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Long Dip-Slip Fault in a Viscoelastic Half Space Model of the Lithosphere

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Abstract Most of the earthquake faults in North- East India, China, mid Atlantic-ridge, the Pacific seismic belt and Japan are found to be predominantly dip-slip in nature. In the present paper a dip -slip fault is taken to be situated in a viscoelastic half space representing the upper lithospheric region of the Earth. A movement of the dip -slip nature across the fault occurs when the accumulated stress due to various tectonic reasons e.g. mantle convection etc, exceeds the local friction and cohesive forces across the fault. The movement is assumed to be creeping in nature, expressions for displacement, stress and strains are obtained by suitable mathematical methods. A detailed study of these expressions may give some ideas about the nature of stress accumulation in the system, which in turn will be helpful in formulating an earthquake prediction programme.

Keywords Aseismic Period, Dip-Slip Fault, Earthquake Prediction, Mantle Convection, Plate Movements, Stress Accumulation, Tectonic Process, Viscoelastic Half Space

1. Introduction

It is the observational fact that while some faults are strike slip (finite or infinite in length) in nature ,there are faults (e.g: Sierra Nevada/Owens valley: Basin and Range faults, Rocky Mountains, Himalayas, Atalanti fault of central Greece-a steeply dipping fault with dip 60, 80(deg)) where the surface level changes during the motion i.e. the faults are dip-slip in nature.

A pioneering work involving static ground deformation in elastic media were initiated by ([37-38]). ([4-7]), ([15-16]). [27],[34],[35] did a wonderful work in analyzing the displacement, stress and strain for dip-slip movement. Later some theoretical models in this direction have been formulated by a number of authors like[2],[31],[23],[32],[41][24],[22],[33],[11],[3],[29], ([18-21]),[8],[26],[28],[12],[30],[44],[42],[43],[25],[45],[36] has discussed various aspects of fault movement in his book. Reference;[13] has discussed stress accumulation near buried fault inlithosphere-asthenosphere system. The work of[10] can also be mentioned.

In most of these works the medium were taken to be elastic and /or viscoelastic, some authors preferred layered model with elastic layer(s) over elastic or viscoelastic half space.

In the present case we consider a long dip-slip fault situated in a viscoelastic half space which reaching upto the free surface. The medium is taken to be under the influence of tectonic forces due to mantle convection or some related phenomena. The fault is assumed to undergo a creeping movement when the stresses in the region exceed certain threshold values.

In our paper, we consider a viscoelastic half space to represent the upper part of the lithosphere-asthenosphere system, with constant rigidity ($2{:}0\times10^5$ Mpa) and viscosity ($10^{20}-10^{21}$ pa:s) following the observational data mentioned by[9],[14]. It may be stated that a thin elastic layer overlying an elastic/viscoelastic half space is likely to be a more preferable model for the system. But, numerical computational works indicate that the presence of a layer does not lead to any significant qualitative changes in the nature of the stress and strain accumulation in the model, only a small 10 percent quantitative change were observed. Analytical expressions for displacements stresses and strains are obtained both before and after the fault movement using appropriate mathematical technique involving integral transformation, Green's function. Numerical computational works have been carried out with suitable values of the model parameters and the nature of the stress and strain accumulation in the medium have been investigated.

2. Formulation

We consider a long dip-slip fault F, width D situated in a viscoelastic half space of linear Maxwell type.

A Cartesian coordinate system is used with a suitable

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point O on the strike of the fault as the origin, the strike of the fault along the Y_1 axis and Y_2 axis is as shown in Figure 1, and Y_3 axis pointing downwards. We choose another coordinate system Y_1', Y_2' and Y_3' axes as shown in Figure. 1 below, so that the fault is given by $F : (y_2' = 0, 0 \le y_3' \le D)$. Let θ be the dip of the fault F.

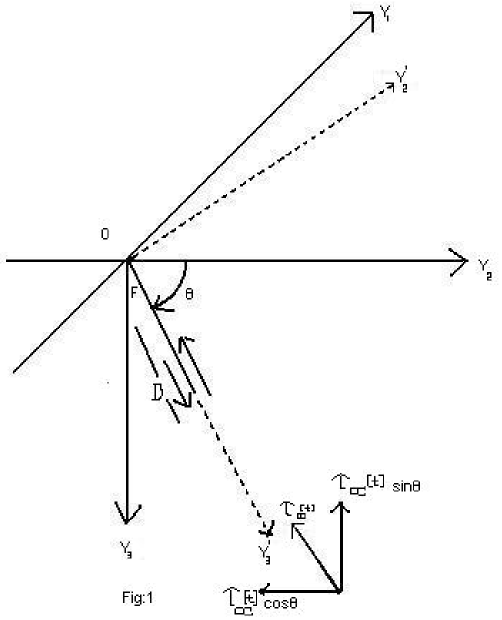


Figure 1. Section of the model by the plane y_1=0

For a viscoelastic Maxwell type medium the constitutive equations are taken as:

$$\left(\frac{1}{\eta}+\frac{1}{\mu}\frac{\partial}{\partial t}\right)\tau_{22}=\frac{\partial}{\partial t}(e_{22})=\frac{\partial}{\partial t}\left(\frac{\partial v}{\partial y_2}\right) \quad (1.1)$$

$$\left(\frac{1}{\eta}+|\frac{1}{\mu}\frac{\partial}{\partial t}\right)\tau_{23}=\frac{\partial}{\partial t}(e_{23})=\frac{1}{2}\left(\frac{\partial v}{\partial y_3}+\frac{\partial v}{\partial y_2}\right) \quad (1.2)$$

$$\left(\frac{1}{\eta}+\frac{1}{\mu}\frac{\partial}{\partial t}\right)\tau_{33}=\frac{\partial}{\partial t}(e_{33})=\frac{\partial}{\partial t}\left(\frac{\partial w}{\partial y_3}\right) \quad (1.3)$$

Where η is the effective viscosity and μ is the effective rigidity of the material.

The stresses satisfy the following equations: (assuming quasistatic deformation for which the inertia terms are neglected).

$$\frac{\partial}{\partial y_1}(\tau_{11})\frac{\partial}{\partial y_2}(\tau_{12})+\frac{\partial}{\partial y_3}(\tau_{13})=0 \quad (1.4)$$

$$\frac{\partial}{\partial y_1}(\tau_{21})\frac{\partial}{\partial y_2}(\tau_{22})+\frac{\partial}{\partial y_3}(\tau_{23})=0 \quad (1.5)$$

$$\frac{\partial}{\partial y_1}(\tau_{31})\frac{\partial}{\partial y_2}(\tau_{32})+\frac{\partial}{\partial y_3}(\tau_{33})=0 \quad (1.6)$$

Where $((-\infty < y_2 < \infty), y_3 \ge 0, t \ge 0)$ (Assuming the body forces do not change during the fault movement).

The boundary conditions are taken as, with t=0 representing an instant when the medium is in aseismic state:

$$\tau_{22}(y_2,y_3,t)=\tau_\infty(t)cos\theta \text{ as}$$
$$|y_2|\to\infty, y_3 \ge 0, t>0 \quad (1.7)$$

On the free surface

$$y_3=0,(-\infty<y_2<\infty,t\ge 0)$$
$$\tau_{23}(y_2,y_3,t)=0 \quad (1.8)$$
$$\tau_{33}(y_2,y_3,t)=0 \quad (1.9)$$

Also, as $y_3\to\infty(-\infty<y_2<\infty,t\ge 0)$

$$\tau_{23}(y_2,y_3,t)=0 \quad (1.10)$$
$$\tau_{33}(y_2,y_3,t)=\tau_\infty(t)sin\theta \quad (1.11)$$

[Where $\tau_\infty(t)$ is the shear stress maintained by mantle convection and other tectonic phenomena far away from the fault].

The initial conditions are: Let$(v)_0$; $(w)_0$,$(\tau_{ij})_0$ and $(e_{ij})_0$ i,j=2,3 be the value of (v); (w),(τ_{ij}) and (e_{ij}) at time t=0 which are functions of y_2, y_3 and satisfy the relations (1.1)-(1.11).

2.1. Solutions in the Absence of Any Fault Dislocation

The boundary value problem given by (1.1)-(1.11), can be solved(as shown in the Appendix-1) by taking Laplace transform with respect to time 't' of all the constitutive equations and the boundary conditions. On taking the inverse Laplace transform we get the solutions for displacements, stresses as:

$$v(y_2,y_3,t)=(v)_0+(cos\theta/\mu)\times$$
$$\times\left[(\tau_\infty(t)-\tau_\infty(0)+\left(\frac{\mu}{\eta}\right)\int_0^t\tau_\infty(\tau)d\tau\right]$$
$$w(y_2,y_3,t)=(w)_0+(y_3/\mu)\times$$
$$\times\left[(\tau_\infty(t)-\tau_\infty(0)+\left(\frac{\mu}{\eta}\right)\int_0^t\tau_\infty(\tau)d\tau\right]sin\theta$$
$$\tau_{22}-\tau_\infty(t)cos\theta-[\tau_\infty(0)cos\theta-(\tau_{22})_0]e^{-(\mu/\eta)t}$$
$$\tau_{23}=(\tau_{23})_0e^{-(\mu/\eta)t}$$
$$\tau_{33}=[\tau_\infty(t)-\tau_\infty(0)e^{-(\mu/\eta)t}]sin\theta \quad (A)$$

From the above solution we find that τ_{22} increases with time and tends to $\tau_\infty(t)\cos\theta$ as t tends to ∞, while τ_{23}, tends to zero, but τ_{33} tends to $\tau_\infty(t)\sin\theta$. We assume that the geological conditions as well as the characteristic of the fault is such that when the stress-component $\tau_{2'3'}$ reaches some critical value, say $\tau_c < \tau(t)\cos\theta$ the fault F starts creeping.

For bounded stress and strains, the creep function should satisfy the following conditions as discussed by Mukhopadhyay et al.(1983,a).

(C_1) Its value will be maximum on the free surface.

(C_2) The magnitude of the creep will decrease with y_3 as we move downwards and ultimately tends to zero near the lower edge of the fault.

$$(y_2'=0, y_3'=D)$$

If $g(x'_3)$ be the creep function, it should satisfy the above conditions.

2.2. Solutions after the Fault Movements

We assume that after a time T_1, the stress component $\tau_{2'3'}$ (which is the main driving force for the dip-slip motion of the fault) exceeds the critical value τ_e, and the fault F starts creeping, characterized by a dislocation across the fault given by (Appendix-2).

$$[(w)]_F = w_1(t_1)g(y_3)H(t_1)$$

Where, $H(t_1)$ is the Heaviside function and $[(w)]_F$ = the discontinuity of w across F given by:

$$[(w)]_F = \lim_{(y_2' \to 0+)}(w) - \lim_{(y_2' \to 0-)}(w)$$
$$(y_2' = 0, 0 \le y_3' \le D)$$

We solve the resulting boundary value problem by modified Green's function method following[16],[31] and correspondence principle (As shown in the Appendix 2) and get the solution for displacements, stresses and strain as :

$$v(y_2, y_1, \theta, t) = (v)_0 + (\cos\theta/\mu)$$
$$\left[\tau_\infty(0)) + (\mu/\eta)\int_0^t \tau_\infty(\tau)d\tau\right]$$
$$w(y_2, y_3, \theta, t) = (w)_0 + (y_3/\mu)$$
$$\left[\tau_\infty(0)) + (\mu/\eta)\int_0^t \tau_\infty(\tau)d\tau\right]\sin\theta$$
$$+ [H(t-T_1)/(2\times\pi)]w_1(t_1)e^{(-\mu/\eta)t}\int_0^D g(x_3')$$
$$[A_1/B_1 + C_1/D_1]dx_3'$$

Where,

$$A_1 = (y_2)\sin\theta - (y_3)\cos\theta,$$
$$B_1 = [(x_3'^2) - 2(x_3')(y_3\cos\theta + y_3\sin\theta) +$$
$$+ (y_3^2) + (y_3^2)]$$
$$\tau_{22}(y_2, y_3, \theta, t) = \tau_\infty(t)\cos\theta - (\tau_\infty(0)\cos\theta -$$
$$-(\tau_{22})_0)e^{(-\mu/\eta)t}$$
$$\tau_{23}(y_2, y_3, \theta, t) = (\tau_{23})_0)e^{(-\mu/\eta)} +$$
$$+ [H(t-T_1)/(2\times\pi)]w_1(t_1)e^{(-\mu/\eta)}]\int_0^D$$
$$g(x_3')[A_2/B_2 + C_2/D_2]dx_3'$$

Where,

$$A_2 = (x_3'^2 - 2x_3'(y_2\cos\theta + y_3\sin\theta) + y_2^2 + y_3^2)$$
$$\sin\theta - (y_2\sin\theta - y_3\cos\theta)(2y_2 - 2x_3'\cos\theta)$$
$$B_2 = x_3'^2 - 2x_3'(y_2\cos\theta + y_3\sin\theta) + y_2^2 + y_3^2,$$
$$C_2 = (x_3'^2 - 2x_3'(y_2\cos\theta - y_3\sin\theta) + y_2^2 + y_3^2)$$
$$\sin\theta - (y_2\sin\theta + y_3\cos\theta)(2y_3 - 2x_3'\cos\theta)$$
$$D_2 = (x_3'^2 - 2x_3'(y_2\cos\theta - y_3\sin\theta) + y_2^2 + y_3^2)$$
$$\tau_{33}(y_2, y_3, \theta, t) = [\tau_\infty(t) - \tau_\infty(0)e^{-(\mu/\eta)t}]\sin\theta$$
$$+ [H(t-T_1)/(2\pi)]w_1(t_1)e^{(-\mu/\eta)}]\int_0^D$$
$$g(x_3')[A_3/B_3 + C_3/D_3]dx_3'$$

Where,

$$A_3 = (x_3'^2 - 2x_3'(y_2\cos\theta + y_3\sin\theta) + y_2^2 + y_3^2)$$
$$(-\cos\theta) - (y_2\sin\theta - y_3\cos\theta)(2y_3 - 2x_3'\cos\theta)$$
$$B_3 = x_3'^2 - 2x_3'(y_3\cos\theta + y_3\sin\theta) + y_2^2 + y_3^2$$
$$C_3 = (x_3'^2 - 2x_3'(y_2\cos\theta - y_3\sin\theta) + y_2^2 + y_3^2)\times$$
$$\times\cos\theta - (y_2\sin\theta + y_3\cos\theta)(2y_3 - 2x_3'\sin\theta)$$
$$D_3 = x_3'^2 - 2x_3'(y_2\cos\theta - y_3\sin\theta) + y_2^2 + y_3^2$$
$$e_{23}(y_2, y_3, \theta, t) = \left(\frac{1}{2}\right)(e_{23})_0 + H(t-T_1)/(2\pi)$$
$$w_1(t_1)\phi_2(y_2, y_3, \theta, t$$
$$e_{33}(y_2, y_3, \theta, t) = (e_{33})_0 + (1/\mu)[(\tau_\infty(t) - \tau_\infty(0))/$$
$$+(\mu/\eta)\int_0^t \tau_\infty(\tau)d\tau]\sin\theta + H(t-T_1)/(2\times\pi)$$
$$w_1(t_1)\phi_3(y_2, y_3, \theta, t)$$

Where, $\phi_2(y_2, y_3, \theta, t), \phi_3(y_2, y_3, \theta, t)$ are given in Appendix 2. (B)

3. Numerical Computations

Following[1] and recent studies on rheological behavior of crust and upper mantle by[9],[14] the values of the model parameters are taken as :

$\mu = 2\times10^{11}$ Dynes/cm^2

$\eta = 3\times10^{20}$ Poise

D = Depth of the fault = 10 km.,[nothing that the depth of all major earthquake faults are in between 10-15 km].

$\tau_\infty(t)$ = constant = $\tau_\infty = 2\times10^8$ dynes/cm^2 (200 bars),[post seismic observations reveal that in most of the cases, stress released in major earthquake are of the order of 200 bars or less, in extreme cases, it may be 400 bars.]

$(\tau_{23})_0 = 5\times10^7$ Dyne/cm^2 (50 bars)

and $\tau_\infty(0) = 0$

We take the creep function $g(x_3') - W[(x_3')^2 - D^2)^2]/(D)^4$, with W = 1 cm/year, satisfying the conditions stated in $(C_1)-(C_2)$.

We now compute the following quantities:

$$W_1(y_2, y_3, \theta, t) = w(y_2, y_3, \theta, t) - (w)_0 - (y_3/\mu)$$
$$[(\tau_\infty(t) - \tau_\infty(0)) + (\mu/\eta)\int_0^t \tau_\infty(\tau)d\tau]\sin\theta \quad (2.1)$$

$$t_{2'3'}(y_2 y_3, \theta, t) = \tau_{23}(y_2, y_3, \phi, t) - (\tau_{23})_0 e^{(-\mu/\eta)t} \quad (2.2)$$

$$t_{3'3'}(y_2 y_3, \theta, t) = [\tau_{23}(t) - \tau_\infty(0)e^{-(\mu/\eta}]\sin\theta \quad (2.3)$$

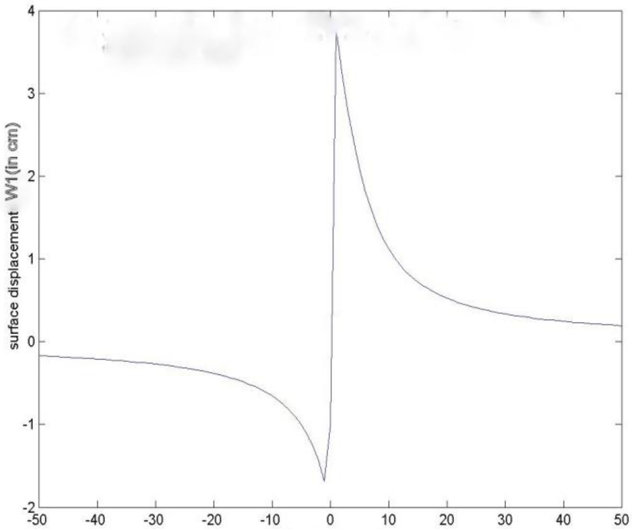
$$E_{23}(y_2, y_3, \theta, t) = [e_{23}(y_2, y_3, \theta, t) - \left(\frac{1}{2}\right)(e_{23})_0]\times10^6 \quad (2.4)$$

$$E_{33}(y_2, y_3, \theta, t) = e_{33}(y_2, y_3, \theta, t) - (e_{33})_0 - (y_3/\mu)$$
$$[(\tau_\infty(t) - \tau_\infty(0)) + \left(\frac{\mu}{\eta}\right)\int_0^t \tau_\infty(\tau)d\tau]\sin\theta \tag{2.5}$$

where τ_{23}, e_{23}, e_{33} are given by (B).

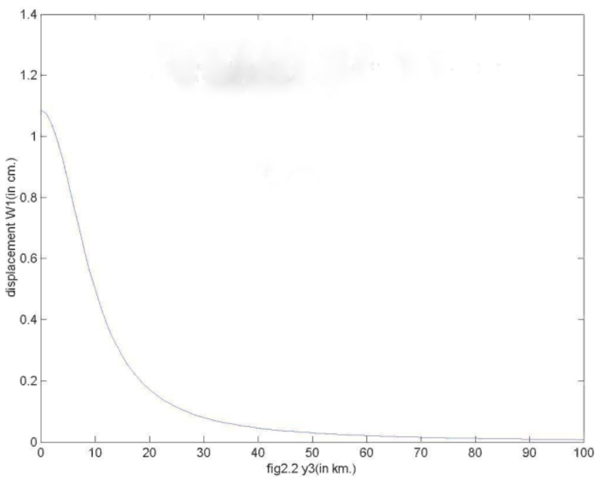
4. Results and Discussion

(A) Variation of vertical component of displacement due to creep across the fault after t=1 year.



y_2 (in km.)

Figure 2. Variation of the vertical component of surface displacement W_1 with y_2 for y_3=0, t=1 year, θ=45 (in deg) due to the fault movement



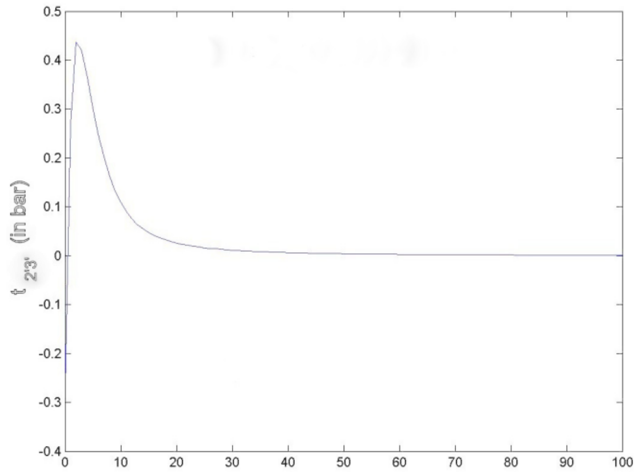
y_3 (in km.)

Figure 3. Variation of the vertical component of surface displacement W_1 with y_3 for y_2=5 km., t=1 year, θ=90 (in deg) due to the fault movement

Equation (2.1) gives us the vertical component of displacement at (y_2, y_3) due to the movement across the fault for different dip angle θ and at different time after the fault movement. We take t=1 year. In Figure 2 the graph shows the nature of surface displacement W_1 against y_2, the distance from the strike of the fault with θ=45 (in deg). It is observed that the displacement are in opposite directions across the strike of the fault. Their magnitudes gradually decrease and tend to zero as we move away from the fault. This is quite expected as the effect of the fault movement gradually die out with distance. The sudden changes of W_1 near y_2 = 0 is due to the dip-slip motion along the fault. This is in good conformity with the results shown in[36]. Figure 3 shows the variation of W_1 with depth y_3 along the vertical through a point y_2 = 5 km for a vertical fault with θ=90(in deg). It shows that W_1 decreases sharply upto a depth of about 15 km and thereafter diminishes to zero at a slower rate, and becomes significantly small for y_3 > 100km.

(B) Variation with depth of the main driving stress $t_{2'3'}$ in the dip-slip direction due to the movement across F.

Figures (4) – (7) show the variation of $t_{2'3'}$ with depth y_3 for various θ and some specific values of y_2.



y_3 (in km.)

Figure 4. Variation of the stress component $t_{2'3'}$ with y_3 for y_2=1 km., t=1 year, θ=90 (in deg) due to the fault movement

In Figure 4, it is found that for a vertical dip-slip fault θ=90 (in degree) and at a point very near to the fault y_2=1 km, $t_{2'3'}$ undergoes a change (in one year) by an amount < 0.5 bar, due to the creeping movement across F. Initially there is a very small region of stress release just below the free surface (0<y_3<1 km). Thereafter, the stress-accumulation region start, a maximum accumulation occurs at a depth of about 2 km below the free surface. This additional stress falls off quickly and becomes negligibly small at a depth of 20-25 km.

In case of the same vertical dip slip fault but a little bit away from the fault, (y_2=5 km), the characteristic of the nature of stress accumulation and release is the similar but of magnitude of much lower order (0:04bar). Figure 6 shows the variation of main driving stress component $t_{2'3'}$ with the depth for y_2 = 5 km, t = 1 year due to the creep movement across the fault at a dip angle θ=60(degree). The graph shows that there is a negligibly small region (<0.1 km) of stress release and then it accumulates upto a depth ≥ 9 km and attains its maximum value 22.5 bar (approx.) for t=1 year and then decreases to 0 at a depth about 100 km.

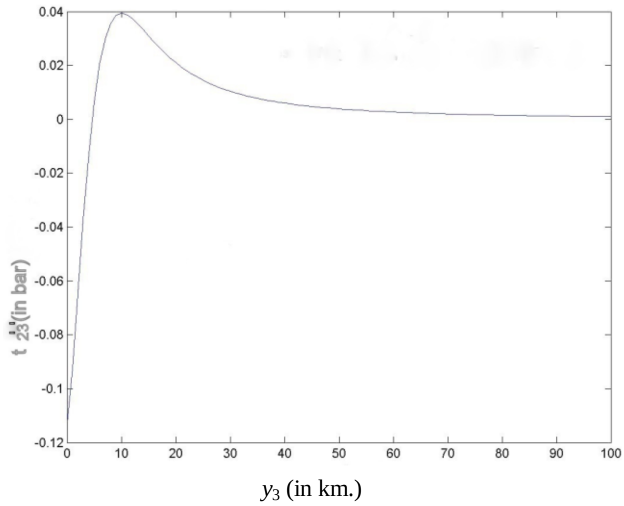


Figure 5. Variation of the stress component $t_{2'3'}$ with y_3 for y_2=5 km., t=1 year, θ=90 (in deg) due to the fault movement

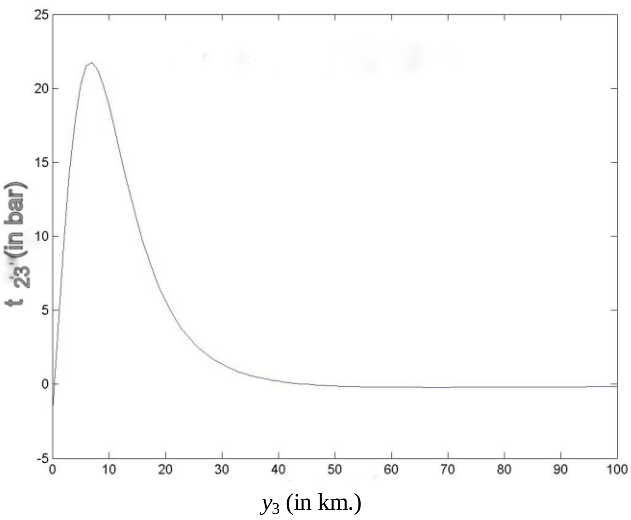


Figure 6. Variation of the stress component $t_{2'3'}$ with y_3 for y_2=5 km., t=1 year, θ=60 (in deg) due to the fault movement

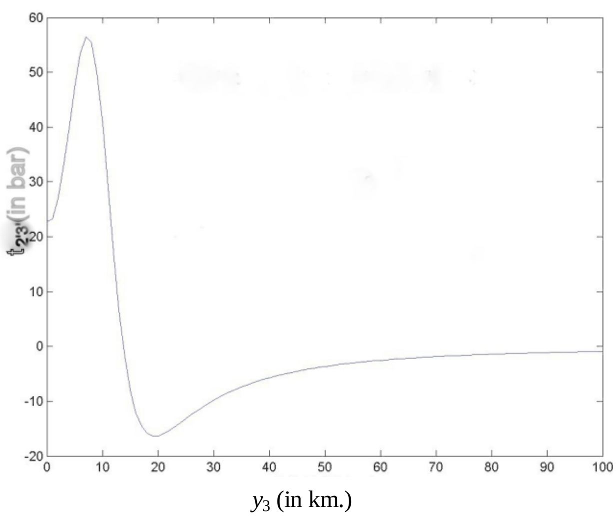


Figure 7. Variation of the stress component $t_{2'3'}$ with y_3 for y_2=9 km., t=1 year, θ=30 (in deg) due to the fault movement

Figure 7 shows the variation of stress component $t_{2'3'}$ for y_2 = 9 km, t = 1 year and dip angle θ=30(degree) with the depth due to the creep across the fault.

We see stress accumulates in the region $0 \le y_3 \le 17$ km then it releases upto 100 km. and become 0 at a depth about 110 km.

Thus in the above discussion we see that due to creep movement in the dip-slip fault, there are regions where stress get released and there are certain other regions where stress accumulates. The rate of stress release/accumulation depends essentially on the dip-angle θ and the distance y_3 from the fault.

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Appendix-1

Solutions for displacements, stresses and strains in the absence of any fault movement:

We take Laplace transform of all the constitutive equations and the boundary conditions (1.1)-(1.11) with respect to time and we get,

$$\overline{\tau_{22}} = \frac{\left(p\frac{\partial \bar{v}}{\partial y_2}\right) - \left(\frac{\partial v}{\partial y_2}\right)_0}{\frac{1}{\eta}+\frac{p}{\mu}} + \frac{\frac{1}{\mu}(\tau_{22})_0}{\frac{1}{\eta}+\frac{p}{\mu}} \qquad (3.1)$$

where, $\overline{\tau_{22}} = \int_0^\infty \tau_{22} e^{-pt} dt(p>0$, Laplace trans-formation variable) and similar other equations. Also the stress equations of motions in Laplace transform domain as:

$$\frac{\partial}{\partial y_1}(\overline{\tau_{11}})+\frac{\partial}{\partial y_2}(\overline{\tau_{12}})+\frac{\partial}{\partial y_3}(\overline{\tau_{13}})=0 \qquad (1.4a)$$

$$\frac{\partial}{\partial y_1}(\overline{\tau_{21}})+\frac{\partial}{\partial y_2}(\overline{\tau_{22}})+\frac{\partial}{\partial y_3}(\overline{\tau_{23}})=0 \qquad (1.5a)$$

$$\frac{\partial}{\partial y_1}(\overline{\tau_{31}})+\frac{\partial}{\partial y_2}(\overline{\tau_{32}})+\frac{\partial}{\partial y_3}(\overline{\tau_{33}})=0 \qquad (1.6a)$$

$\overline{\tau_{22}}(y_1,y_2,y_3,p)=\tau_\infty(p)cos\theta$ as $|y_3|\to\infty$, $y_3 = 0$ (1.7a)

On the surface y_3=0, $(-\infty<y_2<\infty)$

$$\overline{\tau_{23}}(y_1,y_2,y_3,p)=0 \qquad (1.8a)$$

$$\overline{\tau_{33}}(y_1,y_2,y_3,p)=0 \qquad (1.9a)$$

Also $y_3 \to \infty(0\infty < y_2 < \infty)$

$$\overline{\tau_{23}}(y_1,y_2,y_3,p)=0 \qquad (1.10a)$$

$$\overline{\tau_{33}}(y_1,y_2,y_3,p)=\tau_\infty(p)sin\theta \qquad (1.11a)$$

Using (3.1), other similar equations and assuming the initial fields to be zero, we get from (1.4a),and(1.6a)

$$\nabla^2(\bar{v})=0,\nabla^2(\bar{w})=0 \quad (3.2)$$

Thus we are to solve the boundary value problem (3.2) with the boundary conditions (1.7a)-(1.11a).

Let,

$$v(y_2,y_3,\theta,p)=\frac{(v)_0}{p}+A_1y_2+B_1y_3 \quad (3.3)$$

be the solution of (3.2)

where A_1 and B_1 are arbitrary constant which are independent of y_2 and y_3, to be determined using the initial and boundary conditions as above.

Using the boundary conditions (1.10a) – (1.11a) and the initial conditions we get,

$$A_1=\frac{1}{p}\left[\left(\frac{1}{\eta}+\frac{p}{\mu}\right)\right]\left(\tau_\infty(p)-\frac{1}{P+\frac{\mu}{\eta}}\tau_\infty(0)\right)sin\theta \quad (3.4)$$

$$\text{and, } B_1 = 0 \quad (3.5)$$

On taking inverse Laplace transformation we get,

$$v(y_2,y_3,\theta,t)=(v)_0+(y_2 cos\theta/\mu)$$
$$[\tau_\infty(0))+(\mu/\eta)\int_0^t \tau_\infty(\tau)d\tau] \quad (3.6)$$

Similarly we can get the other components of the displacements.

$$w(y_2,y_3,\theta,t)=(w)_0+(y_2 sin\theta/\mu)$$
$$[\tau_\infty(t)-\tau_\infty(0))+(\mu/\eta)\int_0^t \tau_\infty(\tau)d\tau] \quad (3.7)$$

The stresses are given by,

$$\tau_{32}=(\tau_\infty(t)cos\theta-[\tau_\infty(0)cos\theta-(\tau_{22})_0]e^{-(\mu/\eta)t} \quad (3.8)$$

$$\tau_{23}=(\tau_{23})_0e^{-(\mu/\eta)t} \quad (3.9)$$

$$\tau_{33}=(\tau_\infty(t)-\tau_\infty(0)e^{-(\mu/\eta)t})sin\theta \quad (3.10)$$

Using the displacements the strains can also be found out to be,

$$e_{22}(y_2,y_3,\theta,t)=(e_{22})_0 \quad (3.11)$$

$$e_{33}(y_2,y_3,\theta,t)=\left(\frac{1}{2}\right)(e_{23})_0 \quad (3.12)$$

Appendix-2

Solutions after the fault movement

We assume that after a time T_1 the stress component $\tau_{2'3'}$ (which is the main driving force for the dip-slip motion of the fault) exceeds the critical value τ_c and the fault F starts creeping. Then we have an additional condition characterizing the dislocation in w due to the creeping movement as:

$$[(w)]_r=w_1(t_1)g(y_1)H(t_1) \quad (4.1)$$

where, $H(t_1)$ is the Heaviside function and$[(w)]_F$. The discontinuity of w across F given by

$$[(w)]_F=\lim_{(y_2'\to 0+)}(w)-\lim_{(y_2'\to 0-)}(w) \quad (4.2)$$
$$(y_2'-0,0\le y_3'\le D)$$

Taking Laplace transformation in (4.1) we get,

$$[(\bar{w})]_F=w_1(p)q(y_3') \quad (4.3)$$

The fault creep commences across F after time T_1, clearly,

$$[(w)]_F = 0$$

for $t_1 \le 0$, where $t_1 = t-T_1$, F is located in the region ($y_3' = 0$, $0 < y_3' < D$).

We try to find the solution as,

$$v=(v)_1+(v)_2, w=(w)_1+(w)_2,$$
$$\tau_{22}=(\tau_{22})_1+(\tau_{22})_2, \tau_{23}=(\tau_{23})_1+(\tau_{23})_2, \quad (4.4)$$
$$\tau_{33}=(\tau_{33})_1+(\tau_{33})_2$$

Where $(v)_1$; $(w)_1$; $(\tau_{ij})_1$, are continuous everywhere in the model and are given by (A). While the second part $(v)_2$; $(\tau_{ij})_2$ are obtained by solving modified boundary value problem as stated below. We note that $(v)_2$; is continuous even after the fault creep, so that$[(v)]_2=0$,while $(w)_2$ satisfies the dislocation condition given by (4.2).

The resulting boundary value problem can now be stated as:$(w)_2$ satisfies 2D Laplace equation as,

$$\nabla^2(\bar{w})_2=0 \quad (4.5)$$

Where, $(\bar{w})_2$ is the Laplace transformation of $(w)_2$, with the modified boundary condition,

$$\bar{\tau}_{22}(y_2,y_3,\theta,p)=0 \text{ as } |y_2|\to\infty, y_3\ge 0 \quad (1.7b)$$

$$\bar{\tau}_{23}(y_2,y_3,\theta,p)=0 \text{ as}$$
$$y_3\to\infty(-\infty<y_2<\infty,) \quad (1.10b)$$

$$\bar{\tau}_{33}(y_2,y_3,\theta,p)=0 \text{ as}$$
$$y_3\to\infty(-\infty<y_2<\infty,) \quad (1.11b)$$

and the other boundary conditions are same as (1.8a)-(1.9a).

We solve the above boundary value problem by modified Green's function method following[16],[31], and the correspondence principle.

Let $Q(y_2,y_3)$ be any point in the field and $P(x_2,x_3)$ be any point in the fault, then we have,

$$(\bar{w})_2(Q)=\int_F (w_1)(P)g(x_3')[G_{33}^3(P,Q)dx_2- \quad (4.6)$$
$$-G_{32}^3(P,Q)dx_3]$$

where,

$$G_{33}^3(P,Q)=\left(\frac{1}{2\times\pi}\right)\frac{y_3-x_3}{[(y_2-x_2)^2+(y_3-x_3)^2]}- \quad (4.7)$$
$$-\frac{y_3+x_3}{[(y_2-x_2)^2+(y_3+x_3)^2]}$$

$$G_{32}^3(P,Q)=\left(\frac{1}{2\times\pi}\right)\frac{y_2-x_2}{[(y_2-x_2)^2+(y_3-x_3)^2]}- \quad (4.8)$$
$$-\frac{y_2+x_2}{[(y_2-x_2)^2+(y_3+x_3)^2]}$$

$$(\bar{w})_2(Q)=\int_F (w)_1(P)g(x_3')\times$$

$$\frac{(y_2 \sin\theta - y_3 \cos\theta)}{[(x_3')^2 - 2\times x_3'(y_2 \cos\theta - y_3 \sin\theta) + (y_2)^2 + (y_3)^2]} \times dx_3'$$

$$((w)_1(P)/2\times\pi)\phi_1(y_2, y_3, \theta) \text{ (say)} \quad (4.9)$$

Where,

$$\phi_1(y_2, y_3, \theta) = \int_F g(x_3') \frac{(y_2 \sin\theta - y_3 \cos\theta)}{[(x_3')^2 - 2\times x_3'(y_2 \cos\theta - y_3 \sin\theta) + (y_2)^2 + (y_3)^2]} - \frac{(y_2 \sin\theta + y_3 \cos\theta)}{[(x_3')^2 - 2\times x_3'(y_2 \cos\theta - y_3 \sin\theta) + (y_3)^2 + (y_3)^2]}] \times dx_3' \quad (4.10)$$

Taking inverse Laplace transformation,

$$(w)_2(Q) = w_1(t_1)\phi_1(y_2, y_3, \theta)H(t_1)$$

Where, $H(t_1)$ is the Heaviside step function, which gives the displacement at any points $Q(y_2, y_3)$.

We also have,

$$(\bar{\tau}_{22})_2 = 0$$

$$(\bar{\tau}_{23})_2 = \frac{w_1(p)}{(2\times\pi)} \frac{p}{\frac{p}{\mu} + \frac{1}{\eta}} \phi_2(y_2, y_3, \theta)$$

where, $\phi_2(y_2, y_3, \theta) = \frac{\partial}{\partial y_2}\phi_1(y_2, y_3, \theta)$

$$(\bar{\tau}_{33})_2 = \frac{w_1(p)}{(2\times\pi)} \frac{p}{\frac{p}{\mu} + \frac{1}{\eta}} \phi_3(y_2, y_3, \theta)$$

where, $\phi_3(y_2, y_3, \theta) = \frac{\partial}{\partial y_3}\phi_1(y_2, y_3, \theta)$

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