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Role of pore fluids in faulting

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We consider three *in situ* processes which involve fluid flow in the crust: fault creep, aftershocks and dilatancy. Measurements of water level in wells suggest that creep events on the San Andreas fault are coupled with pore pressure changes. Readjustment of transient pore pressure, induced by large shallow earthquakes, possess the correct time constants and magnitudes to explain the occurrence of aftershocks. And finally, temporal changes of travel times in the Gram district (U.S.S.R.) imply that dilatancy may occur *in situ*.

Introduction

There are several known *in situ* phenomena which indicate that pore fluids play an important mechanical role. Extensive removal from, or addition to the ground of, fluids can cause measurable elastic deformation of the ground as well as failure. The reversed process, where deformation causes significant changes in pore pressure has also been observed. For example, the solid tide of the Earth induces pore pressure changes which have been measured in wells (Sterling & Smets 1971).

Recent experimental and theoretical studies yielded three new results which are closely related to *in situ* faulting processes.

Fluctuations of water level due to fault creep

A 150 m well was drilled, in 1971, 270 m north of the Cienega Winery near Hollister, California, 9 m east of the active surface trace of the San Andreas fault. A mechanical system was used to record the level of water in the well continuously with time. Superimposed on the initial rise of water level shown in figure 1 – an adjustment to the drilling – are fluctuations of various periods and amplitudes. A comparison (figure 1) with the barometric pressure record at the well site reveals clear correlation between water level and long period atmospheric pressure fluctuations. A two-point leaky integrator which acts as filter and delay was applied to the raw barometric data so that by summing the level data and filtered atmospheric data the resultant trace is free of atmospheric effects. It can be observed in figure 1 that the corrected level is almost flat, except for three abnormal level changes which are not related to atmospheric pressure. These three level events coincide within hours with the only creep events which were detected at the Winery, 270 m farther south on the fault. Clearly there exists a one-to-one correspondence between water-level changes and creep events on the fault at this site. The level change Δh is approximately proportional to the amount of creep Δu, with $\Delta h/\Delta u \simeq 14$ mm/mm. Although creep has always the same right lateral motion, the water level has increased with the first and third creep event, and decreased with the second event. Furthermore, the level changes preceed the creep in the first and third event, but follow the creep in the second event. The data are summarized in table 1. The simplest explanation for these temporal and sense relations between creep and water level changes can be obtained by invoking a model of a slip zone which propagates along the fault.

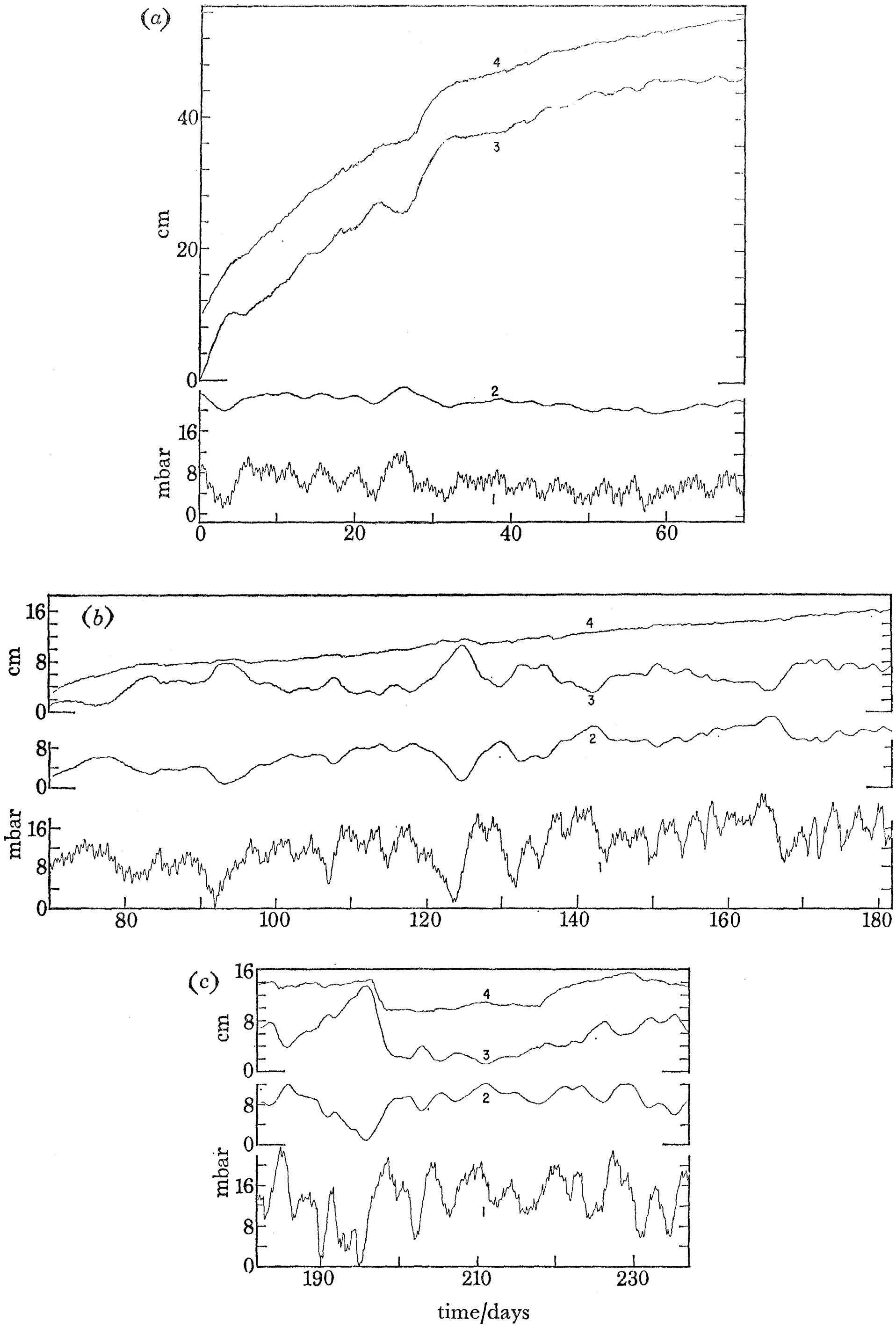

FIGURE 1 *a*, *b*, *c*. Data collected at Cienega Winery from 13.00 P.S.T. 14 June 1971, to 13.00 P.S.T. 7 February 1972. Time is marked in days from the beginning. The barometric pressure (line 1) is filtered (line 2) to obtain the appropriate barometric correction to be applied to the water level data (line 3) and yield the corrected water-level record (line 4). The filling of the well is disturbed near day 30 (12 July 1971), and rather sharp changes occur on days 197 (27 December 1971) and 218 (17 January 1972).

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Table 1. Summary of data for creep events and associated water level fluctuation at Cienega Winery, near Hollister, California

date	creep mm	duration of creep/h	water level change/cm	rise or decline	duration of level change h	time lag of creep after level change h	computed tectonic pressure 10^2 Pa (mbar)
12 July 1971	4	16	5.6	rise	100	−4	+18
27 December 1971	3	54	4.1	decline	48	+30	−15
19 January 1972	2	40	3.2	rise	80	−8	+7

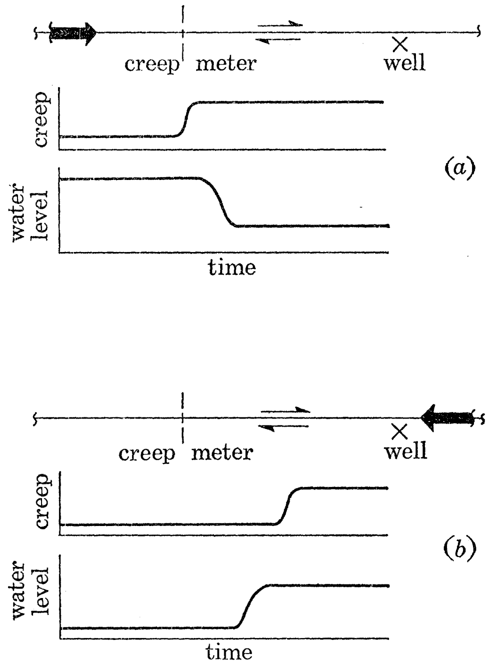

Figure 2. Schematic diagram of creep event propagating from the north (*a*) and from the south (*b*) with the resulting signals shown for the creep offset and the water level.

Figure 2 is a schematic representation of propagating creep event and the response of our water well. The heavy black arrow represents the propagating tip of a growing slip zone. In figure 2*a* creep is propagating to the north such that a response is seen first at the creep recorder site. Because the motion is right lateral there is a zone of dilatation produced at the well site with subsequent lowering of the water level.

In figure 2*b* the sense of propagation is reversed so that there is a compressional zone produced at the well as the event propagates by the well. In this case the water level in the well would rise. According to this interpretation the events described in table 1 could be indicative of the direction of creep propagation.

Aftershocks induced by transient pore pressure

The frequency of occurrence of aftershocks following large shallow earthquakes decays with time. This decay requires a time-dependent process that is much faster than the large-scale tectonic loading and much slower than the propagation of elastic waves.

We decided to investigate the possibility that significant changes in pore pressure can be generated by large earthquakes. The decay of these pressure anomalies as a result of the flow of the pore fluid can provide the necessary viscous element in the form of an apparent time-dependent strength which can cause aftershocks.

A very simple example illustrates our ideas. We approximate the end of the fault break by an edge dislocation in an otherwise homogeneous infinite elastic space. In this model the hydrostatic stress induced by the sudden appearance of an edge dislocation in a coordinate system r, θ about the end of the crack is

$$\bar{\sigma}(\boldsymbol{r}) = [\sigma_{rr}(\boldsymbol{r}) + \sigma_{\theta\theta}(\boldsymbol{r})]/2] = A \sin \theta/r,$$

where

$$A = \mu b/2\pi\,(1-v),$$

b is the offset across the break, and μ and v are, respectively, the shear modulus and the Poisson ratio for the material.

Typical values for μ and v are about 10^{10} Pa (10^5 bar) and 0.25, respectively. At a distance of 1 km from the end of a fault with a 1 m offset, the change of shear and volumetric stress would be about 2.5 mPa (25 bar). The shear stress drop considered reasonable for large earthquakes are of the order of 10^6 to 10^7 Pa (10 to 100 bar). It would therefore be rather surprising if the decay of the pore pressure field did not have profound effects on local seismicity.

During the earthquake the pore pressure will change by an amount equal to the volumetric stress change. Thus the initial condition on the pore pressure for the fluid flow is

$$P(\boldsymbol{r}, 0) = A \sin \theta/r.$$

Later developments in the pore pressure field will be approximately governed by the diffusion equation

$$\frac{\partial P(\boldsymbol{r}, t)}{\partial t} = c\nabla^2 P(\boldsymbol{r}, t), \tag{1}$$

where c is the hydraulic diffusivity. By means of the Hankel transform we find

$$P(\boldsymbol{r}, t) = A\,\frac{1-\exp\{-r^2/4ct\}}{r}\sin\theta. \tag{2}$$

The direction of flow is primarily from the positive peak on one side of the fault to its negative image on the other side.

The time constant associated with the pore pressure decay is

$$\tau = L^2/4c,$$

where L is a typical length. We take $L = 1$ km as a reasonable length. For unfractured solid rock, typical values for compressibility and permeability of $\beta = 5\times10^{-11}\ \text{Pa}^{-1}$ ($5\times10^{-6}\ \text{bar}^{-1}$), $K = 10^{-3}$ darcy yield $\tau = 120$ days. For fractured and jointed rock, $\beta = 5\times10^{-10}\ \text{Pa}^{-1}$ ($5\times10^{-5}\ \text{bar}^{-1}$), $K = 1$ darcy, and corresponds to 1 day. Thus τ will most likely be in the range from several hours to several months. This range of the time constant is similar to the periods observed in aftershock sequences, commonly lasting from a few days to a few months.

We now postulate that the frequency of after shocks in some volume near the main shock is proportional to the time derivative of the local pore pressure. Thus the total number of aftershocks per unit time is

$$\frac{\mathrm{d}N}{\mathrm{d}t} = \frac{1}{a}\int_V \frac{\partial P}{\partial t}\,\mathrm{d}v, \tag{3}$$

where a is a constant. The integration is carried out over the volume where pore pressure is increasing, that is, the region where the main shock produces dilatation. Differentiating equation (2) and substituting in equation (3), we find that the frequency of aftershocks

$$\frac{\mathrm{d}N}{\mathrm{d}t} = -\frac{1}{a}\left(\frac{c\pi}{t}\right)^{\frac{1}{2}}$$

decays as $t^{-\frac{1}{2}}$.

The frequency of aftershock sequences commonly show initial decay of $t^{-\frac{1}{2}}$ to $t^{-\frac{3}{4}}$ gradually shifting to t^{-1} at later time. These initial decay rates are in remarkable agreement with our prediction. There are several good reasons, however, for expecting our simple model to give only an approximate lower bound to the actual decay rate. First, the linear relation between frequency and the pressure-time derivative does not hold when the total possible number of aftershocks is small. Secondly, we have permitted only two-dimensional pressure gradients. Any three-dimensionality would cause both the pore pressure field and the aftershock frequency to decay more rapidly.

We conclude that the flow of groundwater can provide the viscous element necessary to produce aftershock sequences.

Dilatancy inferred from premonitory changes of travel times

Nersesov, Semnnov & Simbireva (1969) and Semenov (1969) reported unusual variations of the ratio $\xi = t_s/t_p$ of the travel times of shear waves to compressional waves in the Garm region of the Tadzhik Socialist Soviet Republic. In several cases, following relatively long periods

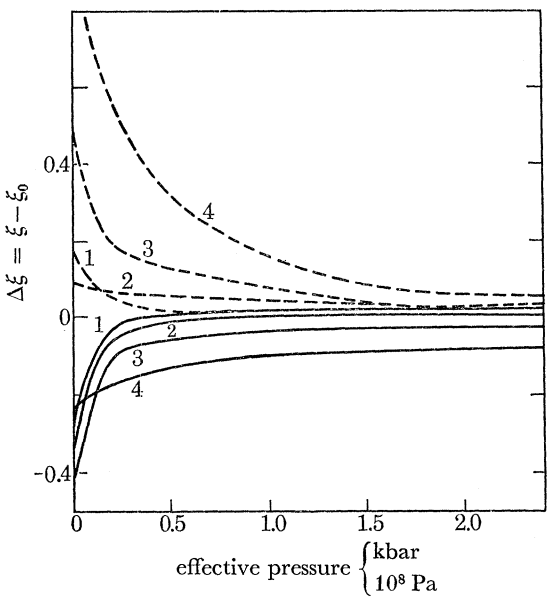

FIGURE 3. Deviation of the parameter $\xi = V_p/V_s$ from its high pressure value ξ_0, as a function of pressure, in dry (—) and saturated (- - -) rock samples. 1, Casco granite; 2, westerly granite; 3, Troy granite; 4, Bedford limestone.

(several years?) of constant ξ, ξ decreases significantly for a month or two, increases again to roughly its initial value or more, closely followed by a local earthquake. Furthermore, Savarensky (1968) reported that in some cases not only t_s/t_p, but also t_p itself changes with time. t_p is greater before an earthquake, than after. This observation suggests the possibility that the mechanical properties of the rocks in the volume affected by the main shock vary significantly with time prior to the earthquake. Thus it can be assumed that $t_s/t_p \approx V_p/V_s$, where the velocities are spatial averages over the same volume. The question now is, what determines the ratio V_p/V_s in rocks? The parameter ξ can be computed from laboratory results. In dry rock samples ξ increases with confining pressure, as shown in figure 3 and table 2, for several rocks. The increase in confining pressure reduces the pore volume V_p in rocks, due to the closure of cracks. In dry rocks, therefore, increasing ξ is associated with decreasing pore volume and decreasing ξ is associated with increasing pore volume.

Table 2. The value of ξ in dry and saturated rock as a function of effective confining pressure

(Data from Nur & Simmons 1969)

	effective pressure (10^5 Pa / bar)								
	0	50	100	200	400	700	1000	2000	3000
	Casco granite, porosity = 0.007								
ξ dry	1.42	1.64	1.71	1.72	1.74	1.75	1.77	1.76	1.75
ξ sat.	2.19	2.11	2.01	1.94	1.89	1.85	1.80	1.77	1.76
	Westerly granite, porosity = 0.009								
ξ dry	1.36	1.49	1.62	1.68	1.69	1.71	1.72	1.74	1.74
ξ sat.	1.83	1.83	1.84	1.83	1.83	1.82	1.81	1.79	1.78
	Troy granite, porosity = 0.002								
ξ dry	1.55	1.75	1.77	1.80	1.81	1.81	1.80	—	—
ξ sat.	1.97	1.93	1.87	1.85	1.82	1.81	1.82	—	—
	Bedford limestone, porosity = 0.123								
ξ dry	—	1.70	1.74	1.74	1.76	1.77	1.77	1.79	—
ξ sat.	—	2.76	2.66	2.45	2.25	2.13	2.04	1.90	—

In many cases *in situ*, existing cracks are saturated with liquids, usually water. The value of ξ in water saturated rocks is shown in figure 1 and in table 2 for several rocks. Note the relatively high value of ξ under these conditions. Furthermore, as effective confining pressure is raised, the pore volume decreases, ξ diminishes, unlike the dry case. Therefore, in saturated rocks unlike dry rocks, increasing ξ is associated with *increasing*, not decreasing, pore volume.

The time dependence of the deviation $\Delta\xi$ *in situ* from the 'steady' value ξ_0, is reproduced from Semenov (1969) in figure 4. A comparison of the 'steady' ξ_0 values with laboratory values suggests that the rocks *in situ* are subject to effective pressure in excess of 100 to 200 MPa (1 to 2 kbar). The anomalous decrease of ξ, cannot be matched with any laboratory values for saturated rocks. Dry rocks, however, do exhibit such low values of ξ. If laboratory measurements are at all indicative of *in situ* conditions, these observations imply at once that before a typical earthquake in Garm, dry cracks are being opened up in the rocks around the focal zone. These cracks which are initially devoid of ground water, cause the observed reduction of ξ. This process would also cause t_p to be anomalously long before an earthquake, as observed by Savarensky (1968).

The appearance of new cracks before failure in brittle rocks is most likely a manifestation of

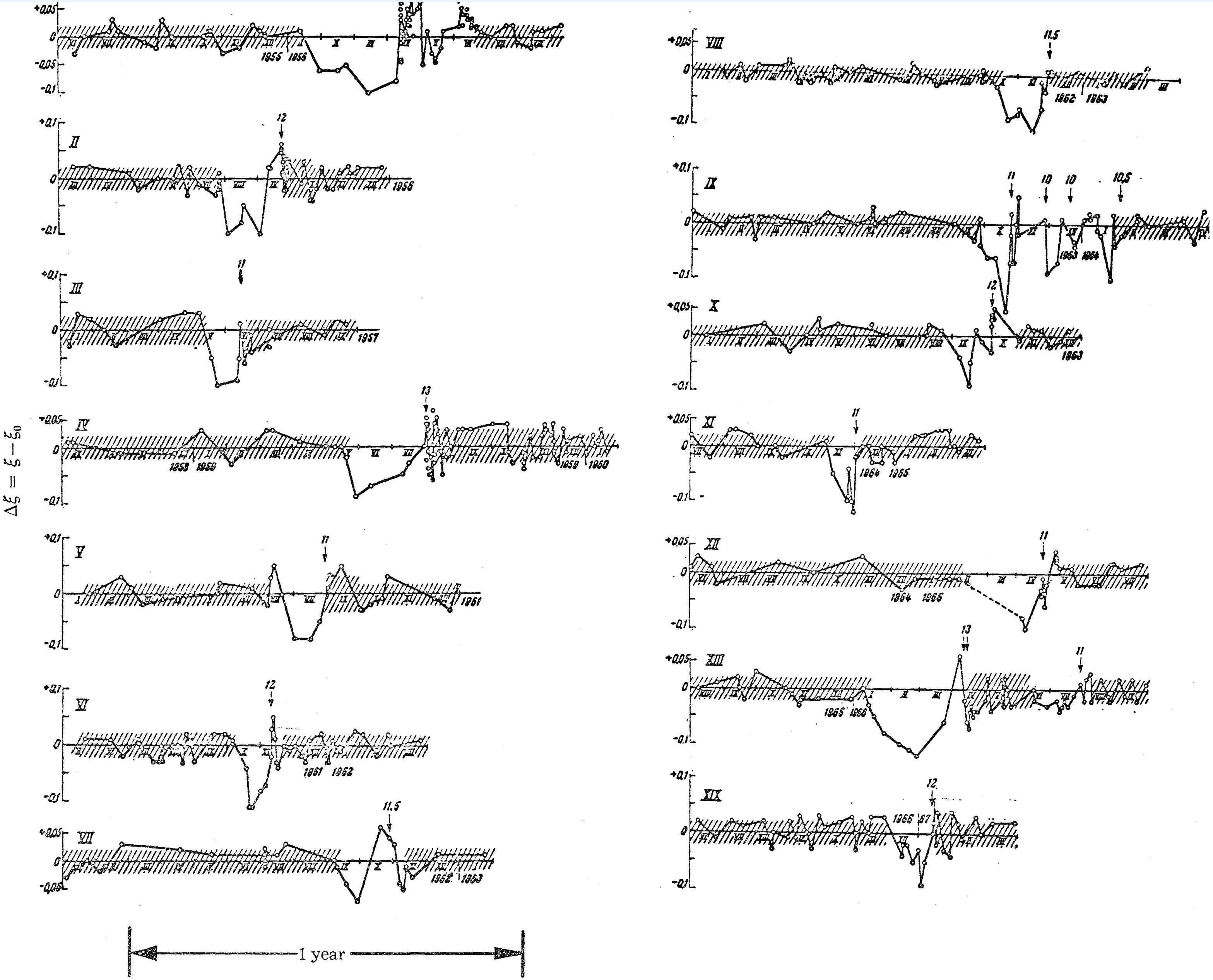

FIGURE 4. Deviation of the parameter $\xi = t_s/t_p$ from its normal value as a function of time, measured *in situ* for several events in Garm (reproduced from Semenov 1969). Values are given in $\Delta\xi = \xi - \xi_0$. Small arrows and numbers indicate earthquakes. Time is indicated by numerals (year) and roman numbers (month).

the well-known phenomenon of dilatancy. If the variations of ξ in Garm are caused by changes of rock properties, the inescapable conclusion is that dilatancy can take place *in situ*.

Assuming that rocks dilate before rupture in Garm, it remains to explain why ξ increases back to ξ_0 or more before sudden rupture occurs. An attractive possibility is the influence of ground water, which will tend to flow into the newly formed cracks. The dry cracks which first caused abnormally low ξ, slowly fill with ground water with two important consequences. First, ξ will markedly increase even beyond ξ_0 as cracks become saturated. Secondly, as pore pressure increases with time the shear strength of the rock diminishes, to the point where finally failure occurs.

Conclusions

We have considered three *in situ* processes which involve fluid flow in the crust: fault creep, aftershocks and dilatancy. Measurements of water level in wells suggest that creep events on the San Andreas fault are coupled with pore pressure changes. Readjustment of transient pore pressure, induced by large shallow earthquakes, possess the correct time constants and magnitudes to explain the occurrence of aftershocks. And finally, temporal changes of travel times in the Garm district imply that dilatancy may occur *in situ*. We conclude that the flow of pore fluids, *in situ*, particularly ground water, may play a major role in the mechanics of earthquakes, the generation of fore, main and aftershocks, creep and pre-rupture deformation.

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