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ON STRESS ACCUMULATION NEAR A CONTINUOUSLY SLIPPING FAULT IN A TWO-LAYER MODEL OF THE LITHOSPHERE

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Intoduction

It has been pointed out by Mukhopadhyay, Pal and Sen (1980) that Savage and Burford (1971, 1973), Spence and Turcott (1976) and others have reported continuous aseismic fault creep ranging from 1 to 6 cms per year in the central section of the San Andreas fault system. In order to study the effect of such creep on the stress accumulation in the fault region and other related phenomenon near active strike-slip faults, Mukhopadhyay, Pal and Sen (1980) considered a theoretical model consisting of a continuously creeping vertical, strike-slip fault in a visco-elastic half space, with uniform visco-elastic properties. This model has the advantage that it is possible to obtain exact solutions in closed form for the displacements and stresses, for various types of creeping dislocation across the fault and for arbitrary distributions of displacements and stresses at the time of commencement of the fault creep. In this connection, we note that the effective viscosity of the lithosphere would be expected to depend on the depth. The lower lithosphere, being at a much higher temperature and pressure would be expected to undergo greater creeping displacements without fracture. We also note that the observed faulting on shallow strikeslip faults, such as the San Andreas fault, is often estimated to extend to depths of about 10 to 15 Kms. This appears to indicate the accumulation of shear stress at greater depth does not reach sufficiently high values to cause further downward extension of the fracture. This phenomenon can be explained easily if we assume that the material of the lower lithosphere below the fault creeps under tectonic stresses without fracture. We again note that according to the results of the laboratory experiments on the deformation of rocks at high temperatures and pressures as reported by Griggs and Handin (1960), Heard (1976) and others, the rocks at depths below 15 Kms., at the strain rates of the order of 0.1 μ strain/year reported in the neighbourhood of the San Andreas fault by Prescott and Savage (1976), would be sufficiently ductile to undergo large creeping deformations without fracture. This effect is likely to be more pronounced in the asthenosphere. Finally, we note, that the tectonic forces causing accumulation of shear stress are likely to be more prominent in the lower lithosphere and asthenosphere. Keeping all these points in view we consider a two-layered model. The upper layer is taken to be elastic, resting on and in welded contact with a visco-elastic half-space. For this model, it is found that analytical solutions can be obtained for the displacements and stresses for certain types of creeping dislocation across a long vertical strike-slip fault in the upper layer. Although the layered model is more realistic, the form of the solution in this case is much more complicated than that of the solution obtained by Mukhopadhyay, Pal and Sen (1980)

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for a creeping strike-slip fault in a visco-elastic half-space and the solution obtained for the layered model is applicable to a more restricted class of creeping dislocations. But much useful information can still be obtained from the solution for a creeping fault in the layered model, as shown later in this paper. It is therefore hoped that the analysis of the layered model considered here would provide a useful addition to the results obtained by Mukhopadhyay, Pal and Sen (1980).

Formulation

We consider a vertical strike-slip fault F of depth D, situated in an elastic layer of thickness H $(0 < D \leq H)$. The layer rests on and is in welded contact with a visco-elastic half-space. The fault reaches the free surface of the elastic layer. We introduce cartesian coordinates (y_1, y_2, y_3) with the y_1-axis on the free surface along the trace of the fault, the y_3-axis vertically downwards and the y_2-axis along surface, perpendicular to the trace of the fault. The section of this theoretical model by the vertical y_2-y_3-plane is shown in Fig. 1. We assume that the length of the fault is large compared to its depth and we take the displacements and stresses to be independent of y_1 and depending on y_2, y_3 and the time t. This displacement component parallel to the y_1-axis in the layer and half-space are represented by ω and ω_1. These components are associated with the strike-slip fault movement. The relevant stress components associated with ω, ω_1 are represented by (τ_{12}, τ_{13}) for the layer and by (τ'_{12}, τ'_{13}) for the half-space. Since, the displacements and stresses are independent of y_1, (ω, ω_1), (τ_{12}, τ_{13}), $(\tau'_{12}\ \tau'_{13})$ are independent of the other component of displacement and stress. For the elastic layer the constitutive equations are taken to be

$$\left.\begin{aligned} \tau_{12} &= \mu_1 \frac{\partial \omega}{\partial y_2} \\ \text{and} \quad \tau_{13} &= \mu_1 \frac{\partial \omega}{\partial y_3} \\ (0 \leq y_3 &\leq H). \end{aligned}\right\} \tag{1}$$

The half-space is taken to be linearly visco-elastic and of the Maxwell type, with the constitutive equations

$$\left.\begin{aligned} \frac{1}{\mu_2} \cdot \frac{\partial}{\partial t} \tau'_{12} + \frac{\tau'_{12}}{\eta} &= \frac{\partial^2 \omega_1}{\delta t\, \partial y_2} \\ \frac{1}{\mu_2} \cdot \frac{\partial}{\partial t} \tau'_{13} + \frac{\tau'_{13}}{\eta} &= \frac{\partial^2 \omega_1}{\partial t\, \partial y_3} \\ (y_3 > H) \end{aligned}\right\} \tag{2}$$

We consider slow quasi-static aseismic deformation of the system when the inertial terms in the stress equations of motion are small and can be neglected as explained by Mukhopadhyay, Pal and Sen (1980). For such aseismic deformation, the stresses satisfy the relation :

$$\left.\begin{aligned} \frac{\partial \tau_{12}}{\partial y_2} + \frac{\partial \tau_{13}}{\partial y_3} &= 0, \quad 0 \leq y_3 \leq H \\ \frac{\partial \tau'_{12}}{\partial y_2} + \frac{\partial \tau'_{13}}{\partial y_3} &= 0, \qquad y_3 > H \end{aligned}\right\} \tag{3}$$

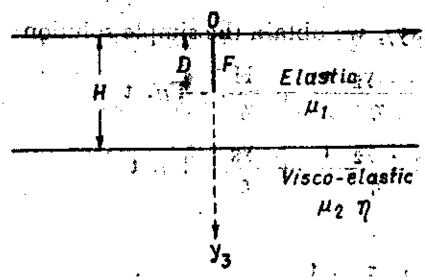

Fig. I Section of the model by the plane $y_1 = 0$.

We assume that the upper surface of the layer is stress-free, and the shear stress τ'_{12} in the half-space $y_3 > H$, at a large distance from the fault plane has a constant value τ_∞, maintained by tectonic forces, while the shear stresses near the fault plane may change from time to time due to fault movements. We also assume that the upper layer is in welded contact with half-space. Then the displacements and stresses would satisfy the following boundary conditions :

$$\left.\begin{aligned} \tau_{13} &= 0 \quad \text{at} \quad y_3 = 0, \\ \tau_{13} &= \tau'_{13} \quad \text{at} \quad y_3 = H, \\ \omega &= \omega_1 \quad \text{at} \quad y_3 = H, \\ \text{and} \quad \tau_{12}' &\to \tau_\infty \quad \text{as} \quad y_2 \to \infty \quad \text{in} \quad y_3 > H \end{aligned}\right\} \tag{4}$$

From (1), (2) and (3), we find that

$$\left.\begin{aligned} &\nabla^2\omega = 0 \text{ and } \frac{\partial}{\partial t}(\nabla^2\omega_1) = 0, \text{ which are satisfied if } \nabla^2\omega = 0, \quad 0 < y_3 < H \\ &\text{and} \quad \nabla^2\omega_1 = 0, \quad y_3 > H \end{aligned}\right\} \tag{5}$$

For the shear type deformations we consider for our model, we take the displacements to be anti-symmetric with respect to the plane $y_2 = 0$, so that

$$\omega(y_1, y_2, y_3) = -\omega(y_1, -y_2, y_3) \text{ and } \omega_1(y_1, y_2, y_3) = -\omega_1(y_1, -y_2, y_3)$$

Displacements and stresses in the absence of fault creep

In this case, we take the displacements and stresses to be continuous throughout the system. We measure the time from a suitable instant after which the conditions (1) — (5) become applicable, so that they are valid for $t > 0$. We assume that at $t = 0$, there are displacements and stresses $(\omega)_0$, $(\omega_1)_0$, $(\tau_{12})_0$, $(\tau_{13})_0$, $(\tau'_{12})_0$ and $(\tau'_{13})_0$ satsifying the relations (1) — (5).

We take Laplace transforms of (1) — (5) with respect to t. This gives a boundary value problem which can be solved easily in some cases. Inversion of the Laplace transform then gives the solution for the displacements and stresses. Since, $(\tau_{13})_0$ and $(\tau'_{13})_0$ satisfy (1) — (5), they have the same value at $y_3 = H$.

Let $\quad T_H(y_2) = [(\tau_{13})_0]\ y_3 = H$

$\quad\quad = [(\tau'_{13})_0]\ y_3 = H.$

If T_H (y_2) is a constant $= T'_H$ (say), we obtain the simple solution

$$\left.\begin{aligned}
\omega &= (\omega)_0 + \frac{\tau_\infty \cdot y_2 \cdot t}{\eta} + \frac{H}{\eta} \cdot T_H \cdot t \\
\omega_1 &= (\omega_1)_0 + \frac{\tau_\infty \cdot y_2 \cdot t}{\eta} + \frac{y_3}{\eta} \cdot T_H \cdot t \\
\tau_{13} &= (\tau_{13})_0 \\
\tau_{12} &= (\tau_{12})_0 + \frac{\mu_1 \cdot \tau_\infty \cdot t}{\eta} \\
\tau'_{13} &= (\tau'_{13})_0 \cdot e^{-\mu_2 t/\eta} + T_H \cdot (1 - e^{-\mu_2 t/\eta}) \\
\tau'_{12} &= (\tau'_{12})_0 \cdot e^{-\mu_2 t/\eta} + \tau_\infty \cdot (1 - e^{-\mu_2 t/\eta})
\end{aligned}\right\} \quad (6)$$

From (6) we find that for $t > 0$, the stress $\tau'_{13} \rightarrow 0$ as $t \rightarrow \infty$. The stress τ_{12} in the layer increases steadily as t increases and becomes large as $t \rightarrow \infty$, but τ'_{12} remains bounded and tends to τ_∞ as $t \rightarrow \infty$. This indicates that fracture or fault creep would be more likely to develop in the upper layer than the lower layer half-space. This is consistent with the observation that most of the major strike-slip faults appear to be confined to fairly shallow depths.

Displacements and stresses after the commencement of fault creep

The accumulation of the shear stress τ_{12} in the upper layer would be expected to lead eventually either to seismic fault movement or to aseismic fault creep. We consider here the case of aseismic fault creep and we now measure the time t from the instant at which the creep commences. The relations (1)—(5) are valid in this case. But there is a dislocation across the fault F which changes with time as long as the creep continues. The corresponding boundary conditions across the creeping fault are taken to be

$$\left.\begin{aligned}
&[\omega] = f(y_3) \cdot U(t) \qquad \text{for } t > 0, \\
&(y_2 = 0,\ 0 < y_3 < D;\ \ U(0) = 0)
\end{aligned}\right. \quad (7)$$

where $[\omega]$ is the discontinuity in ω across F. We take τ_{12}, τ_{13} to be continuous across F. To obtain solutions satisfying the relations (1)—(5), and the additional condition (7) across the fault, we try to obtain the solution in the form

$$\left.\begin{aligned}
\omega &= (\omega)_1 + (\omega)_2 \\
\tau_{12} &= (\tau_{12})_1 + (\tau_{12})_2 \\
\tau_{13} &= (\tau_{13})_1 + (\tau_{13})_2 \\
\omega_1 &= (\omega_1)_1 + (\omega_1)_2 \\
\tau'_{12} &= (\tau'_{12})_1 + (\tau'_{12})_2 \\
\tau'_{13} &= (\tau'_{13})_1 + (\tau'_{13})_2
\end{aligned}\right\} \quad (8)$$

where $(\omega)_1$, $(\tau_{12})_1, \ldots\ldots\ldots\ldots, (\tau'_{13})_1$ satisfy the relations (1)—(5) and are continuous throughout the system while $(\omega)_2$, $(\tau_{12})_2, \ldots\ldots\ldots, (\tau'_{13})_2$ satisfy the conditions (1)—(3), (5), (7) and the following conditions given by (9), which replace (4) :

$$\left.\begin{array}{lll} \tau_{13} = 0 & \text{at} & y_3 = 0 \\ \tau_{13} = \tau'_{13} & \text{at} & y_3 = H \\ \tau'_{13} \to 0 & \text{as} & y_3 \to \infty \end{array}\right\} \tag{9}$$

and

The solutions for $(\omega)_1$, $(\tau_{12})_1$, ..., $(\tau'_{13})_1$ are exactly similar to the solutions (6). The only difference is that $(\omega)_0$, $(\omega_1)_0$, ..., $(\tau'_{13})_0$ in (6) are replaced by $(\omega)_p$, $(\omega_1)_p$, ..., $(\tau'_{13})_p$ which are the values of the functions $(\omega)_1$, $(\omega_1)_1$, ..., $(\tau'_{13})_1$ at $t = 0$. To obtain the solutions for $(\omega)_2$, ..., $(\tau'_{13})_2$, we take Laplace transforms of (1) — (3), (5), (7) and (9) with respect to t. This gives rise to a boundary value problem for $(\bar{\omega})_2$, ..., $(\bar{\tau}'_{13})_2$ which are the Laplace transforms of $(\omega)_2$, ..., $(\tau'_{13})_2$. This boundary value problem can be solved by using a suitable modification of a Green's function technique developed by Maruyama (1966) and Rybicki (1971). This technique has been explained by Mukhopadhyay Pal and Sen (1980). On inverting the Laplace transforms with respect to t, we obtain solutions for $(\omega)_2$, ..., $(\tau'_{13})_2$. Finally, from equation (9), we obtain ω, ω_1, ..., τ'_{13}. It is then clear that they would satisfy (1) — (5) and (7). We note, however, that this method enables us to calculate exact solutions of the displacements and stresses in the form of convergent series of reasonably simple form only for certain simple distributions of the dislocation across the fault. In particular, if $f(y_3)$ = a constant or a polynomial in y_3, and $U(t) = Vt$, where V is a constant, in (7), then the solutions for the displacements and stresses are obtained in the form of convergent series. In the case $f(y_3)$ = constant, when the dislocation across F is independent of the depth it is found as in Mukhopadhyay, Pal and Sen (1980), that there is a singularity of the stresses in the neighbourhood of the lower end of the fault. Analytical investigations similar to those in Mukhopadhyay, Pal and Sen (1980) show that the displacements and stresses would be bounded everywhere including the tip of the fault, if the following conditions were satisfied :

(a) $f(y_3)$ and $f'(y_3)$ are continuous in $0 \leqslant y_3 \leqslant D$.

(b) $f''(y_3)$ is continuous in $0 < y_3 < D$, $y_3^m f''(y_3) \to$ 0 or to a finite limit, as $y_3 \to 0 + 0$, $(D - y_3)^n f''(y_3) \to$ 0 or to a finite limit as $y_3 \to D - 0$, where m, n are constants, each < 1.

(c) $f'(0) = 0 = f'(D) = f(D)$.

The physical implication of these conditions is that the creeping dislocation vary smoothly over the fault and approaches the value zero with sufficient smoothness near the tip of the fault. A simple type of dislocation satisfying the conditions given above is

$$[\omega] = \frac{V.\,t.\,(y_3^2 - D^2)^2}{D^4} \tag{10}$$

In this case, the displacements and stresses can be obtained in the form of convergent series. We find that for $0 \leqslant y_3 \leqslant H$,

$$\omega = (\omega)_p + \frac{\tau_\infty \cdot y_2 \cdot t}{\eta} + \frac{Vt}{2\pi} \cdot \phi_1(y_2, y_3) + \frac{Vt}{2\pi} \cdot \sum_{m=1}^{\infty} \left(\frac{a}{b}\right)^m \cdot \psi_m(y_2, y_3)$$

$$+ \frac{1}{2\pi} \cdot \sum_{m=1}^{\infty} \left(\frac{a}{b}\right) \cdot \sum_{r=1}^{m} \binom{m}{r} \cdot b_1^r \cdot \psi_m(y_2, y_3) \cdot Q_r(t) \tag{11}$$

where

$$Q_r(t) = \frac{Vt}{a_1^r} \cdot \left[1 - e^{-a_1 t} \cdot e_{r-1}(a_1 t)\right] - \frac{Vr}{a_1^{r+1}} \cdot \left[1 - e^{-a_1 t} \cdot e_r(a_1 t)\right] \tag{12}$$

$$\tau_{12} = (\tau_{12})_p + \frac{\mu_1 \cdot \tau_\infty \cdot t}{\eta} + \frac{\mu_1 \cdot Vt}{2\pi} \cdot \Big[\phi_2(y_2, y_3) + \sum_{m=1}^{\infty} \left(\frac{a}{b}\right)^m \cdot \psi_{m1}(y_2, y_3)\Big] + \frac{\mu_1}{2\pi} \cdot \sum_{m=1}^{\infty} \left(\frac{a}{b}\right)^m \cdot \sum_{r=1}^{m} \binom{m}{r} \cdot b_1^r \cdot Q_r(t) \cdot \psi_{m1}(y_2, y_3)$$

$$(\text{for } 0 \leq y_3 \leq H) \tag{13}$$

In equations (11) – (13),

$$a = \frac{\mu_1}{\mu_2} - 1\,; \quad b = \frac{\mu_1}{\mu_2} + 1\,; \quad a_1 = \frac{\mu_2}{\eta(1+S)}\,;$$

$$b_1 = \frac{2\mu_1 \cdot S^2}{\eta(1-S^2)}\,; \quad S = \mu_2/\mu_1\,;$$

$$\phi_1(y_2, y_3) = \frac{2 \cdot y_2 \cdot (3y_3^2 - y_2^2)}{D^3} - \frac{10\, y_2}{3D} + \frac{2\, y_2 \cdot y_3 \cdot (y_3^2 - y_2^2 - D^2)}{D^4} \cdot \log_e \frac{(D-y_3)^2 + y_2^2}{(D+y_3)^2 + y_2^2} + \frac{y_2^4 - y_2^2 \cdot (6\, y_3^2 - 2\, D^2) + (y_3^2 - D^2)^2}{D^4} \times \left[\tan^{-1}\left(\frac{D-y_3}{y_2}\right) + \tan^{-1}\left(\frac{D+y_3}{y_2}\right)\right];$$

$$\psi_m(y_2, y_3) = \phi_1(y_2, p) + \phi_1(y_2, q),$$

where $p = 2^m H + y_3,\ q = 2^m H - y_4,$

$$\phi_2(y_2, y_3) = -\Big[\frac{6 \cdot (y_3^2 - y_2^2)}{D^3} - \frac{10}{3D} + \frac{2\, y_3 \cdot (y_3^2 - 3\, y_2^2 - D^2)}{D^4} \log_e \frac{(D-y_3)^2 + y_2^2}{(D+y_3)^2 + y_2^2} + \frac{4\, y_2^2 \cdot y_3 \cdot (y_3^2 - y_2^2 - D^2)}{D^4} \left\{\frac{1}{(D-y_3)^2 + y_2^2} - \frac{1}{(D+y_3)^2 + y_2^2}\right\} + \frac{4y_2^3 - 2y_2 \cdot (6\, y_3^2 - 2\, D^2)}{D^4} \left\{\tan^{-1}\left(\frac{D-y_3}{y_2}\right) + \tan^{-1}\left(\frac{D+y_3}{y_2}\right)\right\}\Big];$$

$$\psi_{m1}(y_2, y_3) = \phi_2(y_2, p) + \phi_2(y_2, q),$$

where $p = 2^m H + y_3,\ q = 2^m H - y_3,$

$$e_r(a_1 t) = 1 + a_1 t + \frac{(a_1 t)^2}{2!} + \ldots\ldots\ldots + \frac{(a_1 t)^r}{r!}$$

$$(r \geq 1). \tag{14}$$

We note that $(\omega)_p$ and $(\tau_{12})_p$ are, in this case, the values of ω and τ_{12} at $t = 0$, when the fault creep commences. The expression for ω_1 is similar to the expression for ω and the expressions for τ_{13}, τ'_{12}, τ'_{13} are similar to the expressions for τ_{12} except for the fact that there is no term of the form $\tau_\infty \cdot y_2 . t/\eta$, which is present in the expression of τ_{12} and increases monotonically with time. We have not given the detailed expressions for ω_1, τ_{13}, τ'_{12}, τ'_{13}, since the displacement ω alone can be observed on the free surface and the shear stress component τ_{12} is expected to control the strike-slip movement of the fault, so that ω_1 and τ_{12} are the quantities which are of primary importance in this investigation.

Convergence and computation of the series

On examining the behaviour of the mth term of the series giving the displacements and stresses, in the case of any creeping dislocation $[u] = Vt. f(y_3)$ across the fault, where $f(y_3)$ satisfies the conditions (a), (b) and (c) for boundedness of displacement and stress, it is found that as $m \to \infty$, the mth term is of the order $0(1/m^3. (a/b)^m)$ or $0(1/m^3)$. Since $|a/b| < 1$, it follows that all the series converge. On actually computing the terms and the approximate sum of the series after assigning suitable values to the parameters of the model, viz., μ_1, μ_2, η, D, H, τ_∞ ,V (given later in this paper), it is found that the convergence is fairly rapid for the creeping dislocation we have considered. It is found that, to obtain the sum of the series correct to three significant figures (i.e., with error less than 0.1 percent), it is sufficient to compute the *first 20 to 30 terms of the series to secure the appropriate order of convergence*, the number of terms required depend to some extent on the values assigned to the parameters μ_1, μ_2, η, D, H, τ_∞ and V. Although the terms of the series have been obtained explicitly in closed form, the expressions for the terms are fairly long. Hence the computation of the sum of the series has been carried out on an IBM 1130 type computer, after developing a suitable programme in FORTRAN IV for the computation.

Discussion of the results and conclusions

We now study the effect of fault creep in our model on the changes in the surface strain near the fault and the shear stress τ_{12} in the neighbourhood of the fault. In choosing suitable values for the parameters in our model, we keep in view the San Andreas Fault and take D = 10 kms. We take $\mu_1 = \mu_2 = 3.78 \times 10^{11}$ dynes/cm^2, which is the value of the lithospheric rigidity given by Aki (1967). We also note that Cathles (1975) gives estimates of the effective viscosity of the lower lithosphere and asthenosphere in the range 10^{20} to 10^{22} poise. These estimates have been obtained from a comparison of theoretical results of observational data on a post-glacial uplift of Fennoscandia in Canada. We therefore choose values of η for the half-space in the range 10^{20} poise to 5×10^{20} poise. We also note that in our model the creep velocity across the fault on the surface $y_3 = 0$ is V. Since, Spence and Turcotte (1973) report creep velocities in the range 1 to 6 cms/year in the central section of the San Andreas fault, we consider values of V in the range 0 to 6 cms/year. We first calculate the accumulation of shear strain on the surface near the fault. For this, we first calculate the strain component $e_{12} = \frac{\partial \omega}{\partial y_2}$ at $y_2 = 0$, $y_3 = 0$, near the fault on the surface. We net calculate $E_{12} = \{e_{12} - (e_{12})_0\} \times 10^6$, where $(e_{12})_0$ is the value of e_{12} at $t = 0$. E_{12} depends on the time t, and is obtained in the form of a convergent series. This is evaluated numerically, taking D = 10 kms., H = 100 kms., $\eta = 5 \times 10^{20}$ poise, $\mu = 3.78 \times 10^{11}$ dynes/-

cm^2 and $\tau_\infty = 10$ bars. This choice of the parameter τ_∞ gives a value of E_{12}, which gives the rate of increase of $e_{12} - (e_{12})_0 \approx 0.6 \times 10^{-6}$ year, for the case $V = 0$. This is of the same order as the rate of strain accumulation in the neighbourhood of the northern locked section of the Sen Andreas fault, as reported by Savage and Burford (1971, 1973). We therefore adopt this value of τ_∞, and calculate E_{12} for different values of V, for $t > 0$. The computations have been carried out on an IBM 1130 computer. The results are shown in Fig. II. It is seen that, for larger values of V, the rate of accumulation of shear strain on the surface becomes smaller, and when the value of V, the velocity of creep, reaches 0.8 cms/-year, the strain accumulation becomes virtually zero.

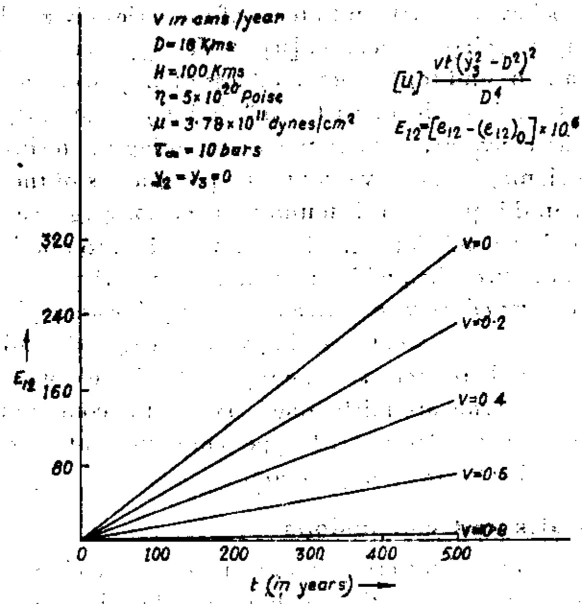

Fig. II Changes of the surface shear strain with time.

We next calculate the average value of the shear stress over the fault, given by

$$(\tau_{12})_A = \frac{1}{D} \int_0^D (\tau_{12})_{y_2=0}\, dy_3$$

$(\tau_{12})_A$ is obtained easily from (13), in the form of a convergent series. We use the same values of the model parameters, and compute $(\tau_{12})_A$. In the first case, we take $(\tau_{12})_0 = 0$, where $(\tau_{12})_0$ is the value of $(\tau_{12})_A$ at $t = 0$. The results for this case are shown in Fig. III. It is found that the rate of increase of $(\tau_{12})_A$ is smaller for larger values of V, and when V reaches the value of 2.9 cms/year, the rate of accumulation of the average shear stress $(\tau_{12})_A$ becomes nearly zero. Thus, the fault creep prevents accumulation of shear stress in this case. We have also considered the case in which $(\tau_{12})_0 = 40$ bars. The corresponding results are shown in Fig. IV. We again find that, for larger values of V, the rate of increase of $(\tau_{12})_A$ becomes smaller. When V exceeds 3 cms/year, there is a steady decrease in the value of τ_{12} with time. For $V = 4.0$ cms/year, $(\tau_{12})_A$ approaches the value zero in about 300 years. Thus, in this case, there is continuous aseismic release of shear stress due to fault creep. We

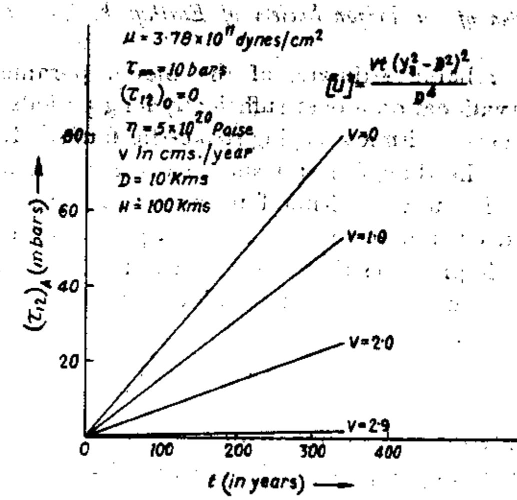

Fig. III Changes of the average shear stress across the fault with time t, where the average stress at t = 0 is zero.

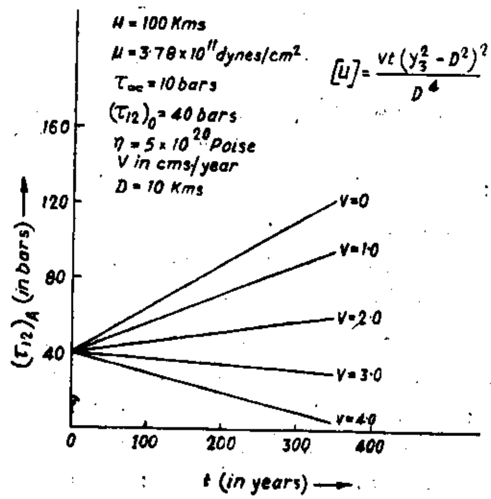

Fig. IV Changes of the average shear stress across the fault with time t, where the average shear stress at t = 0 is 40 bars.

note here that the creep velocities for which complete stress release can be achieved are of the same order (i.e., a few cms/year) as the aseismic creep velocities reported in central section of the San Andreas fault, by Spence and Turcotte (1976). These results are similar to those obtained by Mukhopadhyay, Pal and Sen (1980). But one important difference is that a much smaller value of τ_∞, the shear stress maintained by tectonic forces in the half space far away from the fault, gives the same rates of accumulation of stress and strain as those obtained by Mukhopadhyay, Pal and Sen (1980). In fact, in the layered model we consider here, a comparitively small shear stress, maintained in the half space away from the fault, leads to the accumulation of much larger shear stresses near the fault in the upper layer, unless the aseismic creep velocity is sufficiently large.

We note that, more reliable estimates of the model parameters may be obtained if reliable and detailed observational data over sufficiently long periods of time are available for the ground deformation near aseismic creeping strike-slip faults. It may then be possible to obtain reliable estimates for the changes in the shear stress near the fault. This may enable us to assess the possibility of a sudden seismic fault movement, and to estimate the magnitude of such a movement, if and when it occurs. It is also hoped that this model might add to our insight into the tectonic processes that lead to accumulation of large shear stresses in the upper lithosphere, leading to sudden seismic fault movement, and the release of such shear stresses by aseismic fault creep.

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