RUPTURE MECHANICS OF PLATE BOUNDARIES

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Abstract. A mechanical instability--drastically different from stick-slip type models may be responsible for every large earthquakes, M $\geq$ 7 or so. In the model presented here, the beginning of the rupture cycle consists of aseismic yielding at the bottom part of the lithosphere. The yield zone propagates upwards, initially very slowly, with accelerating speed. It breaks out at sonic speeds, with surface rupture length which is determined by the thickness of the lithosphere. Predicted values for surface rupture range from 100 to 500 km or so for strike-slip events, in agreement with observed values.

Introduction

One of the most important questions in the design of long range and long duration experiments to deterime the deformation of the earth's surface near plate boundaries is the expected time-space pattern of this deformation. Positioning of ground stations, and their density, repeat rates of measurements, and expected spatial coherency can all be greatly improved with some preknowledge of the character of the field to be measured. In this paper we develop a conceptual model for inter-, pre-, and co-seismic surface deformation at and around active plate boundaries, in order to provide the range of surface deformation patterns which we can anticipate from available physical models of rock yielding.

It is generally recognized that large earthquakes (M $\geq$ 7.0 or so) are responsible for most of the seismic slip at plate boundaries. In spite of the dramatic model of instability of these slip events--in the form of destructive earthquakes with rupture lengths extending hundreds of kilometers, most of the deformation at plate boundaries is not seismic. In large thrust earthquakes, seismic rupture occurs to depth of only about 1/2 the plate thickness (e.g., the Alaska 1964 earthquake) leaving the lower part unruptured. In large strike-slip events this depth is typically even smaller--10 or 15 km as compared with 70 to 100 km for the thickness of the lithosphere. Consequently, a major portion of the plate boundary--below the seismogenic zone--must yield aseismically.

Savage (1975) proposed a simple model shown in Figure 1, invoking steady creep in a narrow zone, extending underneath the seismogenic trace of the plate boundary. Turcotte and Spence (1974) suggested a model in which the stress remains constant with time in the aseismic zone, with slip rate which varies with depth. Prescott and Nur (1980) suggested, on the basis of laboratory observed creep behavior of hot rock, and field observations of interseismic crustal deformation, that the aseismic zone is more likely to be of some width in which shear strain rate is anomalously high.

None of these current models treat the problems of seismic yielding as a truly mechanical one--in which motion is more or less rigorously related to the acting forces or the stresses. Such motion must include the time history not only of particles in the boundary region, but also the possible motion of edges of the zone in which plate boundary yielding does actually occur.

Furthermore, the correct mechanism of the process must also link the aseismic yielding to the fact that seismic gaps for large earthquake (M $\geq$ 7 or so) seem to be filled in a systematic and regular fashion--unlike smaller earthquakes (M $\leq$ 6 or so) which fail to show any significant regularity. The mechanical model must also be able to describe the development of yielding and rupture in time and space and provide understanding of the factors which control the length of rupture of very large earthquakes. Typical ruptures are hundreds of kilometers long, several times the thickness of the lithosphere in which they occur.

In this paper we have attempted to extend the Savage model for the rupture at plate boundaries, to include an aseismic as well as the seismic instability. In the next section we outline the basic concepts of the model and show preliminary results. The mechanics considered

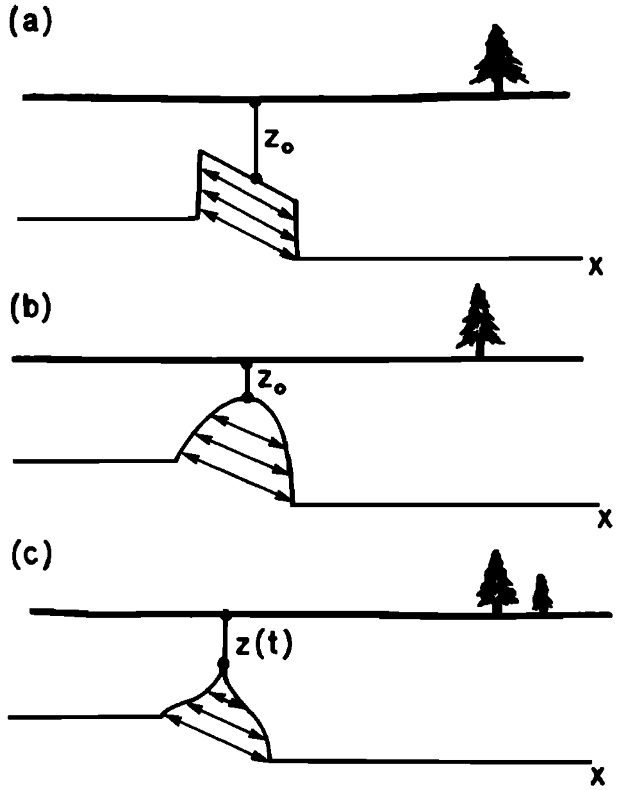

Fig. 1. Models for rupture of the lithosphere: (a) The Savage Model with fixed dislocation at depth Z_o; (b) The Turcotte model with fixed crack at Z_o; (c) Proposed model with moving dislocation/crack or dislocation flux. The depth to tip Z(t) decreases with time.

provide very specific predictions of the time history of surface deformations associated with plate boundary yielding to distances of hundreds of kilometers away from the boundary and along the boundary. The conceptual model as given here is exceedingly simple-minded, and should be developed in the future in much more detail, with the mechanics involved to be refined in order to provide a basis for the design of long and large scale (10-300 km) repeated geodetic deformation measurements which might be necessary to reveal the actual crustal process leading to, during, and following very large events. The development of this model by incorporating more detailed rheology may thus provide working hypotheses for intermediate range prediction of very large earthquakes.

An Instability Model for Large Earthquakes

We begin by considering the simple screw dislocation model of Savage and Burford (1970), which represents an infinitely long transcurrent fault in a flat earth. In the model a dislocation, with Burger's vector b_o, is situated at depth Z_o below the free surface as shown in Figure 1.

In the Savage model the stress at the neighborhood of the tip of the slip zone becomes very high--with the consequence that the tip must become unstable, with a tendency to move. Turcotte's model (1974) attempts to overcome part of this problem, by imposing a constant stress condition on the slip zone for $Z > Z_o$, as shown in Figure 1. However, the stress acting on the tip is again very large, suggesting again that this configuration is not stable.

Since models with fixed slip zones lead invariably to very high tip stresses, we may consider relaxing the assumption that the tip depth is fixed. Instead we assume that the tip can move in response to the forces acting on it. The simplest model is obtained from classical dislocation theory, where motion of a dislocation is related to the force acting on it through some simple law of motion (Head, 1972). For the configuration on Figure 1, the force acting on the dislocation is simply the attraction to the free surface (e.g., Cottrell, 1964)

$$F = \frac{\mu b^2}{4\pi} \cdot \frac{1}{Z} \qquad (1)$$

where F is force per unit length of dislocation line, b is the Burger's vector, and μ is the shear modulus.

A similar expression holds for the attractive glide force acting on an edge dislocation towards the free surface

$$F = \frac{\mu b^2}{4\pi(1 - \nu)} \cdot \frac{1}{Z}. \qquad (2)$$

A noteworthy feature of equations (1) and (2) is that the force acting on the dislocation is inversely proportional to its distance from the free surface to which it is attracted.

A simple law for the motion of a dislocation under force may be given by the power law (Head, 1972)

$$-\frac{dZ}{dt} = A \cdot F^n \qquad (3)$$

where Z is depth, and A is a constant (in the Savage model, A = 0). The power coefficient n must be selected on the basis of either experimental results or additional models. In a rough way, the value of n may be related to laboratory observed power law creep in rocks (e.g., Kirby, 1977). We may consider therefore the range $1 \leq n \leq 5$ for lithospheric behavior. Combining equation (1) and (3) for the screw dislocation, we obtain

$$- \frac{dZ}{dt} = C_o \left(\frac{1}{Z}\right)^n \tag{4}$$

with the constant $C_o = A\left(\frac{\mu b^2}{4\pi}\right)^n$.

Assuming a dislocation source at Z_o, the initial condition $Z = Z_o$ at $t = 0$ and the final state $Z = 0$ at $t = t_1$ we obtain the position of the dislocation as a function of time

$$\overline{Z} = (1 - T)^{1/(n+1)} \tag{5}$$

and the dislocation velocity

$$\overline{V} = - \frac{d\overline{Z}}{dt} = \left(\frac{1}{n + 1}\right)(1 - T)^{-n/(n+1)} \tag{6}$$

where $\overline{Z} = Z/Z_o$ and $T = t/t_1$.

Equation (6) shows that at early time $T << 1$, the dislocation moves slowly towards the surface. However, it greatly accelerates as T approaches 1, and becomes infinite as it breaks out. We suggest that this explosive breakout of the dislocation, for $n \geq 1$, may be useful in describing the instability process which is responsible for very large earthquakes. The effect is quite insensitive to details of stress,

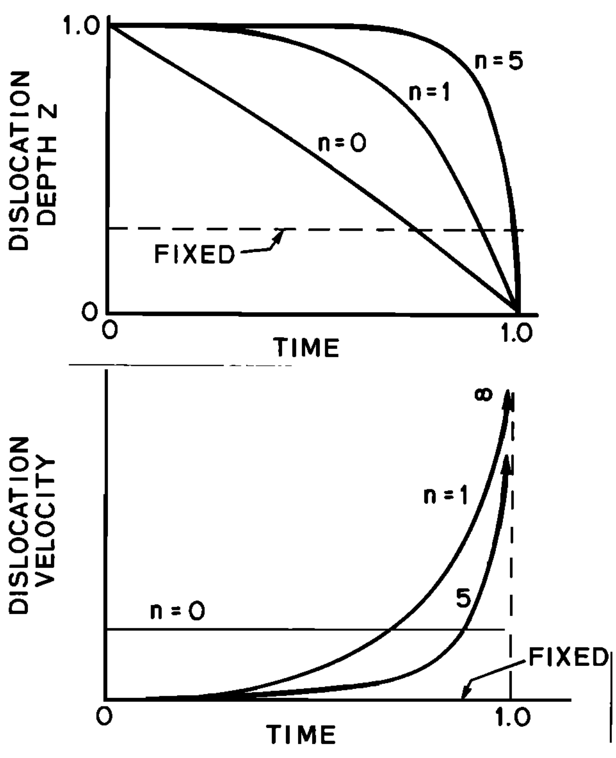

Fig. 2. The depth and velocity of the screw dislocation moving under the attractive surface force. The factor n indicates the motion power law used in the computation. The Savage and Turcotte models correspond to the fixed dislocation depth.

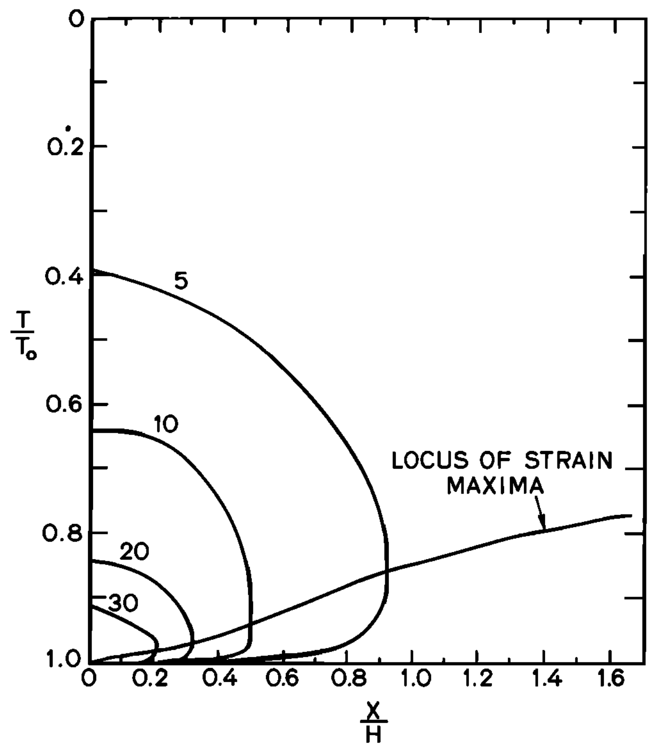

Fig. 3. The time-space distribution of surface strain for an upward moving dislocation with power law coefficient n = 1. Note that surface strain for most points away from the fault ($x \neq 0$), first increase gradually and then decrease before the breakout at $T/T_o = 1$, $X = 0$.

geometry and the exact value of n. It does depend however, on the assumption that the material involved fails by localization of strain, and that the localized zone obeys in some gross sense a power law behavior.

In Figure 2 we show the depth $\overline{Z}$ and the velocity $\overline{V}$ of the screw dislocation for various values for n--invariably showing the breakout instability.

Using the dislocation model we calculate the free surface displacement U_3 and strain fields e_{23} as a function of time.

$$U_3(X,t) = 2b\left[\tan^{-1}\left(\frac{X}{Z(t)}\right)\right]$$

and (7)

$$e_{23}(X,t) = 2b\left[\frac{Z(t)}{X^2 + Z^2(t)}\right]$$

where $X = x/Z_o$, is the distance perpendicular to the fault. Figure 3 shows an example of these fields for n = 1. The strain field is first very broad becoming more concentrated with time.

In general we find that most of the surface strain accumulation occurs only in the last 30% to 10% of earthquake repeat cycle.

The Length of Surface Ruptures

The two-dimensional model can be extended to three dimensions, using the line tension concept. We consider a situation as shown in Figure 4 in which two types of forces act on the bowed dislocation line shown: the attraction towards the free surface, given approximately by equation (3), and the self attraction of the line, given by

$$T = \alpha\mu b^2 \frac{1}{R} \tag{8}$$

where R is the radius of curvature, and α is a constant. Expressing R in terms of the depth Z of the closest point of the line to the surface, normalized with respect to the depth of the source Z_o, and the half length of the slip zone L, we obtain the line tension

$$T = \alpha\mu b^2 \frac{2(1 - Z)}{L^2 + (1 - Z)^2} \; . \tag{9}$$

The configuration of the dislocation line at any given moment is controlled by the equilibrium between the line tension T which tends to keep the line straight and thus opposes the motion towards the free surface, and the image attraction towards the free surface F. Equating the two forces we obtain the relation between the half length L of the slip zone at depth and the depth of the dislocation line Z as shown in Figure 4.

$$L = [(8\pi\alpha + 2)Z - (8\pi\alpha + 1)Z^2 - 1]^{\frac{1}{2}}. \tag{10}$$

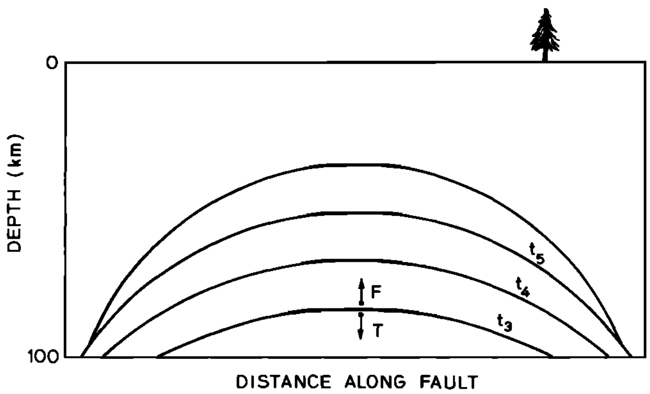

Fig. 4. Equilibrium configurations of bowed dislocation line, subject to line tension T and surface attraction f, at consecutive time t_i. Shortly after t_3 the system becomes unstable and rupture accelerates toward the free surface.

We notice at once that equation (10) yields a half length L which is double valued. This implies that for $Z < Z_c$, where Z_c is some critical depth, $0 < Z_c < Z_o$, the motion of the dislocation line becomes unstable, breaking out at speeds approaching wave speeds. Differentiating equation (10) with respect to Z, we find the critical depth Z_c at which the equilibrium rupture half length L_c is largest

$$Z_c = \frac{4\pi\alpha + 1}{8\pi\alpha + 1} \; . \tag{11}$$

Reasonable values are (Cotrell, 1964) $1/\pi \leq \alpha \leq 4/\pi$, leading to $.5 \leq Z_c \leq .6$. This suggests that the unstable motion of the dislocation line begins at a depth of about 1/2 to 2/3 the depth of the lithosphere. Creep or slip below this depth is slow, involving a gradual increase of the length of the zone. At the onset of the instability, the critical half length is

$$L_c = \frac{2\pi\alpha(8\pi\alpha - 2)}{(8\pi\alpha + 1)} \; . \tag{12}$$

For $\alpha = 1/\pi$, $2/\pi$, and $4/\pi$, the values of the critical length L_c are 1.15, 1.94 and 2.7 times the thickness of the lithosphere. Taking this thickness as 70 km, computed rupture lengths range between 170 and 500 km. These values are in good agreement with observed values for large earthquakes, such as 400 km for the 1906 San Francisco earthquake.

Assuming again that our lithospheric model behaves elastically outside the slip zone, we may use dislocation theory also to estimate the spatial and temporal pattern of surface deformation--displacement, tilt, and strain--preceeding the breakout of rupture, in the form of a major earthquake. Figure 5 shows the computed surface displacement history for the bowing dislocation line model.

Discussion and Conclusion

I have shown in this paper that a mechanical instability--drastically different from stick-slip type models (e.g., Brace and Byerlee, 1970) may be responsible for very large earthquakes $M \geq 7$ or so. In the model, the beginning of the rupture cycle consists of aseismic yielding at the bottom part of the lithosphere. The yield zone propagates upwards, initially very slowly, with accelerating speed. It breaks out at sonic speeds, with surface rupture length which is determined by the thickness of the lithosphere. Predicted values for surface rupture range from 100 to 500 km or so for strike-slip events, in agreement with observed values.

The instability as outlined in the previous sections is at best only a conceptual outline of the process. A comprehensive development of the model must include much more realistic

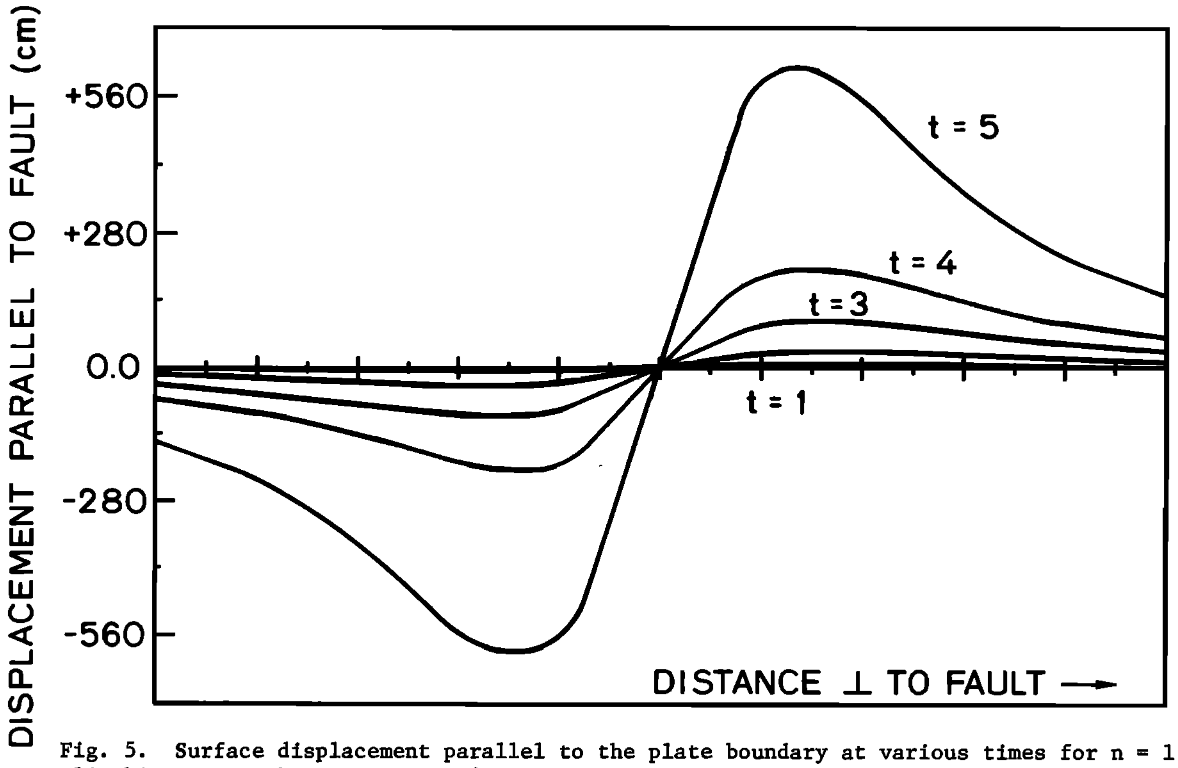

Fig. 5. Surface displacement parallel to the plate boundary at various times for n = 1, and slip history as shown in Figure 4.

and more detailed characteristics of the lithosphere, and comparison with observed inter-, pre-, and co-seismic deformation data such as from California and Japan, including for example, rupture length vs. magnitude. It will also be necessary to analyze the instability in a plate with a lower free boundary, in contrast to the present half space analysis (e.g., Chou, 1965). The net effect of the lower plate boundary is likely to retard the motion of the dislocation at early times in the interseismic cycle, and accelerate the motion in its later part, relative to the half space case.

The entire analysis must also be developed for thrust faulting--using edge dislocations, moving under glide forces, with dependence on the power n in dislocation velocity power law, the tip of the slip plane, and the possible effects of climb forces.

The uniform slip dislocation model can further be generalized to a variable slip model by superposition of dislocations. In particular we need to study the effects of a stiff brittle crust over a more compliant lithosphere. It is well known that additional forces on dislocations are exerted by elastic interfaces. The effect of a stiff crust might be to retard the dislocation motion at depth, but accelerate it once it penetrates the crust, thus enhancing the instability.

The model may also be extended to include post-seismic deformation in which the dynamic stress drop forces the dislocation to move back down away from the free surface. Because this motion just below the seismic zone is downward, it is stable, with a velocity which decreases with depth or time.

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