A dislocation model of microplate boundary ruptures in the presence of a viscoelastic asthenosphere

M. Bonafede, E. Boschi* and M. Dragoni

Dipartimento di Fisica, Settore di Geofisica, Università degli Studi di Bologna,
and Istituto Nazionale di Fisica Nucleare, Sezione di Bologna, Viale Berti Pichat 8,
40127 Bologna, Italy

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Summary. A microplate is modelled as an elastic plate with two long strike-slip boundaries, lying over a Maxwell-type viscoelastic asthenosphere. The microplate is subjected to a constant and uniform shear strain rate by the opposite motions of two adjoining larger plates. After the occurrence of an earthquake at one of the microplate boundaries, the time evolution of shear stress at the other boundary is studied. It is found that stress build-up at the second boundary is delayed due to stress diffusion governed by the asthenosphere relaxation. Earthquake occurrence at this latter boundary would be delayed depending upon both the microplate width and the ratio between the Maxwell relaxation time of the asthenosphere and a characteristic time required for tectonic strain to recover rupture conditions. It turns out that the parameters which determine the occurrence of seismic activity along the microplate boundaries are more strictly constrained in the presence of a viscoelastic asthenosphere than in the case of an elastic half-space model.

1 Introduction

Plate margins are often well-defined belts, which are characterized by intensive tectonic activity. The relative motions of the adjoining plates can be easily determined in these cases (Minster & Jordan 1978). In some regions, however, no sharp boundary has been found. The common belief is that, due to a complex geological history, the lithosphere along the boundary is fragmented into several smaller plates, which are usually termed 'microplates'. Microplates have horizontal dimensions which are usually less than $10^3$ km. Their presence makes it particularly difficult to understand the geodynamical processes occurring in the region. In particular, the boundaries between microplates can slip past one another under the action of tectonic forces, so that the pattern of seismic activity may be extremely complex.

*Also at Istituto Nazionale di Geofisica, Rome, Italy,
An example of fragmentation can be found in California, where the plate boundary appears to be composed of several parallel strike-slip strands. Extending north from the San Francisco Bay, the Humboldt microplate is bounded by the San Andreas fault on the west, by the Calaveras–Hayward-Rodgers Creek–Maacman–Lake Mountain faults on the east and by the Gorda plate on the north (Herd 1978).

An extensive fragmentation occurs in the Mediterranean region, which marks the boundary between the Eurasian and the African plates (McKenzie 1970, 1972). Here there is no single seismic lineament along which earthquakes are seen to occur more or less regularly and notions such as the ‘earthquake recurrence time’, which are fruitfully applied to other parts of the world, lose much of their meaning. The distribution of earthquake epicentres is in fact spread over large areas. However, they are often concentrated along closed contours, which define relatively aseismic areas and are interpreted as microplate boundaries (see, e.g. Papazachos 1973; Udías & López Arroyo 1972). For instance, the Apennines on the west and the Dinarides on the east seem to mark the boundaries of the almost rectangular Adriatic microplate, where large earthquakes take place respectively through tensile and compressive mechanisms (see, e.g. Mantovani & Boschi 1983). In many cases, the microplate boundaries are only guessed, but not conclusively recognized.
Furthermore, no regular migration pattern can be identified, except in a very few cases (Richter 1958; Mogi 1968). Even the definition of what a microplate is has never been clearly stated in the literature. A major plate can be defined, according to plate tectonics, in terms of the shear stress, exerted by a viscoelastic asthenosphere upon the base of the plate, being capable of building up fracture stresses along the plate boundaries (Hanks 1977; Bonafede & Dragoni 1982); it thus seems natural to define a microplate as a lithosphere fragment which, because of its 'small' basal area, would not be capable by itself of building up fracture stresses along its margins, so that its dynamics are mostly governed by the motions of the adjoining major plates. These would be the Pacific and the North American plates in the case of the Humboldt microplate and the African and the Eurasian plates in the case of the Mediterranean region. From this point of view, the microplate margins are weakness zones defined by a closed pattern of seismicity which encloses a relatively aseismic region with 'small' horizontal dimensions in the sense stated above.

A study of stress redistribution after a large earthquake is therefore necessary to ascertain which other microplate boundary section is a probable candidate for the next large earthquake to occur. In this connection, a major role is played by the asthenosphere, due to its viscoelastic properties. In fact, the asthenosphere gradually relaxes the stress as imposed on it by earthquakes occurring in the overlying lithosphere and so it alters the state of stress in the lithosphere, with a time-scale depending upon its rigidity and viscosity. The role of the asthenosphere in controlling the stress evolution along major plate boundaries has been emphasized by many authors (see, e.g. Nur & Mavko 1974; Melosh 1976; Savage & Prescott 1978a, b; Spence & Turcotte 1979; Dragoni, Bonafede & Boschi 1982). Obviously, the asthenosphere may significantly influence the earthquake occurrence times also on microplate boundaries.

In the present paper, we consider the two-dimensional model of a microplate with two long, parallel transcurrent margins, embedded between two larger plates (Fig. 1), overlying a viscoelastic asthenosphere with Maxwell rheology. A uniform shear stress is transmitted to the microplate by the two adjacent plates. This model has the great merit of lending itself to an analytical treatment, which allows us to study those parameters which are relevant in governing microplate dynamics and also to determine the necessary conditions for the occurrence of seismic activity at the microplate margins. We shall consider the case when a rupture occurs at boundary A due to the ambient shear stress exceeding a threshold value and we shall study the subsequent evolution of the state of stress within the microplate with particular attention to boundary B.

The effects of stress relaxation in the asthenosphere become particularly important if the whole lithospheric thickness is ruptured. Complete rupture at a plate boundary over the whole lithospheric thickness may occur suddenly, in connection with a major earthquake, or, more plausibly, in a finite time though a sequence of creep and brittle fracture events (Davies & Brune 1971). There are indeed indications that aseismic slip may play an important role both in California (e.g. Savage & Burford 1973) and in the Mediterranean region (North 1974). We assume however that the time-scale over which the relative plate displacement occurs is much shorter than the characteristic relaxation time \( \tau \) of the asthenosphere, which may be a few years (Nur & Mavko 1974). Moreover, we shall be interested in relatively short time-scales (a few characteristic times at most): otherwise the effect of other earthquakes in the surrounding region on the state of stress at boundary B should be taken into account. A further point to be noted is that stress drop values in the Mediterranean region are comparatively low (North 1977) with respect to other, more active seismic areas. We describe pure strike-slip boundaries, since the mathematical treatment is simpler in this case. While this may be a reasonable model of the Humboldt plate, the results — in particular the time-scale of the phenomena considered here — will not be appreciably
altered if dip-slip or mixed-type relative plate motions are considered as in the case of the Mediterranean region.

2 The model

The lithospheric plates are modelled as elastic slabs of thickness $h$ and shear modulus $\mu_1$ (Fig. 1). The microplate is placed between two larger plates and its horizontal width (the distance between its two boundaries) is $l$. We consider the 2-D problem of two long, parallel plate margins. A viscoelastic half-space representing the asthenosphere is placed under the lithosphere, in welded contact with it. A Maxwell rheology is assumed for the asthenosphere, characterized by a shear modulus $\mu_2$ and a viscosity $\eta$. We assume that the microplate is subjected to a uniform and constant shear strain rate produced by the motion of the larger plates, as indicated by the arrows in Fig. 1:

$$\varepsilon_{z1}(x_2, x_3; t) = K.$$  

Since the relative plate velocity is very low (typically a few centimetres per year) it is safe to assume that the strain rate $\varepsilon_{z1}$ is nearly uniform throughout the microplate. The strain rate $\dot{\varepsilon}_{z1}$ produces a uniform increase in the corresponding shear stress component $\tau_{z1}$ (which we call ‘tectonic’ stress). If we assume that the tectonic stress has the value $\tau_0$ at $t = 0$, then

$$\tau_{z1}(x_2, x_3; t) = \tau_0 + 2\mu Kt.$$  

We consider the possibility that tectonic stress $\tau_{z1}$ is high enough to produce relative strike-slip displacements at the plate margin $A$.

A uniform relative displacement of two plates can be represented by a Volterra dislocation of screw type (Steketee 1958; Maruyama 1964). The solution for the viscoelastic problem is found by starting from the solution for the elastic problem and then applying the correspondence principle (see, e.g. Fung 1965). The displacement field produced by a dislocation occurring in a layered elastic half-space can be obtained from the relations given by Rybicki (1971). We consider a dislocation with uniform amplitude $\Delta u$, reaching a depth $d$ in an elastic layer of thickness $h$ and shear modulus $\mu_1$ (Fig. 2), overlying an elastic half-

Figure 2. Coordinate frame and quantities employed in the equivalent elastic problem.
The displacement field \( u_1^d \) produced by the dislocation in the upper elastic layer is

\[
u_1^d(x_2, x_3) = \frac{\Delta u}{2\pi} \left[ \arctan \frac{x_3 + d}{x_2} - \arctan \frac{x_3 - d}{x_2} + \sum_{n=1}^{\infty} \Gamma^n \left( \arctan \frac{x_3 - 2nh + d}{x_2} - \arctan \frac{x_3 + 2nh - d}{x_2} \right) \right] \]

where

\[
\Gamma = (\mu_1 - \mu_2)/(\mu_1 + \mu_2).
\]

By Hooke's law, the stress component \( \sigma_{21}^d \) is given by

\[
\sigma_{21}^d(x_2, x_3) = \frac{\mu_1 \Delta u}{2\pi} \left[ F_0(x_2, x_3) + \sum_{n=1}^{\infty} \Gamma^n F_n(x_2, x_3) \right]
\]

where

\[
F_0(x_2, x_3) = \frac{x_3 - d}{x_2^2 + (x_3 - d)^2} - \frac{x_3 + d}{x_2^2 + (x_3 + d)^2}
\]

\[
F_n(x_2, x_3) = \frac{x_3 - 2nh - d}{x_2^2 + (x_3 - 2nh - d)^2} - \frac{x_3 - 2nh + d}{x_2^2 + (x_3 - 2nh + d)^2} + \frac{x_3 + 2nh - d}{x_2^2 + (x_3 + 2nh - d)^2}
\]

The time-dependent stress for the viscoelastic problem (in the case of a Maxwell solid) can be now obtained in the space of Laplace transforms by the substitution

\[
\mu_2 = \mu_2 \eta s/(\eta s + \mu_2)
\]

where \( \mu_2 \) and \( \eta \) are the shear modulus and the viscosity of the asthenosphere and \( s \) is the complex variable employed in the Laplace transforms. Since the average shear moduli of the lithosphere and of the asthenosphere are not very different, we assume for simplicity that they are the same, i.e. we take

\[
\mu_1 = \mu_2 = \mu.
\]

Since the time during which the dislocation process occurs is much shorter than the times in which we are interested here (a few years), we may assume that the dislocation takes place suddenly, i.e.

\[
\Delta u(t) = \Delta u H(t)
\]

where \( H(t) \) is the Heaviside step function. The Laplace transform of (10) is

\[
\Delta \tilde{u}(s) = \Delta u/s.
\]
Figure 3. Shear stress at the Earth’s surface as a function of the distance from boundary A. Different curves refer to different times. The uniform stress increment due to tectonic loading is omitted to emphasize the stress diffusion effect.

We now use (8), (9) and (11) in equation (5) and obtain the Laplace transform $\tilde{\sigma}_{21}(x_2, x_3; s)$ of the time-dependent stress field:

$$\tilde{\sigma}_{21}(x_2, x_3; s) = \frac{\mu \Delta u}{2\pi s} \left[ F_0(x_2, x_3) + \sum_{n=1}^{\infty} \tilde{\Gamma}^n(s) F_n(x_2, x_3) \right]$$  \hspace{1cm} (12)

where

$$\tilde{\Gamma}(s) = (\tau s + 1)^{-1}$$  \hspace{1cm} (13)

and we have put

$$\tau = 2\eta/\mu.$$  \hspace{1cm} (14)

Equation (10) is easily inverted to give

$$\sigma_{21}(x_2, x_3; t) = \frac{\mu \Delta u}{2\pi} \left[ F_0(x_2, x_3) + \sum_{n=1}^{\infty} A_n(t/\tau) F_n(x_2, x_3) \right]$$  \hspace{1cm} (15)

where $t > 0$ and

$$A_n(t) = \frac{1}{(n-1)!} \int_0^t \xi^{n-1} \exp(-\xi) d\xi$$  \hspace{1cm} (16)

are incomplete gamma functions. The coefficients $A_n(t)$ can be written as finite series (Abramowitz & Stegun 1964):

$$A_n(t) = 1 - \exp(-t) \sum_{k=0}^{n-1} \frac{t^k}{k!}.$$  \hspace{1cm} (17)
Figure 4. (a) Shear stress at the Earth's surface as a function of time, for different distances from boundary A, in the absence of tectonic stress ($K = 0$ in equation 18). (b) Shear stress at the Earth's surface as a function of time, for different distances from boundary A and $4K\eta/\sigma_o = 10^{-1}$ (see equation 18).
Figure 5. (a) Shear stress at boundary B as a function of depth $x_3$, in the absence of tectonic stress ($K = 0$ in equation 18) and for a microplate width $l/h = 1$, where $h$ is the lithosphere thickness. Different curves refer to different times, after the earthquake occurrence at boundary A, in terms of the asthenosphere characteristic relaxation time $\tau$. (b) The same as (a), for $l/h = 2$. (c) The same as (a), for $l/h = 3$. (d) The same as (a), for $l/h = 4$. 
A very good approximation of the stress values is obtained by taking the first 100 terms in the series appearing in (15).

3 Time evolution of shear stress in the microplate

As shown in Fig. 1(b), we assume that the dislocation event involves the entire plate thickness, i.e. $d = h$ in (3), (6) and (7). With the term 'dislocation event', we also include a sequence of brittle and aseismic faulting processes, provided they occur on a time-scale much shorter than the characteristic relaxation time $\tau$ of the asthenosphere.

In Fig. 3, the shear stress at the Earth's surface is plotted as a function of the horizontal distance $x_2/h$ from boundary A. By assumption, stress drops to zero at the time of the earthquake ($t = 0$) at $x_2 = 0$ and is gradually recovered as time goes on. It is interesting to note that stress increases monotonically for $x_2/h \leq 2$, while it has an initial decrease at larger distances. This is made evident in Fig. 4(a), where the shear stress at boundary B is plotted as a function of time for different microplate widths $l$, in the absence of tectonic strain rate (i.e. $K = 0$ in equation 2). Stress is fairly rapidly increasing when $l/h = 1$, but it decreases at first already for $l/h = 2$ and the subsequent stress recovery is extremely slow. This kind of behaviour is enhanced for larger microplate widths. In Fig. 4(b) a tectonic strain rate component has been added, according to the formula

$$\sigma_{\text{tot}}(x_2, x_3; t) = \sigma_0 + \sigma_{21}^d(x_2, x_3; t) + 2\mu K t.$$ (18)

Figs 3 and 4 show the effect of stress diffusion which takes place perpendicularly to the plate boundary A, after an earthquake has occurred there. The phenomenon of stress diffusion was initially suggested in the framework of models with a purely viscous asthenosphere (Elsasser 1969; Bott & Dean 1973; Melosh 1976). In the more realistic viscoelastic model for the asthenosphere, stress diffusion is less apparent (see also Savage & Prescott 1978b), but it will be seen later in this paper that diffusion still plays an important role at some distance from boundary A.

In Fig. 5(a–d) the shear stress at boundary B is plotted as a function of depth at various times after the earthquake at boundary A. For all microplate widths $l$, the coseismic stress drop ($t = 0$) is greater at the Earth's surface than at some depth. We note that the effect of a preliminary stress decrease due to diffusion from the boundary A is enhanced at depth with respect to the Earth's surface and is already present at the base of the lithosphere for $l/h = 1$ (Fig. 5a).

4 Discussion

The results give rise to the following considerations. We have seen (Fig. 5) that $\sigma_{21}^d(x_2, x_3; t)$ does not vary much with depth $x_3$ when $t \geq \tau$ and, in any case, when $x_2/h \geq 2$. Since for the microplate width $l$ we can reasonably assume $l/h \geq 2$ at least, for the sake of simplicity we shall deal in the following with stress values at the Earth's surface only, $x_3 = 0$. We have assumed that the initial stress field (at $t = 0$) is uniform throughout the microplate (equation 2) and that the plate boundary A ruptures first, because of its lower strength (Fig. 1b). Alternatively, we might assume that the two boundaries have the same strength, but the initial stress field is lower at boundary B than at boundary A. The same conclusions are reached in both cases. For definiteness, we consider the first case: let $\sigma_0$ be the strength of boundary A and $\sigma_0 + \delta \sigma$ be the strength of boundary B, with $\delta \sigma > 0$. An earthquake occurring at boundary A at $t = 0$ produces a coseismic stress drop at the Earth's surface

$$\Delta \sigma(x_2) = -\sigma_{21}^d(x_2, 0; 0)$$ (19)
Figure 6. (a) The time $T_B$ required for earthquake occurrence at boundary B, as a function of the strength difference $\delta \sigma$ between boundaries A and B, for the elastic half-space model. Different curves refer to different microplate widths $l/(4K\eta/\sigma_0 = 10^{-1})$. (b) The time $T_B$ required for earthquake occurrence at boundary B as a function of the strength difference $\delta \sigma$ between boundaries A and B, for the two-layer viscoelastic model (solid curves). The dashed curves are the same appearing in (a) and are shown for comparison. All curves are labelled with the corresponding microplate width $l/h (4K\eta/\sigma_0 = 10^{-1})$. 
which lowers the stress level at boundary B by a quantity

\[ \Delta \sigma(l) = \frac{\mu \Delta u}{\pi} \frac{h^2}{l^2 + h^2} \]  

(20)

Let us call \( T_A \) and \( T_B \) the time intervals required for the occurrence of the next earthquake at the boundaries A and B, respectively. That is

\[ \sigma^{\text{tot}}(0, 0; T_A) = \sigma_0 \]  

(21)

\[ \sigma^{\text{tot}}(l, 0; T_B) = \sigma_0 + \delta \sigma. \]  

(22)

Under these assumptions, the condition for the occurrence of the next earthquake at the microplate boundary B can be stated as

\[ T_B < T_A, \]  

(23)

otherwise the next earthquake will occur again at boundary A.

4.1 Elastic Half-Space Model

Let us first consider the case of no stress relaxation in the asthenosphere. If no earthquake took place at boundary A, the shear stress at boundary B would reach the threshold value \( \sigma_0 + \delta \sigma \) when \( t = T \), where, from equation (2),

\[ T = \delta \sigma/(2\mu K). \]  

(24)

Since, however, boundary A is weaker than boundary B by assumption, the time \( T_B \) needed to reach the same threshold value \( \sigma_0 + \delta \sigma \) after an earthquake has occurred at boundary A is, in the absence of stress relaxation,

\[ T_B = \frac{1}{2\mu K} \left[ \delta \sigma + \Delta \sigma(l) \right] \]  

(25)

from (18) and (22). According to (23), \( T_B \) must be less than

\[ T_A = \sigma_0/(2\mu K) \]  

(26)

from (18) and (21). In Fig. 6(a), \( T_B \) is plotted as a function of the strength difference \( \delta \sigma \) between the two boundaries, for several microplate widths \( l \). It can be seen that the relative importance of the stress change produced at boundary B by earthquakes occurring at boundary A is determined by the ratio \( \delta \sigma/\sigma_0 \): from Fig. 6(a), a necessary condition for the next earthquake to occur at boundary B according to (23) is in fact \( \delta \sigma < \sigma_0 \). If, for instance, \( \delta \sigma/\sigma_0 = 1/100 \), the stress change \( \Delta \sigma(l) \) at boundary B is 10 times greater than \( \delta \sigma \). For \( l/h = 3 \) the coseismic stress drop \( \Delta \sigma(l) \) at boundary B is 10 times greater than \( \delta \sigma \), from (20). From (25) \( T_B \) is 11 times longer than \( T \). Thus, obviously, an earthquake at a microplate margin delays the earthquake occurrence on an adjacent parallel margin, as considered in the present model. If, however, the difference in strength \( \delta \sigma \) between the two plate margins is greater, the time delay is smaller. For instance, if \( \delta \sigma/\sigma_0 = 1/10 \), \( T_B \) is only two times longer than \( T \).
Figure 7. The earthquake recurrence time $T_A$ at boundary $A$, as a function of the asthenosphere viscosity $\eta$ (see equations 21 and 18, with $x_2 = x_3 = 0$); $\tau_o$ is the asthenosphere characteristic relaxation time (equation 14) with $\eta = 5 \times 10^{19}$ Pa and $\mu = 6 \times 10^{11}$ dyne cm$^{-2}$. The different curves refer to different values of the ratio $K/\sigma_o$. If $\sigma_o = 10$ bar, the curves refer to $K$ values from $10^{-15}$ to $10^{-14}$ s$^{-1}$. The dashed line $T_B = 10 \tau_o$ is drawn to illustrate an example given in the text.

4.2 TWO-LAYER VISCOELASTIC MODEL

Let us now consider the effect of stress relaxation in the asthenosphere. Another element now enters the problem, i.e. the ratio $T_B/\tau$, which makes the constraint (23) more stringent for the occurrence of seismic activity at boundary B. If we take the stress drop $\sigma_B$ at boundary A of the order of 10 bar, which is a plausible value for the Mediterranean region (North 1977), $\Delta\sigma(l)$ is typically 1 bar for $l$ of the order of a few lithospheric thicknesses (Fig. 3); if moreover the tectonic strain rate $K$ is taken as $0.5 \times 10^{-14}$ s$^{-1}$ (North 1974) and $\eta = 5 \times 10^{19}$ Pa (Nur & Mavko 1974), we have $T_B/\tau = 1$. The fact that $T_B$ and $\tau$ are of the same order of magnitude makes stress relaxation an important factor in the earthquake mechanism at plate boundaries.
The above values for $\sigma_0$, $K$ and $\eta$ have been employed to draw Figs 4(b) and 6(a, b). The last two figures show the time required for earthquake occurrence at boundary B for the elastic half-space model and for the two-layer viscoelastic model, respectively. These delay times are plotted as functions of the strength difference $\delta \sigma$, for several values of the microplate width $l$. Due to stress relaxation in the asthenosphere, now $T_A$ is significantly less than $T_A$. In Fig. 7, the recurrence time $T_A$ is plotted as a function of the asthenosphere viscosity $\eta$. The different curves refer to different values of the ratio between the tectonic strain rate $K$ and the stress drop $\Delta \sigma$. It can be seen that $T_A$ shows the largest variations corresponding to the plausible values for the asthenosphere viscosity.

From a comparison between Figs 6(a) and 6(b), it is easy to see that the condition $T_B < T_A$ lowers significantly the admissible $T_B$ values with respect to the corresponding $T_B$ values. Furthermore, while in the elastic half-space model $\delta \sigma$ can be as high as $\sigma_0$, when $T_B$ approaches $T_A$, the constraints on $\delta \sigma$ are more severe in the case of the viscoelastic model. All the above-mentioned effects depend on how large $l$ is. Referring to Fig. 6(b), the asthenosphere relaxation delays the earthquake occurrence time at boundary B for $l/h > 2$, while the opposite holds for $l/h \leq 2$. It can be seen from Fig. 3 that at distances $l \geq l_0$, the decrease in stress due to the asthenosphere relaxation may be of the same order of magnitude as the coseismic stress drop $\Delta \sigma$. In a microplate, where both boundaries A and B are seismically active, $T_A$ may be either interpreted as a lower bound for the earthquake recurrence time on boundary A or, according to condition (23), as an upper bound for $T_B$. Referring to Fig. 7, the observed $T_B$ value might provide both an upper bound for $K/\sigma_0$ and a lower bound for the local value of the asthenosphere viscosity. For instance, if $T_B = 10 \tau_0$, $K/\sigma_0$ must be lower than $10^{-21}$ cm$^2$ (dyne s)$^{-1}$; moreover, if from other considerations $K/\sigma_0$ is plausibly higher than, say, $10^{-22}$ cm$^2$ (dyne s)$^{-1}$, $\eta$ must be higher than about $10^{20}$ P.

5 Conclusions

The model proposed in the present paper allows us to study in detail the stress evolution at a microplate margin in response to the stress perturbation produced by an earthquake at a parallel margin, in the presence of a viscoelastic asthenosphere. The main result obtained is that, for plausible microplate widths, the asthenosphere relaxation acts in the same sense as the coseismic stress drop in lowering the stress level at the microplate boundary and so delays the occurrence time of the next earthquake on that boundary. For seismic activity to occur at both microplate boundaries, a distance $l/h \geq 2$ apart, one must have both a low relative strength difference (or initial stress field inhomogeneity) between the two boundaries, i.e. $\delta \sigma/\sigma_0 \approx 10^{-1}$, and a sufficiently high tectonic strain rate $K$. As to the latter point, earthquake occurrence at a microplate boundary (B in the model) can take place only if the tectonic strain rate is high enough for the rupture stress to be reached in a relatively short time after an earthquake has occurred at a weaker boundary (A); otherwise after some relaxation times $\tau$, stress diffusion will make the stress level in the region almost uniform (Fig. 3). With $\eta = 5 \times 10^{19}$ P and $\mu = 6 \times 10^{41}$ dyne cm$^{-2}$, which is the average for lithosphere and asthenosphere, we have $\tau \approx 5.2$ yr for a possible Maxwell relaxation time. In summary, the effects of the asthenosphere relaxation in determining the earthquake occurrence at boundary B are found to be relevant only if the coseismic stress drop there is at least of the same order of magnitude as the stress gap to be filled for rupturing and if the time required for tectonic stress increase to fill this gap is not smaller than the Maxwell relaxation time of the asthenosphere.

Of course, the model proposed in this paper is only an idealized description of existing microplates. The model is 2-D and rupture is assumed to take place along the entire micro-
plate margin. Furthermore, most existing microplates are in compressive regime (e.g. those in the Mediterranean area) while the model assumes two transform margins. However, this model, while lending itself to an analytical treatment, offers the opportunity of understanding the role played by different parameters in determining the microplate dynamics. One can easily be convinced that the previous quantitative conclusions may be at least qualitatively extended to actual microplates.

For a quantitative study of real microplate margins, one should consider more complex geometries, finite fault extensions and compressive tectonic environments. This can possibly be achieved by use of numerical methods. A major difficulty is, however, our poor knowledge of the direction and intensity of the tectonic forces exerted on the microplates. These should be inferred by studies of focal mechanisms, but in many cases satisfactory results are difficult to obtain. A further difficulty in matching theoretical models to observations is that the microplate boundaries are often not well defined. In any case, the attempts to develop detailed models of microplate interactions seem to be the only practicable way to understand the complex geodynamical processes which occur in the intensively fragmented regions typical of the Mediterranean area.

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