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Michele Dragoni, Maurizio Bonafede, Enzo Boschi

Earthquake recurrence time on a long strike slip fault subject to uniform strain rate

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Sismologia. — *Earthquake recurrence time on a long strike slip fault subject to uniform strain rate.* Nota (*) di MICHELE DRAGONI, MAURIZIO BONAFEDE e ENZO BOSCHI, presentata dal Corrisp. E. BOSCHI.

RIASSUNTO. — Si propone un modello in cui la litosfera nei pressi di una lunga faglia verticale a scorrimento orizzontale è considerata come un solido viscoelastico di Maxwell. Imponendo condizioni al contorno di velocità di deformazione uniforme, si ottiene una relazione per il periodo di ritorno dei grandi terremoti. Questo modello porta a previsioni diverse da quelle di modelli simili con condizioni al contorno di sforzo applicato uniforme.

1. INTRODUCTION

It is well known that fault regions are intensively fractured. Faulting appears to occur at all scales: faults vary in length from thousands of kilometres to cristalline dimensions and intersect one another in a complex pattern (King, 1978). This suggests that modelling the fault zone as a homogeneous elastic medium with a single fault surface in it may be inadequate, at least for long-term deformation. In particular, the presence of a large number of secondary faults and smaller fractures, on which creep occurs in response to applied stresses, makes the crust effectively more compliant. The change in compliance is gradual and the overall effect may not be distinguishable from viscoelastic relaxation (Mavko, 1981). It seems therefore meaningful to introduce an effective viscosity in describing the rheology of highly fractured zones. Budianski and Amazigo (1976) examined the interaction between slip on a long strike-slip fault and lithospheric creep, by modelling the lithosphere as a Maxwell viscoelastic plate. By imposing on the plate a uniform shear stress applied at a very large distance, laterally with respect to the fault, they obtained a solution for the periodic occurrence of earthquakes on the fault. A relation for the recurrence time of earthquakes was given. On the basis of this model, they inferred an effective viscosity of the order of 10^{21} P for the San Andreas fault region. Burridge (1977) generalized Budianski and Amazigo's (1976) model by representing the lithosphere as a half-space with the rheology of a general linear viscoelastic solid and introducing a more realistic fracture criterion on the fault. When specialized to a Maxwell solid, his result for the earthquake recurrence time

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does not differ appreciably, however, from the relation obtained by Budianski and Amazigo (1976).

Both these models impose boundary conditions on stress. However, the state of stress in the lithosphere is poorly known (see e.g. Hanks, 1977). In connection with the earthquake occurrence, we often have information on the stress drop, but not on the absolute values of stress before and after the seismic event. Moreover, the stress field is very sensitive to local inhomogeneities. On the contrary, we have a fairly detailed knowledge of the strain rate field at the Earth's surface. Both geodetic and paleomagnetic data provide a measure of this field. The long-term displacement at depth must be the same as at the surface for plate motions to be accommodated at depth; this displacement takes place mostly aseismically at depth (Davies and Brune, 1971). In modelling the ground deformation along a plate margin, it might then be more appropriate to impose boundary conditions on strain rate, rather than on stress. This may drastically alter the estimate of the recurrence time of great earthquakes.

2. THE MODEL

We model the fault region as a Maxwell viscoelastic half-space characterized by a shear modulus μ and a viscosity η. A long, vertical strike-slip fault lies on the plane $x_2 = 0$ and extends from the Earth's surface ($x_3 = 0$) to depth $x_3 = d$. The model is two-dimensional, all quantities being uniform in the x_1 direction (fig. 1). Moreover, only the u_1 component of the displacement field is non-vanishing. Accordingly

$$\varepsilon_{21} = \frac{1}{2}\, \partial u_1 / \partial x_2 \tag{1}$$

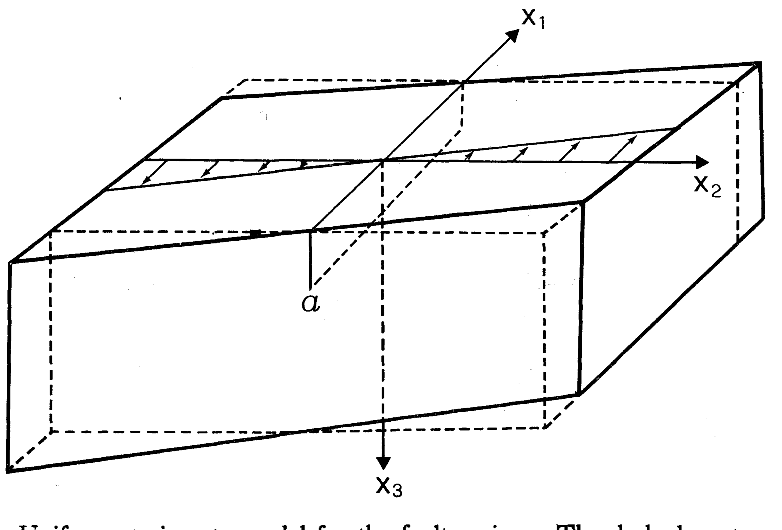

Fig. 1. – Uniform strain rate model for the fault region. The dashed contours denote the unstrained state, the arrows denote the displacement field.

is the relevant shear strain component in our model. The corresponding stress component is

$$\sigma_{21} = 2\,\mu\varepsilon_{21}\,. \tag{2}$$

The constitutive equation for a Maxwell solid is

$$\dot{\sigma}_{21}/\mu + \sigma_{21}/\eta = 2\,\dot{\varepsilon}_{21} \tag{3}$$

where dots denote differentiation with respect to time. Following Budianski and Amazigo (1976), we suppose that slip on the fault occurs when the applied shear stress reaches a statical friction resistance σ_S and that, due to this slip, stress on the fault drops to a dynamical friction resistance σ_D. Free-surface boundary conditions are assumed at the Earth's surface ($x_3 = 0$):

$$\sigma_{31}(x_2\,,0) = 0\,. \tag{4}$$

We assume that the fault region is deformed by a uniform, constant strain rate field:

$$\dot{\varepsilon}_{21}(x_2\,,x_3) = K \tag{5}$$

as sketched in fig. 1. K is half the velocity gradient in the x_2 direction. If we assume that the last earthquake occurred on the fault at $t = 0$, the displacement field at a later time t is given by the displacement of a viscous medium plus an arbitrary initial displacement:

$$u_1(x_2\,,x_3\;;t) = 2\,Kx_2\,t + u_1(x_2\,,x_3\;;0+)\,. \tag{6}$$

From (1) and (2) it follows that

$$\dot{\sigma}_{21}/\mu + \sigma_{21}/\eta = 2\,K\,. \tag{7}$$

The solution is

$$\sigma_{21}(x_2\,,x_3;t) = 2\,K\eta\,(1 - e^{-t/\tau}) + \sigma_{21}(x_2\,,x_3\;;0+)\,e^{-t/\tau} \tag{8}$$

where

$$\tau = \eta/\mu \tag{9}$$

is the Maxwell relaxation time. On the fault surface, $0 \leq x_3 \leq d$,

$$\sigma_{21}(0\,,x_3;t) = 2\,K\eta\,(1 - e^{-t/\tau}) + \sigma_D\,e^{-t/\tau}\,. \tag{10}$$

This stress is plotted in fig. 2 for various times. The shear stress $\sigma_{21}(0, x_3; t)$ on the fault surface remains uniform so that the next earthquake will occur at time $t = \mathrm{T}$, when σ_{21} reaches again the values σ_S:

$$\sigma_{21}(0, x_3; \mathrm{T}) = \sigma_S . \tag{11}$$

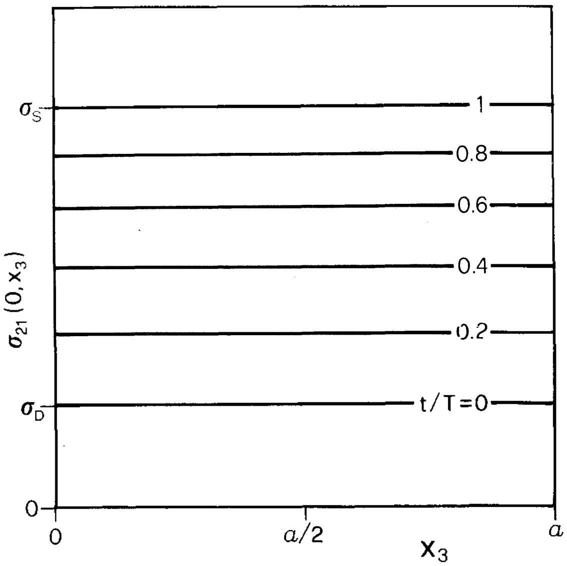

Fig. 2. – Shear stress on the fault surface as a function of depth x_3, for various values of time in units of the recurrence period T. The graph is drawn for $\sigma_D = \sigma_S/4$ and $\mathrm{T} = \tau/2$.

From (10) and (11) we obtain the relation

$$\mathrm{T} = \tau \log\left[(2\,\mathrm{K}\eta - \sigma_D)/(2\,\mathrm{K}\eta - \sigma_S)\right] \tag{12}$$

for the recurrence time of great earthquakes on the fault. The recurrence time T as a function of the effective viscosity η is plotted in fig. 3. At time $t = \mathrm{T}$, we have a uniform stress drop $\Delta\sigma$ on the fault,

$$\Delta\sigma = \sigma_S - \sigma_D \tag{13}$$

and a displacement discontinuity

$$\Delta u_1(x_3) = 2\,\Delta\sigma\,(d^2 - x_3^2)^{1/2}/\mu . \tag{14}$$

If the slipped fault has horizontal length L ($\mathrm{L} \gg d$), this gives a seismic moment

$$\mathrm{M} = \pi\,\mathrm{L}d^2\,\Delta\sigma/2 . \tag{15}$$

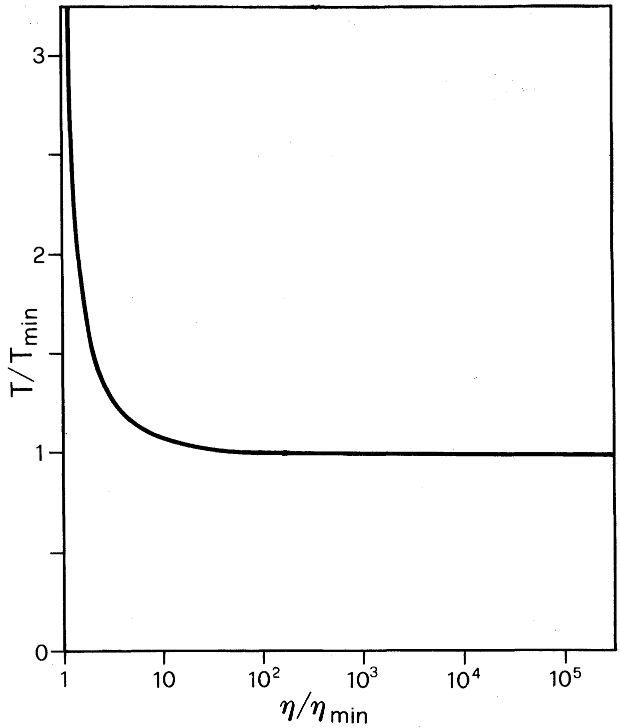

Fig. 3_1. – The recurrence time T of great earthquakes as a function of effective viscosity η, according to our model. The graph is drawn for $\sigma_D = \sigma_S/4$.

As deduced from Eq. (12) and fig. 3, the earthquake recurrence time T has a lower limit

$$T_{min} = \Delta\sigma/(2\,K\mu) \tag{16}$$

which is proportional to the stress drop $\Delta\sigma$. There is also a lower limit for the effective viscosity:

$$\eta_{min} = \sigma_S/(2\,K)\,. \tag{17}$$

For most viscosity values, the recurrence time is very close to its minimum value T_{min}. It is substantially higher only when the effective viscosity is close to its minimum value η_{min}. As can be seen from fig. 3, T tends to infinity as η approaches η_{min}. The physical meaning of the existence of a lower limit for η is that for too low a viscosity, stress relaxation is too fast and the shear stress on the fault surface can never reach the critical value σ_S required for slipping. This is evident from Eq. (10), where the maximum shear stress which can be attained on the fault surface is $2\,K\eta$. It is therefore clear why η_{min} depends on σ_S. We must have $2\,K\eta > \sigma_S$ in order that the earthquake ' engine ' may work.

Had we assumed a boundary condition of uniform stress σ_A applied at infinity,

$$\sigma_{21}(x_2, x_3) = \sigma_A \tag{18}$$

the displacement field would be

$$u_1(x_2, x_3; t) = \sigma_A x_2 t/\eta + u_1(x_2, x_3; 0+) . \tag{19}$$

By comparison with Eq. (6), the solution to this problem is simply obtained by substituting σ_A/η to 2 K in all equations. In particular, the earthquake recurrence time would be

$$\mathrm{T} = \tau \log [(\sigma_A - \sigma_D)/(\sigma_A - \sigma_S)] \tag{20}$$

which is the relation obtained by Budianski and Amazigo (1976). In this case T depends linearly on η.

3. Discussion and conclusions

We have obtained a relation for the recurrence time T of great earthquakes on the fault, by imposing boundary conditions of assigned velocity or velocity gradient. This relation, Eq. (12), is different from the one obtained by Budianski and Amazigo (1976), Eq. (20), who imposed stress boundary conditions, so that a markedly different behaviour of T as a function of η is obtained (fig. 3). We now apply this model to the Northern 'locked' segment of the San Andreas fault and try a numerical estimate of some of the quantities involved in the model. We use throughout $d = 10$ km for the maximum fault depth (see e.g. Barker, 1976) and $\mu = 3 \times 10^{11}$ dyne/cm² for the shear modulus.

Savage and Burford (1973) inferred a relative plate velocity across the San Andreas fault of about 3 cm/yr from geodetic measurements carried out at a maximum distance $|x_2| \simeq 50$ km from the fault trace. This gives a strain rate $\mathrm{K} \simeq 3 \times 10^{-7}\,\mathrm{yr}^{-1}$ (see also Thatcher, 1975b). From Eq. (14), a stress drop $\Delta\sigma \simeq 60$ bar corresponds to a surface slip $\Delta u(0) \simeq 4$ m, which is the average value observed after the 1906 San Francisco earthquake (Thatcher, 1975a). From Eq. (15), with $\mathrm{L} = 500$ km, a seismic moment $\mathrm{M} = 4 \times 10^{27}$ dyne cm is obtained, which agrees with the one determined from long-period surface wave amplitudes (Thatcher, 1975a). With the same stress drop, we obtain from Eq. (16) a minimum recurrence time $\mathrm{T}_{\min} \simeq 330$ yr. Smaller earthquakes would have comparatively shorter recurrence times. We note that, if we take $\sigma_S \simeq 100$ bar as the order of magnitude of the critical stress, we obtain $\eta_{\min} \simeq 5 \times 10^{21}$ P from Eq. (17), which is of the same order of magnitude as the effective viscosity obtained by Budianski and Amazigo (1976). In our case this is only the minimum value and all higher values are admissible.

The estimate obtained for the minimum earthquake recurrence time is considerably larger than the value $T \simeq 100$ yr which is commonly quoted for the San Andreas fault. In the framework of our model, with assigned strain-rate boundary conditions, if the strain accumulation process is uniform in time, the only possibility that great earthquakes have shorter recurrence times is to have a somewhat larger strain rate K. It seems, however, that stress accumulation is non-uniform in time. During the 50 years preceding the 1906 San Francisco earthquake, the strain accumulation rate measured at the Earth's surface was considerably higher than nowadays, ranging from 4 to 9×10^{-7} yr^{-1} (Thatcher, 1975a). As to the origin of the frequently quoted 100-year recurrence period for great earthquakes on the San Andreas fault, it seems to come from Reid's (1910) inferences based on measurements taken before the 1906 San Francisco earthquake. As pointed out by Thatcher (1975a), this is correct only if strain accumulated uniformly in time: in fact, from Eq. (16), if $K = 9 \times 10^{-7}$ yr^{-1} and $\Delta\sigma = 60$ bar, we get $T_{min} \simeq 110$ yr. Since shear straining was considerably greater before the 1906 earthquake than after it, any recurrence time estimated on the basis of pre-earthquake data is likely to be inaccurate. If the present value of the strain rate is representative of the sverage inter-seismic value. the acceleration observed during the 50 years preceding the 1906 San Francisco earthquake might be due, among other possibilities, to an increasing rate of aseismic slippage at depth, or to a gradual extenaion of the aseismic slippage zone to shallower depths (Bonafede and Dragoni. 1982). A possibility is that a similar evolution of the fault region may repeat in the future: in this case, the next large earthquake on the Northern 'locked' segment of the San Andreas fault might be announced by an increase in strain rate. The pre-seismic acceleration in straining would have the effect of decreasing the earthquake recurrence time with respect to the values found in the present work, still keeping it above the 100-years estimate.

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