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Hydrologic Precursors to Earthquakes: A Review

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Abstracts—This review summarizes reports of anomalous flow rates or pressures of groundwater, oil, or gas that have been interpreted as earthquake precursors. Both increases and decreases of pressure and flow rate have been observed, at distances up to several hundred kilometers from the earthquake epicenter, with precursor times ranging from less than one day to more than one year. Although information that might rule out nontectonic causes does not appear in many published accounts of hydrologic anomalies, several recent studies have critically evaluated the possible influences of barometric pressure, rainfall, and groundwater or oil exploitation. Anomalies preceding the 1976 Tangshan, China, and the 1978 Izu-Oshima-Kinkai, Japan, earthquakes are especially well-documented and worthy of further examination.

Among hydrologic precursors, pressure head changes in confined subsurface reservoirs are those most amenable to quantitative interpretation in terms of crustal strain. The response of pressure head to earth tides determines coefficients of proportionality between pressure head and crustal strain. The same coefficients of proportionality should govern the fluid pressure response to any crustal strain field in which fluid flow in the reservoir is unimportant. Water level changes in response to independently recorded tectonic events, such as earthquakes and aseismic fault creep, provide evidence that a calibration based on response to earth tides may be applied to crustal strains of tectonic origin.

Several models of earthquake generation predict accelerating stable slip on part of the future rupture plane. If precursory slip has moment less than or equal to that of the impending earthquake, then the coseismic volume strain is an upper bound for precursory volume strain. Although crustal strain can be only crudely estimated from most reported pressure head anomalies, the sizes of many anomalies within 150 kilometers of earthquake epicenters appear consistent with this upper bound. In contrast, water level anomalies at greater epicentral distances appear to be larger than this bound by several orders of magnitude.

It is clear that water level monitoring can yield information about the earthquake generation process, but progress hinges on better documentation of the data.

Key words: Earthquake prediction; hydrologic precursors; water level.

Introduction

Changes in pressure, flow rate, color, taste, smell, and chemical composition of surface and subsurface water, oil, and gas have been termed hydrologic precursors to earthquakes. Only changes in fluid pressure or flow rate are considered in this review. Such changes have been reported before scores of earthquakes in many countries. Older reports describe 'macroscopic' phenomena that can be observed

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without special equipment, such as spouting of oil from a previously dry well. In the last 15 years, water level anomalies only a few centimeters in amplitude have been detected, using pressure transducers and digital filtering techniques. Although some scientists still doubt that a relationship exists between hydrologic anomalies and subsequent earthquakes, increasing effort is being invested in water level monitoring for earthquake prediction purposes.

This review is organized as follows. First, the theory describing how subsurface fluid pressure responds to crustal strain is outlined. Then, the observations supporting the theory are summarized. Although more evidence is needed, a working hypothesis proposes that water wells in which earth tidal fluctuations occur can be used as calibrated strainmeters. A compilation of the amplitudes of earth tidal fluctuations in wells is presented as the best currently available basis for estimating the strain sensitivity of wells.

After developing this quantitative framework for interpreting water level fluctuations, the types of measurements currently being made are listed, calling attention to successful predictions in which hydrologic anomalies have played a role. The subsequent section is an inventory of popular generalizations about hydrologic precursors: where they are best observed, their migration with time, and their form. Counterexamples to most of these generalizations are cited.

Hypotheses as to the mechanisms that may cause anomalies during the earthquake generation process include the development of a fractured volume near the hypocenter, the passage of a propagating deformation front, and aseismic slippage of part of the fault plane. The first two possibilities are briefly considered, and the third, aseismic slip, is more fully explored. Many of the anomalies that have been perceived could be explained by at least one of these mechanisms.

If water wells are to be useful for monitoring crustal strain or predicting earthquakes, then better documentation of the data is required. A list of suggested supporting information is included in the final section.

Water Level Changes due to Crustal Strain: Theory

Interpretation of hydrologic data in terms of crustal strain requires mathematical models for two effects: first, for the response of fluid pressure in the reservoir to crustal strain, and, second, for the response of the observed quantity (usually fluid level or fluid pressure in a borehole) to the changing pressure field in the reservoir.

Unless there is evidence to the contrary, it is usually assumed that the saturated rock of the fluid reservoir behaves as a porous elastic material. A general mathematical description of a porous elastic material was given by BIOT (1941), and conveniently reformulated by RICE and CLEARY (1976). Using the notation of RICE and CLEARY (1976), the porous elastic medium can be characterized by a diffusivity, c, drained and undrained Poisson's ratios, ν and ν_u, Skempton's coefficient, B, and a shear modulus, G. This formulation is quite general, in that it allows for compressibility of the grains in the rock material, and can describe the response of the reservoir to arbitrary changes in stress or strain.

Fluid pressure change produced by volume strain: Confined reservoir

When fluid flow can be neglected, the constitutive relations as presented by RICE and CLEARY (1976) give

$$\Delta p = -(2GB/3)[(1 + \nu_u)/(1 - 2\nu_u)]\Delta\varepsilon \tag{1}$$

where Δp is the change in reservoir fluid pressure and $\Delta\varepsilon$ is the increment of volumetric strain in the reservoir (expansion positive). For example, in a porous elastic reservoir with $G = 3\ Gpa$, $B = 0.8$, and $\nu_u = 0.3$, (1) predicts a pressure change of 52 cm of water per microstrain.

Earth tides or barometric disturbances of large spatial extent are usually expected to produce uniform volumetric strain in a subsurface fluid reservoir. In the absence of spatial pressure gradients, fluid flow in the reservoir can be neglected. The reservoir volume strain associated with earth tides is usually derived by assuming that the tides impose known horizontal strains with no change of vertical stress. Substitution of the resulting expression for the tidal volume strain into (1) gives the pressure change in the reservoir due to earth tides. BREDEHOEFT (1967) analysed the tidal response of a porous medium with incompressible grains. VAN DER KAMP and GALE (1983) studied the tidal response for a medium with compressible grains, and showed that it reduces to the result of BREDEHOEFT (1967) if the grains are assumed incompressible. BODVARSSON (1970) also gives a relationship between reservoir pressure and volume strain.

The principle, underlying the use of water level measurements to measure crustal strain is that the pressure change in a reservoir depends on the time- and space-varying strain field, but not on the nature of the strain source. Consequently, an estimate of the coefficient in (1) obtained by studying tidal water level fluctuations, will also govern the response to tectonic strain in a similar frequency range.

If strain is spatially nonuniform and is applied slowly, then horizontal flow may not be negligible. Vertical flow toward the water table will also occur, and may be significant if the strain rate is sufficiently low or if confinement is inadequate. In most cases where flow is not negligible, the change in fluid pressure will be smaller than that given by (1). In particular, strains estimated from water level observations using (1) will usually be lower bounds.

Provided the reservoir tapped by the well is adequately described as a porous elastic medium, the coefficient or proportionality in (1) can be estimated by determining the amplitude of a single tidal constituent for which the volume strain is known. However, if the fluid in the well is coming from a small number of fractures of permeability much greater than the bulk rock, it will probably be incorrect to assume that the amplitude of a tidal constituent can yield a coefficient of proportionality between fluid pressure and reservoir volume strain. Such a system will be more sensitive to compression normal to the fracture plane (HANSON and OWEN, 1982; BOWER, 1983). In principle, it is possible to distinguish whether a single fracture is dominating the well response, or whether a uniformly porous reservoir is an adequate model, by examining the amplitudes and phases of the M_2 and O_1 tidal constituents (BOWER, 1983).

Fluid pressure response produced by barometric pressure: Confined reservoir

Ascertaining the response of water level in a well to changes in barometric pressure is important because fluctuations of barometric pressure can produce water level changes that might be mistaken for precursors. Furthermore, the presence of diurnal and semidiurnal components of barometric origin complicates determination of the well response to earth tides. The barometric response of a confined porous elastic reservoir can be derived by assuming that the vertical stress changes by an amount equal to the change in barometric pressure, Δb, that horizontal strains vanish, and that there is no flow. An increase of barometric pressure, Δb, at the surface of the earth compresses the reservoir, an effect that by itself would cause fluid pressure to rise by an amount

$$\Delta p = (B/3)[(1 + \nu_u)/(1 - \nu_u)]\Delta b. \tag{2}$$

However, in an open well, an increase of barometric pressure also exerts a downward force on the fluid surface, so the net effect of an increase in barometric pressure is to cause a drop, Δh, in water level by an amount

$$\Delta h = -(1/\rho g)[1 - (B/3)(1 + \nu_u)/(1 - \nu_u)]\Delta b \tag{3}$$

where ρ is fluid density and g is gravitational acceleration. For the same set of material properties used in the previous section, equation (3) predicts 0.52 cm of water level drop per 1 millibar rise in barometric pressure.

Unconfined reservoirs

Volume strain in an unconfined reservoir produces much smaller water level variations than in a well tapping a confined reservoir. If a volume strain, $\Delta\varepsilon$, is applied to an unconfined layer of saturated thickness H, then the water table will rise by an amount

$$\Delta h = -(H/n)\Delta\varepsilon \tag{4}$$

where n is porosity (BREDEHOEFT, 1967). For a 100 m thick saturated interval of rock with 2% porosity, (3) predicts 0.5 cm/microstrain water table rise, compared with 52 cm/microstrain for the confined example above. Clearly, a confined reservoir should be more sensitive to crustal strain.

Frequency dependence of water level response to reservoir pressure changes

Measurements of pressure in a packed-off section of well should closely track reservoir fluid pressure. However, most measurements are of water levels in open wells. For sufficiently slow changes in reservoir pressure, fluid level tracks reservoir pressure perfectly as

$$\Delta h = (1/\rho g)\Delta p. \tag{5}$$

However, ability of well water level to track changes of pressure in the reservoir depends on the rate of change, because any change of water level in the well requires the flow of water into or out of the well. The frequency dependence is, in principle, the same regardless of the source of reservoir pressure change. Response of water level to reservoir pressure change is influenced by the diffusivity and compressibility (or, equivalently, transmissivity and storage coefficient) of the reservoir. The radius of the well casing and the surface area of the reservoir is open to the well also enter the expression, since less water must flow into a narrow well to equilibrate its water level with the surrounding reservoir.

The dynamic response of a well-aquifer system was first analysed by COOPER *et al.* (1965) for the purpose of studying water level fluctuations produced by seismic waves; their analysis, which includes the effects of inertia on the motion of the water column, treats the well as a line fluid source in an infinite porous elastic reservoir. BREDEHOEFT (1967) illustrates the nature of the well frequency response for earth tidal periods, for which inertia is insignificant. HSIEH *et al.* (1987) have extended the work of COOPER *et al.* (1965) to show that the effect of a finite well radius is negligible, and to allow the well casing to have a different radius from the section of the well that is open to the reservoir. A similar analysis was carried out by BODVARSSON (1970), who derived an expression for the response of water level in a well in a confined reservoir to harmonically varying reservoir pressure by modeling the reservoir as a sphere, and the well as a spherical cavity. BODVARSSON (1970) showed that water level response falls off rapidly for fluid pressure changes having frequencies greater than a critical frequency. That critical frequency can range from cycles per minute to cycles per day for sandstones, down to less than a cycle per month for relatively intact crystalline rock.

Water Level Changes due to Crustal Strain: Observations

The theory described in the previous section is a basis for estimating precursory strain from precursory water level changes. To have confidence in these estimates, it must be shown that the theory successfully accounts for water level changes produced by known strains. Earthquakes and episodic fault creep are relatively well-understood tectonic sources of crustal strain, and the response of water levels to these strain sources offers some evidence that water wells can respond to tectonic strain as predicted by theory. Barometric fluctuations in water level also appear to be in approximate agreement with theory.

Water level changes associated with earthquakes and creep events

The North Kettleman Hills earthquake of 4 August 1985 ($M_L = 5.5$) produced coseismic water level drops in four wells near Parkfield, California, approximately 35 km from the epicenter (ROELOFFS and BREDEHOEFT, 1985). The drops of 2.1 to 7.8 cm, after correction for the tidal sensitivity at each well, agreed to within 50% with calculated values of elastic strain produced by the Kettleman Hills earthquake. Furthermore, the time dependence of each drop agreed with the theoretical step response of a flat confined aquifer having the frequency response calculated by HSIEH *et al.* (1987), for plausible values of transmissivity and storativity.

WAKITA (1975) studied coseismic water level changes associated with the 9 May 1974 Izu-Hanto-Oki, Japan, earthquake ($M_s = 6.8$), and found that the pattern of rising and falling water levels approximated the quadrantal distribution of volume strain expected from the earthquake focal mechanism.

Unfortunately, these two examples, in which coseismic water level changes have been successfully reconciled with estimates of coseismic strain, are the exception rather than the rule. Numerous reports document coseismic water level changes that are much too large, or in the wrong direction, to be explained in this way (e.g., VORHIS, 1968). The possibility of unexpectedly large coseismic water level changes suggests that even small foreshocks should be evaluated as potential causes of water level anomalies. Furthermore, coseismic water level changes by themselves should not be used to calibrate the strain sensitivity of water wells.

There are also several reports of water level fluctuations believed to be associated with episodes of fault creep (MERIFIELD and LAMAR, 1985; LIPPINCOTT *et al.*, 1985); all of them are from the state of California, USA. However, only at one location have propagating creep events been independently recorded by a creepmeter and a water well: several events were recorded in a well near the town of Hollister, California, located 10 m from the trace of the San Andreas fault, and 300 m from the nearest creep meter (JOHNSON, 1973; JOHNSON *et al.*, 1973). Three subsequent studies (WESSON, 1981; ROELOFFS and RUDNICKI, 1985, 1986) have shown that the sizes of the water level changes associated with an event in December, 1971, can be reconciled with a response of a well tapping a reservoir with plausible material properties. However, the persistent nature of the water level drop is not accounted for by the model of a steadily propagating creep event.

The few well-documented examples of water level changes in response to known crustal strains are valuable, but more work is needed in this area.

Effect of barometric pressure

The simplest approach to removing the effects of barometric pressure from water level data, in order to search for signals of tectonic origin, is to use a least-squares procedure to determine a single coefficient of proportionality between water level and barometric pressure, which is presumably an estimate of the coefficient in equation (3). Figure 2 shows water level data from the Turkey Flat well at Parkfield. In the filtered record, barometric pressure effects have been removed, using a single coefficient of proportionality. (Earth tides have also been removed by subtracting constituents determined by least-squares fits to each of eight tidal frequencies.) This

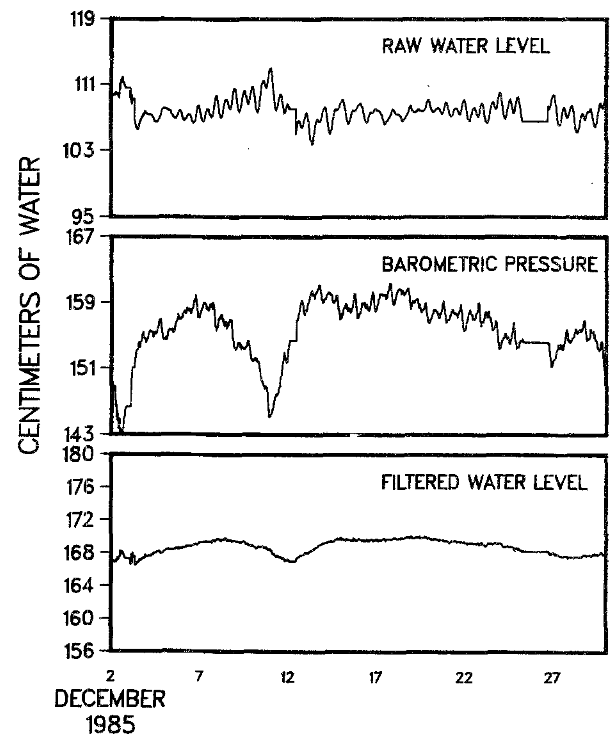

Figure 2
Barometric pressure, and raw and filtered water level data, for the well at Turkey Flat, near Parkfield, California, during the month of December, 1985.

procedure, although fairly effective, has overcompensated for the water level rises that correlate with barometric lows on December 2 and 11.

It is possible that the response of water level to barometric pressure is better matched by convolving the barometric pressure record with a number of filtering coefficients. This approach has been used by JOHNSON *et al.* (1973), and by KISSIN *et al.* (1983). However, these studies have not discussed the effectiveness of this procedure, nor have they reconciled the transfer functions they used with the theoretical frequency responses mentioned above.

Some data are presented with corrections for barometric pressure already applied (e.g., KOVACH *et al.*, 1975; KISSIN *et al.*, 1983). However, a barometric record should be provided with the water level data, even when an attempt has been made to filter out the barometric effects. Where no barometric record is available, the following facts should be born in mind. Barometric efficiencies are theoretically in the range from 0 to 1. The largest barometric pressure variations routinely observed at Parkfield, for example, are approximately 20 millibars, or, changing pressure units, about 20 centimeters of water. Typically, barometric pressure variations have periods on the order of several days. Consequently, if water level falls several centimeters and recovers over a period of several days, barometric pressure must be evaluated as a possible cause. On the other hand, a larger or more persistent water level change cannot reasonably be ascribed to barometric pressure.

Effects of rainfall

The response of reservoir pressure to rainfall is often delayed by a length of time that depends on the thickness and permeability of the overburden and on the distance to the reservoir outcrop. This delay can be as long as several months (e.g., LIPPINCOTT *et al.*, 1985). A threshold amount of rainfall may be required in order to initiate reservoir recharge; thus the absence of recharge in response to the first rainfall in a season is not evidence that recharge will not occur after later precipitation. Figure 3 shows data from the Flinge Flat well near Parkfield. A delay of about ten days exists between rainfall and water level rise. In addition to the delayed rise in water level due to reservoir recharge, short-period fluctuations in water level, of unknown mechanism, occur simultaneously with rainfall events. Small rises in water level at the time of precipitation have also been observed by other investigators and can sometimes be attributed to strain changes produced by surface loading (S. ROJSTACZER, personal communication).

Several years of rainfall and water level records at a well help to distinguish possible tectonic anomalies from changes produced by reservoir recharge. Lack of precipitation at the time of the anomaly is not sufficient to rule out recharge as a possible cause of a water level rise. Although a long record is required, measurements recorded only once per day are adequate to study the recharge behavior of most reservoirs.

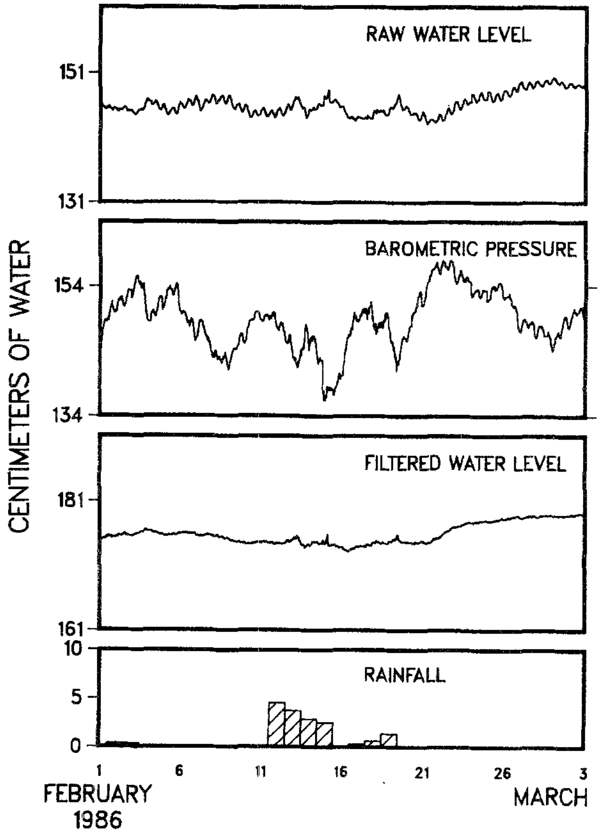

Figure 3
Raw and filtered water level, barometric pressure, and rainfall data for the Flinge Flat water well near Parkfield, California, for the month of February, 1986.

Approximate Strain Sensitivity of Water Wells

At the present time, no analysis of earth tidal fluctuations has been performed for the same well in which a precursor was observed. In lieu of such analyses, this section attempts to provide an approximate relationship for expressing water level changes in terms of strain. The relationship is based on published determinations of the amplitude of the M_2 tidal constituent in well tides.

Measured amplitudes of the M_2 tidal constituent, corrected to the equator, are shown in Figure 1 as a function of well depth; the points are listed in Table 1. Published determinations of the amplitude of the M_2 tide were not included if the period of observation was less than 28 days, if the reservoir was described as unconfined, or if the amplitudes of other tidal constituents were reported but were not in approximately the correct ratios to the amplitude of the M_2 constituent. Figure 1 suggests that deeper wells are more sensitive to tidal strain, and that crude

Table 1

Locations of wells for which M_2 amplitude is plotted in Figure 1. All amplitudes listed in the table have been reduced to the Equator

Location	*Latitude, degrees*	*Depth, meters*	M_2, cm	Reference
Kiabukwa, Zaire	−7.78	2400	7.67	MELCHIOR (1983)
Carlsbad, New Mexico, USA	32.30	86	0.63	
Iowa City, Iowa, USA	41.65	252	2.06	
Duchov, Czechoslovakia	50.62	600	3.45	
Turnhout, Belgium	51.32	2175	3.79	
Heibaart, Belgium	51.38	1660	2.93	
Gaocun, Tianjin, China		3402	8.13	LIU & ZHENG
Fanxian, Fanxian, China		2267	2.73	(1985)
Wali, Beijing, China		657	4.28	(see note 1)
Gold Hill, California, USA	35.83	88	0.78	ROELOFFS
Flinge Flat, California, USA	35.93	122	0.37	& BREDEHOEFT,
Turkey Flat, California, USA	35.89	177	0.93	unpublished
Crystallaire, California, USA	34.47	853	6.87	BREDEHOEFT *et al* (1986)
Chalk River 3, Ontario, Canada	46.01	161	4.29	BOWER (1983)
Chalk River 8, Ontario, Canada	46.01	300	3.52	
Marysville, Montana, USA	46.50	2000	4.85	NARASIMHAN *et al.* (1984)
CB136, Virginia, USA	37.13	136	2.14	RHOADS & ROBINSON (1979)

Note. 1. Tidal amplification is stated in this reference and was used to calculate the amplitude of the M_2 constituent.

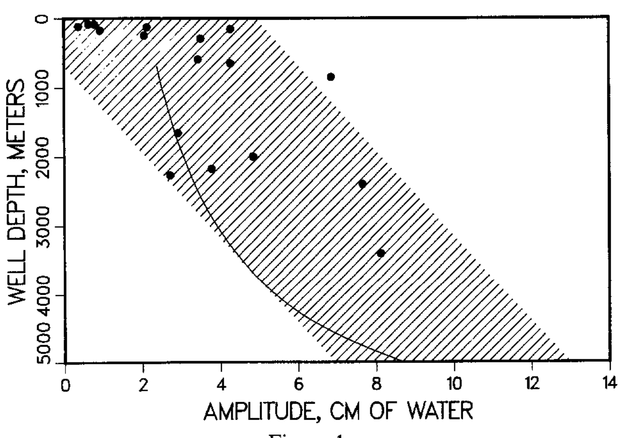

Figure 1

Amplitude of M_2 tidal constituent versus well depth for the wells listed in Table 1. The shaded region represents the relationship given in equation (6). Curve shows estimates of M_2 amplitude as a function of depth obtained from equation (7) and a porosity-depth relation for sandstones given by MAGARA (1980).

bounds can be placed on the amount of crustal strain that is required to produce a given change in water level in a well of given depth. Most points on the plot are within the shaded region defined by the bounding lines

$$\Delta h/\Delta\varepsilon = -(77d + 90 \pm 142) \text{ cm/microstrain} \tag{6}$$

where d is depth in kilometers.

BREDEHOEFT (1967) showed that tidal response should be approximately inversely proportional to porosity when the rock grain compressibility is negligible, compared to the compressibility of the reservoir. This approximation, which is not valid for very low porosities, gives

$$\Delta h = -\Delta\varepsilon\, K/n\rho g \tag{7}$$

in which K denotes the bulk modulus of water. Equation (7) can be used with a porosity versus depth relation to show that an increase of tidal sensitivity with depth is plausible. For example, MAGARA (1980) cites data for sandstones that show porosity decreasing by 4.53% per kilometer of depth, from a value of 27% at 750 meters. The curve in Figure 1 was obtained by substituting that relation into equation (7). Evaluation of equation (7), using a porosity-depth relationship for shales, results in an increase of sensitivity with depth more rapid than that observed. The reason is probably that the assumption of negligible grain compressibility becomes poorer as porosity decreases.

It is reasonable to assume that fluid level or pressure in a borehole responds to tectonic strain in the same way that it responds to tidal strain. This working hypothesis is adopted for the remainder of this review.

The Search for Hydrologic Precursors

Water levels are being monitored to detect earthquake precursors in Japan, the People's Republic of China, the Soviet Union, and the United States. Smaller programs are underway in Egypt and India. The observation networks are briefly described in the following paragraphs, and successful predictions are highlighted. Additional reports are summarized chronologically in the Appendix.

Japan

Although water level monitoring was not initially included in the Japanese earthquake prediction program, water levels are now monitored in 11 deep wells as part of the effort to make a short-term prediction of the expected Tokai earthquake (EARTHQUAKE ENGINEERING RESEARCH INSTITUTE, 1984).

The best documented occurrences of precursory water level changes in Japan preceded the 14 January 1978 Izu-Oshima-Kinkai earthquake ($M_S = 6.8$) (WAKITA, 1978; OKI and HIRAGA, 1979, 1987), although no short-term prediction was issued for this event. These water level anomalies were recorded in three deep observation wells, as well as at over 30 shallower wells monitored by a largely volunteer organization known as the Catfish Club. In addition to the changes in water level, there appear to have been precursory changes in water temperature and radon concentration. Anomalous deformation was also recorded by borehole volume strainmeters and surface geodetic measurements. The hydrologic anomalies preceding this earthquake are being re-examined in an effort to assess the effects of groundwater pumping and to establish the uniqueness of the anomalies in the long-term observation records (SHIMAZAKI, 1987).

People's Republic of China

Hydrologic observations constitute an important part of the earthquake prediction effort in the People's Republic of China. Observation wells in China are concentrated in the eastern half of the country, where the population density and, consequently, the seismic risk, are greatest. Four hundred wells are monitored in the Beijing-Tianjin-Tangshan area, most of which are shallow, but about 30 of which tap confined groundwater, are relatively deep, and have long observation histories of at least daily water level observations (WANG *et al.*, 1979). Many of these wells were specially drilled for the earthquake prediction program, and abandoned oil exploration wells are also utilized. Comparable numbers of wells are monitored in Sichuan and Yunnan provinces in southern China.

Among the numerous accounts of hydrologic precursors in China, three cases stand out. The 4 February 1975 Haicheng earthquake was preceded by hundreds of groundwater anomalies, which contributed to the successful short-term prediction of that event (DENG *et al.*, 1981). Macroscopic groundwater anomalies played a role in successfully predicting the 1976 Songpan-Pingwu earthquake, in spite of the absence of foreshocks (WALLACE and TENG, 1980). The best documented hydrologic precursors in China were those that preceded the disastrous 28 July 1976 Tangshan earthquake, for which no short-term prediction was issued. These anomalies have been carefully examined by WANG *et al.* (1979) and by WU *et al.* (1984), and while some of them appear to have been influenced by rainfall or by groundwater exploitation, others do not have simple explanations unrelated to the earthquake. For the Tangshan earthquake, several other types of data are available: coseismic water level changes, credible precursory resistivity changes (MADDEN, 1987), and pre- and postseismic leveling data (ZHANG, 1982).

USSR

In the Soviet Union, earthquake prediction observations are clustered in 'polygons', in which a variety of different measurements are made. There are at least eight of these polygons, located in the seismically active regions of the Causasus, Central Asia, Kamchatka, and the Kuril Islands. Water level or flow measurements are made in all these regions (SIDORENKO *et al.*, 1979).

The time and magnitude of an earthquake in the Pamir Mountains on 1 November 1978 were successfully predicted on the basis of many kinds of anomalies, including groundwater anomalies in the Fergana Valley polygon, Uzbek SSR (SIMPSON, 1979; ASIMOV *et al.*, 1979).

United States

Currently, the most intensive water level monitoring program in the United States is a part of the Parkfield, California earthquake prediction experiment being carried out under the auspices of the U.S. Geological Survey (BAKUN and LINDH, 1985). It appears that nearly identical earthquakes of magnitude about 6 have occurred on the San Andreas fault near Parkfield approximately every 22 years since 1881. In anticipation that the next characteristic Parkfield earthquake will take place within a few years, several types of monitoring have been undertaken in the Parkfield area. Water levels are currently recorded at 7 wells that range in depth from 100 to 200 meters. Pressure head in each well, together with barometric pressure and rainfall, are recorded every 15 minutes and telemetered via satellite to the U.S. Geological Survey laboratory in Menlo Park, California. At several sites, a second, shallower well is monitored in order to study the relationship between the deeper water pressure and the water table. Other crustal deformation sensors operating at Parkfield include creepmeters, tiltmeters, borehole strainmeters, a two-color laser geodimeter network, and two leveling lines. Although the Parkfield experiment is yielding valuable information about the behavior of water wells as strainmeters, no precursory anomalies have yet been identified.

Several other studies of water levels in California have been made for earthquake prediction purposes. Lamar-Merifield Geologists, Inc. collected water level data in southern California beginning in 1976 (LAMAR and MERIFIELD, 1979; MERIFIELD and LAMAR, 1978, 1980, 1981; MERIFIELD, LAMAR, and BEAN, 1982, 1983). Over thirty wells were eventually monitored. Several were equipped with a telemetry system that transmitted data to a central observatory once per day; water levels in these wells were sampled once a minute. Other water levels were measured by volunteers, at intervals ranging from once a day to once a month. MERIFIELD and LAMAR (1985) decribe anomalous water level changes in several wells before two earthquakes, on 25 February 1980 ($M_L = 5.5$) and on 22 March 1982 ($M_L = 4.5$).

In another experiment, a 152 m deep well in the Cienega Winery near Hollister was monitored from 1971 thru 1975 (KOVACH *et al.*, 1975). Small water level minima were observed in this well before three nearby earthquakes ($4.0 < M_L < 5.0$)

Three wells in the Mojave Desert, southern California, have intermittently been monitored by U.S. Geological Survey personnel since 1981 (HEALY and URBAN, 1985).

Since 1984, the U.S. Geological Survey has been monitoring three wells with depths from 80 to 150 m in the Long Valley Caldera, with the goal of detecting water level changes related to changes of deformation rate. Water level data are recorded at time intervals ranging from 15 to 30 minutes (S. ROJSTACZER, personal communication).

In South Carolina, a 100 m deep observation well in the vicinity of Lake Jocassee has been monitored for at least one year. Anomalies that were possible precursors to earthquakes having magnitudes between 1.0 and 2.3 were detected (TALWANI and VAN NIEUWENHUISE, 1980).

Other countries

Several wells have been monitored since 1985 in a search for precursors to earthquakes near the Aswan Dam in Egypt (D. W. SIMPSON, personal communication).

In Shillong, India, water levels in several dug wells and discharges from two springs have been monitored once per day since 1980. Hydrologic anomalies that occurred before four earthquakes in March thru July, 1977 were interpreted as precursors (NAYAK *et al.*, 1983). GUHA (1984) discusses the influence of rainfall on these anomalies.

Frequently Reported Patterns

Observers of hydrologic anomalies have suggested a number of features they believe to be characteristic of these precursors. Next, these features are summarized, and conclusions drawn by different investigators are contrasted.

Distance from epicenter

A remarkable aspect of many of the precursory groundwater anomalies listed in the Appendix is that they were observed at distances of several hundred kilometers from the earthquake epicenter. MONAKHOV *et al.* (1983) and SADOVSKY *et al.* (1979) state that small 'rebound-type' anomalies can be observed up to 100 km away for events of magnitude less than 4, and up to 900 to 1000 km away for events of magnitude 7 or 8. Before the 1976 Tangshan earthquake (magnitude 7.8), groundwater anomalies were observed as much as 200 km from the epicenter (WANG *et al.*, 1979). Anomalies at distances between 250 and 360 km from the epicenter were reported for the 1982 Lulong ($M_S = 5.3$) and 1983 Heze ($M - 5.9$) earthquakes in China (WANG and WU, 1986). Groundwater anomalies 150 km from the epicenter have been reported for three successfully predicted earthquakes: 1975 Haicheng, China (magnitude 7.3) (DENG *et al.*, 1981), 1976 Songpan-Pingwu, China (magnitude

7.2) (WALLACE and TENG, 1980), and 1978 Pamir Mountains, USSR (magnitude 7) (SIMPSON, 1979).

Sensitive locations

Wells drilled into fault zones and, ideally, at intersections of two or more fault zones, may be especially sensitive to premonitory groundwater variations (e.g., ZHU and ZHONG, 1979; WANG and WU, 1986). For example, before the 1976 Songpan-Pingwu earthquakes, groundwater and other 'macroscopic' anomalies appeared to approach the epicentral area along the Longmenshan fracture zone (WALLACE and TENG, 1980). Some concentration of groundwater anomalies along active faults was reported prior to the 1975 Haicheng earthquake (DENG *et al.*, 1981). The more distant groundwater anomalies preceding the 1976 Tangshan earthquake were observed along faults (WANG *et al.*, 1979). In the United States, precursory water level anomalies reported by KOVACH *et al.* (1975) were observed in a well drilled into the San Andreas fault zone.

There are at least two possible reasons why wells drilled into fault zones may be more sensitive to precursory anomalies. First, the preseismic strain field may be concentrated near the fault, either because it is directly involved in generating the impending earthquake, or because it slips in the intensifying stress field around the future epicenter. On the other hand, fault zone materials may respond with greater water pressure changes to any source of strain, perhaps because the porosity of fractured materials is especially sensitive to changes in the stress field. The first possibility could be distinguished from the second by studying the response of sensitive wells in fault zones to earth tides.

Sense of precursory water level changes

Although both rising and falling water levels have been interpreted as precursors (see Appendix), many authors have expressed the view that falling water levels predominate. According to MONAKHOV *et al.* (1983) and SADOVSKY *et al.* (1979), the most common precursor is a water level drop of several centimeters several days before the earthquake. Anomalies having the same form have also been reported to occur over longer time scales (e.g., KOVACH, 1975). Typically, the drop is beginning to recover when the earthquake occurs; this type of anomaly has been referred to as a 'rebound' anomaly. WANG (1985) also concludes that water levels commonly drop before earthquakes, with rebounds frequent near the epicenter. These water level drops are believed to be related to increase of porosity and permeability due to fracturing, with the subsequent recovery attributable either to influx of fluid or to compression (e.g., MONAKHOV *et al.*, 1983; SADOVSKY *et al.*, 1979; WANG *et al.*, 1979).

Nonetheless, there are also many reports of rising water levels that have been interpreted as precursors. DENG *et al.* (1981) state that many more rising than falling water levels were observed before the 1975 Haicheng, China earthquake. Both rising and falling water levels would be expected if the anomalies represent the response to volumetric strain produced by fault slip. The possibility that aseismic precursory slip is the source of water level anomalies is discussed more fully below.

Time migration

Researchers differ in their opinions as to where groundwater anomalies ought to first be observed. WANG and LI (1979) reviewed water level anomalies before the Haicheng, Tangshan, and Songpan-Pingwu earthquakes. They concluded that anomalies first appeared at large distances from the epicenter, and then migrated toward the epicenter, particularly along the direction of future fault rupture, or along fault lines parallel to that direction. ZHU and ZHONG (1979) discuss this pattern in detail for the 1975 Haicheng, China earthquake, although DENG *et al.* (1981) perceived a more complex time-space pattern in the same set of anomalies. For both the Haicheng and Songpan-Pingwu earthquakes, the migration was of macroscopic anomalies and occurred over a distance of 150 to 200 km and a time period of about 3 months. The explanation for the converging of anomalies toward the epicenter is the gradual concentration of stress in the epicentral area, which is believed to be the last to fail because it is a region of relatively high strength (WANG and LI, 1979).

On the other hand, SADOVSKY *et al.* (1979) believe that water level anomalies begin sooner, and are larger, close to the epicentral area than further away. Because the anomalies discussed in SADOVSKY *et al.* (1979) occur several days before earthquakes 500 or more km away, the time and distance scales of this migration are not comparable to those proceeding the Haicheng and Songpan-Pingwu earthquakes. A case that can be directly contrasted with the Chinese examples is the 1978 Izu-Oshima-Kinkai earthquake in Japan. Anomalies appeared 3 to 5 weeks before this earthquake within 50 km of the epicenter, but only several days before the earthquake at distances between 50 and 150 km (OKI and HIRAGA, 1979).

Water Level Anomalies and Preseismic Deformation

The quantity we hope to measure with water wells is aseismic precursory strain. Opinions as to the nature of this strain fall into three broad categories. One class of theories conjectures that irreversible fracturing occurs throughout a volume containing the future hypocenter. Another class postulates the existence of propagating deformation fronts that can trigger earthquakes. A third class suggests that precursory slip will occur in the plane of the fault that will rupture in the future earthquake. This section will focus on the types of anomalies that would be expected, based on each of these models. More detailed consideration is given to the models that predict precursory slip.

Strains caused by development of a fractured volume

Many observers have ascribed water level changes to the opening of fresh fractures, especially where the change consists of a drop in water level, followed by a recovery that begins shortly before the earthquake. However, no attempts have been made to estimate the size of the volume that is being fractured, or the magnitude of the strain associated with the fracturing. The possiblity that development of a fractured volume ('dilatancy') might occur before the onset of an earthquake was extensively discussed in the mid 1970's, as an explanaton for observed drops in compressional wave velocity before earthquakes in California, New York State, and in the Soviet Union (e.g., WHITCOMB *et al.*, 1973; NUR, 1972). Drops in compressional wave velocity were ascribed to undersaturation resulting from an increase of void space as the result of fracturing; compressional wave velocity returned to normal as fluid from surrounding non-dilatant regions flowed in to fill the new void space. Although undersaturation is necessary to cause a significant drop in compressional wave velocity, it is not required for the production of a water level drop.

RICE and RUDNICKI (1979) have studied a model of the earthquake generation process that includes a strain-softening inclusion in which dilatant hardening occurs. Dilatant hardening of the inclusion delays the onset of dynamically unstable slip. Precursors may be detected that are due either to the period of self-driven slip preceding dynamic instability, or to the pore pressure drop in the dilatant inclusion. However, they found that rapid drops in pore pressure were calculated to occur only toward the latter part of the period of precursory slip. Furthermore, their calculations offer no explanation for the return of pore pressure to its predilatancy value at any time before the earthquake.

One objection to the idea that fresh fractures are generated before an earthquake is that the response to increasing stress is more likely to be sliding on pre-existing faults than breakage of intact material. HADLEY (1973) showed that for samples of granite and gabbro at room temperature, sliding on pre-existing planes occurred before significant microcracking for confining pressures below about 2 kilobars. If fracture growth occurs primarily at great depth, large water level anomalies would not be expected to be produced near the surface.

Although water level drops may plausibly be expected in wells that are located in a zone undergoing fracture, it is not obvious that wells outside that zone would also experience drops. As the fractured zone undergoes extensional strain, it will exert compressive forces on the surrounding material that will tend to cause water level rises if the strain rate is fast, relative to the speed with which fluid can drain from the region. In particular, wells located more than several source dimensions from the hypocenter are not likely to be located within the zone undergoing fracture.

Strain due to a propagating deformation front

Propagating deformation fronts are a possible mechanism for water level anomalies that are very distant from the epicenter, and, especially, for anomalies that propagate toward the epicenter over large distances. Perceived migration of earthquakes, fault creep, and ground deformation are evidence for the existence of such fronts; KASAHARA (1979) lists these observations. SCHOLZ (1977) suggests that apparent migrations of epicenters, microseismicity, and tilt anomalies were manifestations of a deformation front traveling northward at a speed of 110 km/year that triggered the 1975 Haicheng, China earthquake.

Two mechanisms have been proposed for such deformation fronts. SAVAGE (1971) suggested that dislocation distributions may travel as kinematic waves along transform fault zones filled with gouge having constitutive properties such that strain rate is proportional to a power of shear stress. The tendencies of the dislocation distribution to both propagate and spread out diffusively can result in concentration of stress even in the absence of fault zone heterogeneity. IDA (1974) developed a solution for a crack advancing at a steady speed along a fault filled with viscous gouge. Without addressing the mechanism of propagation, LEHNER *et al.* (1981) studied the stress field ahead of a crack traveling along a plate boundary in a lithosphere overlying a viscoelastic asthenosphere. For a plausible choice of material properties, they showed that slowing of such a crack below a velocity of about 100 km/year could be expected to produce stress concentrations ahead of the crack tip, due to the reduced load-carrying capacity of the asthenosphere at lower speeds. Consequently, such a traveling front might trigger an earthquake upon encountering an obstacle to its propagation.

If the traveling deformation front takes the form of a propagating dislocation distribution, then it would be expected to produce water level anomalies consisting of either a rise or a fall in water level that returns to its previous value after passage of the front. The size of the anomaly would depend on the propagation speed, the distance of the well from the path of the front, and the diffusivity of the crust between the well and the path of the front (see the analysis for propagating creep events in ROELOFFS and RUDNICKI, 1985).

Strains caused by preseismic slip

The sizes of water level anomalies that might be associated with preseismic slip on the future fault plane are easier to quantify than those that might be associated with fracturing. Below, the amplitude of the anomalies that have been reported are compared with the strains produced by fault slip. Then several models of the earthquake process that predict precursory slip at different time scales and over different fault areas are summarized.

Comparison of anomaly amplitudes with strain produced by fault slip. If the material around a fault is assumed to behave elastically, then the volume strain produced by slip on a portion of the fault can be calculated. The moment of precursory slip is unknown, but the moment of the expected coseismic slip may plausibly be taken to be an upper bound. Although no firm theoretical basis exists for requiring preseismic slip to be less than coseismic slip, many workers believe that preseismic slip is probably only a small fraction of coseismic slip. This belief is supported by data from borehole strainmeters during several moderate earthquakes, including the 1978 Izu-Oshima-Kinkai earthquake, that show no evidence of preseismic slip, and consequently suggest that the moment of preseismic slip must be less than a few percent of the seismic moment of the earthquake (JOHNSTON *et al.*, 1987).

Volume strain may be estimated from observed water level change, using the relationship suggested by the plot of tidal sensitivity versus depth (Figure 1 and equation (6)). The strain inferred from the water level change should be less than the largest coseismic volume strain predicted for the distance of the observation point from the eventual hypocenter. In Figure 4, this hypothesis has been tested for earthquakes of different moments by plotting strain per unit moment for the data set listed in Table 2. The data set consists of water level anomalies for which well depth, epicentral distance, earthquake magnitude, and size of anomaly were reported, and consequently omits many of the anomalies listed in the Appendix. The vertical error bars include the uncertainty in the tidal calibration from equation (6), as well as a factor of 10 uncertainty in the seismic moment. Although no horizontal error bars are plotted in Figure 4, an estimate of rupture length would be an appropriate

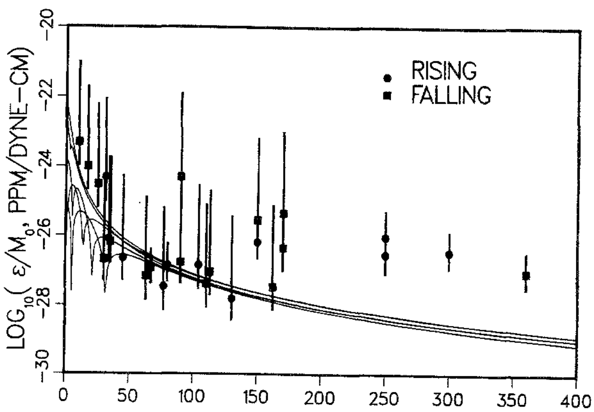

Figure 4

Comparison of strain represented by water level anomalies with predicted maximum coseismic volume strain. Symbols: strain represented by the water level anomalies listed in Table 2, converted to strain using equation (6), and divided by the moment of the subsequent earthquake. Curves: Absolute value of volume strain as a function of distance along the azimuth of maximum volume strain, for several source mechanisms: vertical strike slip, at surface, and 9 km deep; 45 degree dip slip, at surface, and 9 km deep, and 20 degree dip slip, 20 km deep. Curves represent strain per dyne-cm of moment, assuming a shear modulus of 3×10^{11} dynes/cm^2.

measure of the uncertainty in distance. Empirical relationships between magnitude and rupture length (e.g., SCHWARTZ *et al.*, 1984; WYSS, 1979) suggest that for all but two of the earthquakes listed in Table 2, the rupture length is less than 10 km. The two exceptions are the 1976 Tangshan earthquake, for which a rupture length of 140 km is reported by BUTLER (1979), and the 1978 Izu-Oshima-Kinkai earthquake, for which a rupture length of 17 km is reported by SHIMAZAKI and SOMERVILLE (1978).

The comparisons made in Figure 4 must be viewed with caution, because the conversion of water level to strain is crude, and because the data have not been corrected for azimuth relative to the strike of the earthquake fault. Bearing these inadequacies in mind, Figure 4 suggests that, out to 150 km from the epicenter, many reported precursory anomalies can be accounted for by preseismic slip of moment less than or equal to the future earthquake moment. In particular, the sizes of water level anomalies listed in Table 2, preceding the 1976 Tangshan and 1978 Izu-Oshima-Kinkai earthquakes, can be accounted for in this way. Further than 150 km from the epicenter, most reported anomalies appear to correspond to an amount of strain greater than the calculated coseismic strain, even allowing for the considerable uncertainty in seismic moment, well sensitivity, and distance of the well from the initiation point of the rupture. If the anomalies at such great epicentral distances are genuinely related to the subsequent earthquakes, then they may be caused by a mechanism such as a traveling deformation front.

Models that predict precursory slip. DIETERICH (1986a, b) has studied behavior of spring-slider models embodying a slip-rate and rate-history dependent frictional constitutive law. In such a system, slip instability is always preceded by accelerating stable slip over a patch of the fault, which might give rise to premonitory signals in the interval from 10 days to 10 minutes before an earthquake. The moment of premonitory slip is governed by a characteristic slip distance and is not necessarily related to the moment of the impending earthquake, which may rupture an area larger than the nucleation patch.

The characteristic slip distance is believed to be related to the scale of roughness of the fault surface (e.g., TULLIS, 1987). Estimates of this distance from laboratory measurements are less than 50 microns, but larger values may be appropriate for earthquake faults. If laboratory values of characteristic distance are appropriate, then only a fraction of a millimeter of precursory slip would be expected over a radius of several meters. However, if the characteristic distance is as large as 50 mm, then precursory accelerating slip of 250 mm could occur over a patch of radius several km, yielding precursory slip with a moment of 5.5×10^{24} dyne-cm (DIETERICH, 1986a). This moment is not large enough to account for many of the anomalies listed in Table 2, even when only anomalies with precursor times of 10 days or less are considered.

Models based on slip-rate and rate-history dependent friction laws predict that once slip has begun to accelerate over the nucleation patch, displacement on that

Table 2

Precursory water level anomalies plotted in Figure 4. Delta: distance from well to epicenter. Delta: distance from well to epicenter. dh: maximum amplitude of water level change.

Earthquake location	Date (*YYMMDD*)	Magnitude (note 1)	Seismic moment (note 2)	Delta (km)	Well depth (m)	dh (cm)	Reference
Nazarbek, USSR	801211	5.1	8.9E23	150	295	−3.3	Kissin
				170	235	−0..5	*et al.* (1983)
				170	235	−5.0	
S Kuril'sk, USSR	810621	4.5	5.6E22	90	500	−4.0	Monakhov *et al.* (1983)
Liyang, China	790709	6.0 M_S	5.6E25	45	367	4.4	Wang & Wu
				130	369	0.3	(1986)
Fengzhen, China	810813	5.8 M_S	2.2E25	113	217	−1.0	Wang & Wu (1986)
Lunyao, China	811109	5.8 M_S	2.2E25	67	3034	−3.9	Wang & Wu
				150	2694	19.6	(1986)
Lulong, China	821019	5.3 M_S	2.2E24	34	301	3.0	Wang & Wu
				250	1735	7.0	(1986)
				300	2694	3.4	
Heze, China	831107	5.9 M_S	3.5E25	80	2267	5.0	Wang & Wu
				250	1088	7.0	(1986)
				360	2694	−3.6	
Bear Valley, CA	720224	5.0 M_L	1.6E23	25	152	−5.0	Kovach *et al.* (1975)
Stone Canyon, CA	720904	4.7 M_L	7.9E22	17	152	−8.0	Kovach *et al.* (1975)
Lewis Ranch, CA	730115	4.0 M_L	1.0E22	10	152	−5.0	Kovach *et al.* (1975)
Izu Pen., Japan 780114	780114	6.8 M_S	1.1E26	30	500	−30.0	Wakita
			(note 3)	35	600	−100	(1978)
				90	500	−25.0	
Tangshan, China	760728	7.7 M_S	1.8E27	77	60	60.0	Wu *et al.*
			(note 4)	63	61	−120.	(1984)
				104	100	260.	
				110	63	−73.0	
				162	283	−70.0	
California, USA	800225	5.5	1.0E24	31	32	45.0	Merifield and Lamar (1985)

Notes. (1) magnitude scale omitted where unknown.

(2) unless otherwise indicated, seismic moment was estimated from the moment-magnitude relation given by Hanks and Boore (1984), Figure 2, for California earthquakes, or from Nuttli (1985), equations (6a, b, c) and (7a, b, c), for earthquakes not in California. Magnitudes of unspecified scale were assumed to be m_b.

(3) Seismic moment from Shimazaki and Somerville (1978).

(4) Seismic moment from Butler *et al.* (1979).

patch increases rapidly. For a network of water level sensors distributed around the fault, each with a finite strain detection threshold, this increasing displacement will be detected over time at increasingly greater distance from the nucleation patch. Consequently, the anomaly would appear to emanate from the future epicenter. Strain anomalies would be expected to increase monotonically in a sense consistent with the future coseismic volume strain and begin within 10 days before the earthquake (the interval during which slip would be expected to accelerate to instability). A possible example is the accelerating water level drop at Funabara, preceding the 14 January 1978 Izu-Oshima-Kinkai earthquake in Japan (WAKITA, 1981), although this anomaly began about one month before the earthquake. Nevertheless, it is clear that many reported anomalies do not fit this description.

STUART *et al.* (1985) have studied a model of the earthquake generation process in which the fault zone, taken to be a vertical plane in an elastic half-space, has spatially varying resistance to frictional slip. In particular, a patch of the future earthquake rupture area is assumed to obey a strain softening relation between slip and frictional resistance, once a peak stress has been exceeded. Outside this patch, the fault slips freely at a constant stress, or is completely locked. When such a system is subjected to steadily increasing tectonic shear stress, four distinct stages of behavior result. During the first stage, the patch behaves similarly to other areas of the fault. Its increasing resistance to frictional sliding begins to become significant at the beginning of the second, 'slow load' stage, but during this stage, all points on the patch are still in a state of strain hardening. Increasing tectonic load and concentration of stress on the patch gradually bring its outer edges past their peak stress, at which time the third, 'precursory' stage of accelerating slip begins. Earthquake instability results when tectonic loading and stress concentration combine to load the central, unfailed, portion of the patch more quickly than stress can be relieved with any finite rate of slip.

The stage at which accelerating, aseismic fault slip might be observed is the third, 'precursory', stage. For the model parameters obtained by calibrating the model to agree with observed creep and line-length changes at Parkfield, CA, STUART *et al.* (1985) suggest that these changes might become observable about one year before the predicted moderate earthquake. The length of the precursor time in other locatons would be expected to depend on the tectonic loading rate, as well as on the slope of the unloading portion of the slip versus resistance relation. Clearly, there are few contexts where these quantities can be constrained, even to the degree they are at Parkfield.

Accelerating slip occurs first at the edges of the patch and spreads inward toward the future hypocenter. The creep and line-length changes at Parkfield are most consistent with a patch radius of 3 km. In this situation, although the locus of accelerating slip converges toward the hypocenter, it does so over a small enough distance scale that the inward migration might not be apparent in water level records. Unless the patch dimension is very large for large earthquakes, this mechanism cannot explain the movement of hydrologic anomalies toward the epicenter over distances on the order of 100 kilometers, such as was observed before the 1976 Songpan-Pingwu earthquakes.

LI and RICE (1983) visualize earthquakes as the result of unstable propagation of slip upward from the base of the lithosphere. Stress is thereby concentrated in a narrowing region of the crust; the details of the eventual failure of this region may be governed by the processes studied by DIETERICH (1986a, b) and STUART *et al.* (1985). According to this model, the length of fault involved in precursory slip is comparable to the earthquake rupture length.

In this model, as slip propagates upward along the plate boundary, the shear stress averaged through the lithosphere drops. Instability begins when the load-carrying capacity of the lithosphere drops more rapidly than load is transferred to it by the relaxed lithosphere-asthenosphere system. As slip accelerates, the lithosphere-athenosphere system responds in a stiffer, unrelaxed manner. Earthquake instability occurs when even the completely unrelaxed system loads the plate boundary more rapidly than it can sustain the load. The precursor time is the time of accelerating slip between initial and final instabilities. For the material properties and thicknesses of the asthenosphere and lithosphere chosen by LI and RICE (1983), this period can be from 2 months to 5 years long, and would be a time during which increasing strain rates would be observed. Because this time period is longer than that predicted by studies of the details of rupture initiation in the upper part of the crust, it might be difficult to distinguish these anomalies from hydrologic trends. However, gradual declines in water level before the 1976 Tangshan earthquake (WANG *et al.*, 1979) may be an example of this type of process.

The three models discussed here address the earthquake generation process at different spatial scales. Consequently, they yield different estimates of the length of time during which precursors should be observed, and of the size of the area that should undergo preseismic slip. The size and lead time of carefully documented hydrologic anomalies might help ascertain the appropriateness of these models.

Recommendations for Improved Documentation of Hydrologic Precursors

Water level monitoring shows considerable promise as a technique for measuring crustal strain, and reliable observations should be able to help evaluate proposed fault models. Poor documentation of data is the primary obstacle to greater utilization of water level monitoring. In the hope of improving this situation, a list is given here of the information that should be collected and reported in every case of a possible hydrologic precursor:

1. Depth of the well.
2. Record of rainfall for at least one year.
3. Record of barometric pressure, measured at least once every three hours, even if data have been corrected for barometric pressure.

4. Information about wells being pumped in the same vicinity.
5. The longest available record of observations should be shown.
6. Measurement technique: pressure transducer, float recorder, tape soundings.
7. Sampling interval.
8. Response to earth tides.
9. Coseismic response of water level to the subsequent earthquake.
10. Magnitude of earthquake.
11. Depth and focal mechanism of earthquake.
12. Times, magnitudes, and hypocenters of any recorded foreshocks.
13. Distance of azimuth of well from epicenter.
14. A graph of raw water level versus time during the anomaly.
15. Descriptions of any other wells in the vicinity that were monitored but did not show anomalies.

Information on the geology of the site, particularly its proximity to any fault zones, is also useful. Some assessment should be given of whether the reservoir is confined.

Most of the items on this list should be available in every situation, except for the tidal response and the record of barometric pressure. If the other information is available, then these data can be collected after the anomaly occurs. In situations where the cost of continuously recording such information at every well is prohibitive, the required data could be collected using a portable recorder rotated among the wells in an observation network, remaining at each well for about two months.

Conclusions

Clearly, there is a solid theoretical basis for expecting water level anomalies if precursory aseismic deformation produces volume strain in the crust. Increasing evidence shows that water wells can be used as calibrated strainmeters.

Well-documented premonitory groundwater anomalies have been reported to occur years (e.g., Tangshan, 1976), months (e.g., Haicheng, 1975; Izu-Oshima, 1978), and or days (e.g., Pamirs, 1978) before the subsequent earthquakes. Although many investigators state that falling water levels are more common, both rising and falling water levels have been reported, and are expected if there are regions of contraction and extension, respectively, in the crust. Anomalies occur as far as several hundred kilometers from earthquake epicenters, and are often reported to be concentrated in fault zones. Migration both toward and away from the epicenter has been discerned.

Real progress in the use of water levels to measure crustal strain requires better documentation of the observations. Although all reports of water level anomalies to date are deficient in some aspect, the cases of the 1976 Tangshan, China, and 1978 Izu-Oshima-Kinkai, Japan, earthquakes are the best documented and should be thoroughly studied.

Figure 4 suggests that many water level anomalies within 150 km of earthquake epicenters are not too large to be explained by preseismic slip of moment comparable to the earthquake. Although this conclusion must be regarded as tentative, it appears quite likely that water level anomalies can help answer the question of the moment and timing of accelerated preseismic slip, if it occurs.

Appendix

Reports of Premonitory Variations in Level or Flow of Water, Oil, or Gas

Following is a brief summary of published accounts that report premonitory variations in the level of flow of water, oil, or gas. These reports have been obtained primarily from English language literature. Supporting information on well depth, barometric pressure, rainfall, and length of the period of observation is critical in evaluating the relation between any reported anomalies and the earthquake preparation process. Availability of such information is noted in the descriptions.

The list of reports is in chronological order. The word 'magnitude' is used to denote value on an unspecified magnitude scale. An attempt has been made to refer to Chinese place names using the current official Mandarin transliteration; other transliterations under which information appears are included in parentheses.

18 January 1964, magnitude 6.7, Chia-yi, Taiwan: Tubing pressure in oil-field wells increased between 15% and 60%, beginning on 9 January (WANG and LI, 1975).

8 March 1966, magnitude 6.8, and 22 March 1966, magnitude not stated, Xingtai, Hebei, ('Hsingtai, Hopei') China: groundwater levels rose or fell within a 230 km by 190 km area before the first event, and over a 300 km by 220 km area before the second event. (WANG and LI, 1979.) Scientists measuring well water levels with tapes in the epicentral region after the first event observed drops in water level before aftershocks (ZHANG, Y.-Z., personal communication).

3 December 1970, magnitude 5.5, Hai-yuan, Ningxia, China: Water level rises of as much as 0.8 m were observed (WANG and LI, 1979).

24 February 1972, $M_L = 5.0$, Bear Valley, California, USA: In a 152 m deep well, 25 km from the epicenter, a water level drop of approximately 5 cm occurred about 60 days before this earthquake and persisted for about 25 days. The water level drop does not appear to be due to the seasonal effect of rainfall. A correction has been applied for barometric pressure variations. This well was monitored beginning in May, 1971 (KOVACH *et al.*, 1975).

4 September 1972, $M_L = 4.7$, Stone Canyon, California, USA: In the same well for

which the precursor of the 24 February Bear Valley event was reported, a water level drop of about 8 cm was observed about 50 days before the earthquake, with a duration of about 40 days. This earthquake occurred 17 km from the well (KOVACH *et al.*, 1975).

15 January 1973, $M_L = 4.0$, Lewis Ranch, California, USA: In the same well for which the precursor of the 24 February Bear Valley event was reported, there was a 5 cm water level drop 40 days before the Lewis Ranch earthquake, which persisted until the earthquake occurred. The epicenter of the Lewis Ranch earthquake was 10 km from the well (KOVACH *et al.*, 1975).

23 April 1974, magnitude 5.5, Li-yang, Kiangsu, China: rapid water level rises occurred in two wells (one rise was about 1 meter); water overflowed from a third well (WANG and LI, 1979).

4 February 1975, magnitude 7.3, Haicheng, Liaoning, China: About 240 water level anomalies were reported between 16 November, 1974, and the time of this earthquake. These anomalies are described in detail by ZHU and ZHONG (1979) and DENG *et al.* (1981)

8 April 1976, $M_S = 7.1$, and 17 May 1976, $M_S = 7.1$, Gazli, USSR: There was a sharp drop in water level (SADOVSKY *et al.*, 1979).

28 July 1976, magnitude 7.8, Tangshan, Hebei, China: Numerous water level anomalies were observed, perhaps as much as 6 years before this earthquake. They are described in detail in WANG *et al.* (1979), WANG and LI (1979), ZHANG and QIUO (1979), and WU *et al.* (1984). WANG *et al.* (1979) and WU *et al.* (1984) discuss the effects of rainfall and groundwater exploitation on the data.

16 August 1976, magnitude 7.2, Songpan-Pingwu, Sichuan province, China: 395 groundwater changes were observed beginning in March, 1976, becoming larger in June 1976, and clustering along the fault in the days before the earthquake. (WANG and LI, 1979; WALLACE and TENG, 1980; RIKITAKE, 1982; WANG, 1985)

24 November 1976, $M_S = 7.3$, Caldiran, Van province, Turkey: About 1 day before the earthquake, increased discharge from a spring, accompanied by oil seepage, occurred at the eastern end of the 55 km long section of fault along which surface displacement was mapped (TOKSOZ, 1979).

12 May 1977, magnitude 6.5, Ningho, Hebei, China: a usually dry well in Chinghsien county, 110 km from the epicenter, discharged oil and gas between 27 March and 12 April, 1977; oil production in this well prior to the 28 July 1976 Tangshan earthquake was also believed to have been a precursor. Similarly, a well in Jenchiu county, 160 km from the epicenter, produced water from 10 March to 12 March, and gushed oil from 22 March to 24 March; it had also gushed oil before the 1976 Tangshan earthquake (WANG and LI, 1979).

23 July 1977, magnitude 5.5, Xinjiang ('Singkiang'), China: A 380 m deep well, 50 km from the epicenter, began to produce four times its normal amount of fluid, with double the normal content of oil, beginning on 10 July, 1977. The same well began to produce only water and mud on 21 July, and stopped flowing after the earthquake. Flow rates increased in two other wells on 19 July, and another previously dry well yielded gas on 23 July (WANG and LI, 1979).

14 January 1978, $M_S = 6.8$, Izu-Oshima-Kinkai, Japan ('West Off Izu-Oshima earthquake'): Three water level anomalies were observed in deep monitoring wells. Water level in a 500 m deep well at Nakaizu, 30 km from the epicenter, fell approximately 30 cm at the beginning of December, 1977, and had risen again by about 20 cm by the time the earthquake occurred. In a 600 m deep well at Funabara, a rising trend of water level reversed itself in the middle of December, 1977; water level fell about 1 m before the earthquake. Funabara is 35 km from the epicenter. 90 km from the epicenter at Omaezaki, water level in a 500 m deep well dropped by 25 cm, beginning on 28 December, 1977, and then began to rise on 5 January, 1978. These three water level anomalies are described by WAKITA (1978), who refers to more detailed reports in the Japanese literature. In addition to the deep-well anomalies, water level anomalies were observed at over 30 locations, mostly shallow wells, at which groundwater levels were being monitored by the Catfish Club. These anomalies are described in OKI and HIRAGA (1978), in a very complete report that includes information on the response of the monitoring wells to barometric pressure, rainfall, and groundwater pumping.

27 February 1978, magnitude 4.6, and 7 September 1978, magnitude 4.5, Ashkhabad, USSR: About two months prior to the first of these events, water level in a well 120 km from the epicenter dropped about 1 meter, and had recovered by about half that amount when the earthquake occurred. In the same well, a drop of about 1.5 m began 5 months before the second event, and was beginning to recover when the earthquake occurred, at a distance of 20 km from the well. Water level observations had been made in this well since 1970 (ASIMOV *et al.*, 1979).

22–23 March 1978, sequence with magnitudes up to 8, Kuril'sk, USSR: In a well at Goryachiye Kluchi, 170 km from the epicentral region, flow of water increased from about 1100 to over 1200 liters/hour 4 days before the first magnitude 7 event; the flow dropped to its previous rate 1 day before the earthquake. A day or two after this event, the flow again decreased, and a magnitude 8 event occurred several days later. It is stated that atmospheric pressure and precipitation do not affect the flow rate in this well (MONAKHOV *et al.*, 1983).

1 November 1978, magnitude 7, Pamirs, USSR: Six hours before this event occurred, its time was successfully predicted to within a few hours, and its size was predicted to within 1/2 of a magnitude unit. The location, however, was predicted only to within a few hundred kilometers. A long-term decrease in water levels near Khodz-

haabad, in the Fergana Valley, beginning the previous July, was a possible intermediate-term hydrologic precursor. In the same area, cessation of flow in the deepest of 11 wells being monitored and decrease of flow in the remaining wells at the end of October was interpreted as a short-term precursor. These hydrologic precursors occurred at a distance of 150 to 200 km from the epicenter (ASIMOV *et al.*, 1979; D. W. SIMPSON, 1979).

7 July 1979, $M_S = 6.0$, Liyang, China: Of six wells being monitored within 200 km of the epicenter, 2 wells about 370 m deep showed several centimeter rises in water level in the 2 weeks before the event. The water level rise was observed in a well 45 km from the epicenter, 13 days before the earthquake; it was observed in a well 130 km from the epicenter, 5 days before the earthquake. Another well, 600 m deep, showed a drop in flow rate 5 months before the event; this well was 60 km from the epicenter (WANG and WU, 1986).

25 February 1980, $M_L = 5.5$, southern California, USA: Two wells located 31–32 km from the epicenter showed anomalous rises of water level lasting about 4 hours, 4 days before the earthquake. The rise totaled 45 cm in a 32 m deep well, and 2.3 cm in a 130 m deep well (MERIFIELD and LAMAR, 1985).

11 December 1980, magnitude 5.1, Nazarbek, USSR: Between 30 and 90 days before this event, a 5 cm drop in water level was observed in a 235 m deep borehole at Kim, 170 km from the epicenter; this drop persisted until about 2 months after the earthquake. In the same borehole an additional 0.5 cm drop occurred 5 days before the earthquake, and recovered after 3 days. Two days before the earthquake, there was a 3.3 cm drop in a 305 m deep well at Asht, which is located 150 km from the epicenter. Water level observation in these wells began in early 1980; their barometric response and their response to rainfall have been studied. Tidal signals have been observed at Kim, but do not appear to occur at Asht (KISSIN *et al.*, 1983).

13 August 1981, $M_S = 5.8$, Fengzhen, China: a water level drop of one centimeter was observed in a 217 m deep well situated 113 km from the epicenter, 18 days before the event; water level rose in the two days before the earthquake (WANG and WU, 1986).

9 November 1981, $M_S = 5.8$, Lunyao, China: of 9 wells being observed within 200 km of the epicenter, 2 artesian wells showed changes of water pressure on the order of 4 to 20 cm of water in amplitude. The first anomaly occurred 77 days before the earthquake in a 2694 m deep well 150 km from the epicenter, and the second occurred 11 days before the earthquake in a 3034 m deep well 67 km from the epicenter (WANG and WU, 1986).

22 March 1982, $M_L = 4.5$, southern California, USA: Rapid fluctuations of water level occurred in a 133 m deep well 13 km from the epicenter. These fluctuations

occurred over a 4 day period and are tentatively attributed to gas bubbling (MERIFIELD and LAMAR, 1985).

3 July 1982, $M_S = 5.4$, Jianchuan, China: In a well of unspecified depth 10 km from the epicenter, water level fell 13 cm 13 days before the earthquake (WANG and WU, 1986).

19 October 1982, $M_S = 5.3$, Lulong, China: Three of the five wells within 30 km of the epicenter at which water levels were being observed showed water level rises of several centimeters. The change was observed 7 days before the earthquake in a well 300 km away, and 2 days before the earthquake in wells 250 and 34 km away. Well depths range from 300 to 2700 m. In the well 300 km from the epicenter, water level had been monitored for at least the previous year (WANG and WU, 1986).

3 March 1983, $M_S = 5.4$, Xinjiang, China: Two wells 11 and 12 km from the epicenter showed water level changes of 13 and 20 cm, respectively, two days before the earthquake (WANG and WU, 1986).

7 November 1983, magnitude 5.9, Heze, Shandong, China: Of 58 wells within 400 km, 8 wells, with depths between 1000 and 2700 m, showed anomalies of water level, oil production, and flow. In two wells showing water level changes, the change occurred 27 days before the earthquake in a well 80 km from the epicenter, and occurred 13 days before the earthquake in a well 360 km from the epicenter (WANG and WU, 1986).

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