 175–192.

Drakopoulos, J. (1974). Conditions and triggering mechanism of earthquake activity in the region of Kremasta-Kastraki dams, Greece *National Observatory of Athens,* Athens, Greece, (in Greek with English summary).

Gough, D. I. and W. I. Gough (1970a). Stress and deflection in the lithosphere near Lake Kariba, 1, *Geophys. J. R. Astro. Soc.* **21,** 65–78.

Gough, D. I. and W. I. Gough (1970b). Load-induced earthquakes at Lake Kariba, 2, *Geophys. J. R. Astro. Soc.* **21,** 79–101.

Gough, D. I. and W. I. Gough (1976). Time dependence and trigger mechanisms for the Kariba (Rhodesian) earthquakes, *Eng. Geol.* **10,** 211–217.

Gupta, H. K. (1983). Induced seismicity hazard mitigation through water level manipulation at Koyna, India: a suggestion, *Bull. Seismol. Soc. Am.* **73,** 679–682.

Gupta, H. K. and B. K. Rastogi (1976). *Dams and Earthquakes,* Elsevier, Amsterdam, 229 pp.

Kebeasy, R. M., M. Maamoun, E. Ibrahim, A. Megahed, D. W. Simpson, and W. S. Leith (1987). Earthquake studies at Aswan reservoir, *J. Geodynamics* **7,** 173–193.

Keith, C., D. W. Simpson, and O. V. Soboleva (1982). Induced seismicity and style of deformation at Nurek Reservoir, Tadjik SSR, *J. Geophys. Res.* **87,** 4609–4624.

Kurita, K. and N. Fujii (1979). Stress memory of crystalline rocks in acoustic emission, *Geophys. Res. Lett.* **6,** 9–12.

Lahr, K. M., J. C. Lahr, A. G. Lindh, C. G. Bufe, and F. W. Lester (1976). The August 1975 Oroville earthquake, *Bull. Seismol. Soc. Am.* **66,** 1085–1099.

Langston, C. (1976). Body wave synthesis for shallow earthquake sources: inversion for source and earth structure parameters, *Ph.D. Thesis,* California Institute of Technology, Pasadena, CA, 214 pp.

Leblanc, G. and F. Anglin (1978). Induced seismicity at the Manic-3 reservoir, Quebec, *Bull. Seismol. Soc. Am.* **68,** 1469–1485.

Morrison, P. W., Jr., B. W. Stump, and R. Uhrhammer (1976). The Oroville earthquake sequence of August 1975, *Bull. Seismol. Soc. Am.* **66,** 1065–1084.

Nur, A. and J. Booker (1972). Aftershocks caused by pore fluid flow, *Science* **175,** 885–887.

Pavlin, G. B. and C. A. Langston (1983). An integrated study of reservoir-induced seismicity and Landsat imagery of Lake Kariba, Africa, *Photogrammetric Eng. and Remote Sensing* **49,** 513–525.

Rastogi, B. K. and P. Talwani (1980). Relocation of Koyna earthquakes, *Bull. Seismol. Soc. Am.* **70,** 1849–1868.

Rice, J. R. and M. P. Cleary (1976). Some basic stress diffusion solutions for fluid-saturated elastic porous media with compressible constituents. *Rev. Geophys. Space Phys.* **14,** 227–241.

Roeloffs, E. A. (1988). Fault stability changes induced beneath a reservoir with cyclic changes in water level, *J. Geophys. Res.* **93,** 2107–2124.

Savage, W. U., D. Tocher, and P. C. Birkhahn (1976). A study of small aftershocks of the Oroville, California, earthquake sequence of August, 1975, *Eng. Geol.* **10,** 371–385.

Scholz, C. H., L. R. Sykes, and Y. P. Aggarwal (1973). Earthquake prediction: a physical basis, *Science* **181,** 803–810.

Simpson, D. W. (1976). Seismicity changes associated with reservoir impounding, *Eng. Geol.* **10,** 371–385.

Simpson, D. W. (1985). Induced seismicity at Kariba Reservoir—a re-examination, *EOS* **66,** 314.

Simpson, D. W. (1986). Triggered earthquakes, *Ann. Rev. Earth Planet. Sci.* **14,** 21–42.

Simpson, D. W. and S. K. Negmatullaev (1981). Induced seismicity at Nurek Reservoir, *Bull. Seismol. Soc. Am.* **71,** 1561–1586.

Snow, D. T. (1972). Geodynamics of seismic reservoirs, *Proc. Symp. Percolation through Fissured Rock*, Stuttgart: Ges. Erd- und Grundbau, **T2J:** 1–19.

Sykes, L. R. (1967). Mechanism of earthquakes and nature of faulting on the mid-oceanic ridges, *J. Geophys. Res.* **72,** 2131–2152.

Talwani, P. (1981). Hydraulic diffusivity and reservoir induced seismicity, Final technical report, U.S.G.S., Reston, Virginia, 48 pp.

Talwani, P. and S. Acree (1985). Pore-pressure diffusion and the mechanism of reservoir-induced seismicity, *Pageoph.* **122,** 947–965.

Timmel, K. E. and D. W. Simpson (1972). Seismic events during filling of Talbingo Reservoir, *ANCOLD (Australian National Committee on Large Dams) Bull.* **36,** 27–33.

Toppozada, T. R. and P. W. Morrison (1982). Earthquakes and lake levels at Oroville, California, *Calif. Geol.* **35,** 115–118.

Witherspoon, P. A. and J. E. Gale (1977). Mechanical and hydraulic properties of rocks related to induced seismicity, *Eng. Geol.* **11,** 23–55.

Zoback, M. D. and S. Hickman (1982). *In situ study* of the physical mechanisms controlling induced seismicity at Monticello reservoir, South Carolina, *J. Geophys. Res.* **87,** 6959–6974.

LAMONT-DOHERTY GEOLOGICAL OBSERVATORY
OF COLUMBIA UNIVERSITY
PALISADES, NEW YORK 10964

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