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Stress Transfer and Nonlinear Stress Accumulation at the North Anatolian Fault, Turkey

G.-AKIS TSELENTIS[1] and J. DRAKOPOULOS[2]

Abstract—An analysis is presented of the accumulation of stress along the North Anatolian fault. The analysis is based on the time-dependent reloading of the plate boundary by using a modified Elsasser model of a coupled lithosphere-asthenosphere system.

It is found that many of the North Anatolian fault earthquakes are likely to have been triggered by adjacent ruptures, while the time intervals between large earthquakes may have been partly modulated by the relaxation of the viscoelastic asthenosphere.

Key words: Stress, North Anatolia, Turkey, Elsasser.

1. *Introduction*

The degree to which stress builds up between major earthquakes has important implications for the timing of earthquake occurrences and the physical mechanism of crustal deformation (strain accumulation).

In fact the two most striking features of crustal deformation are its variability in time and its tendency to localize at specific areas of the crust, which mainly coincide with the tectonic plate boundaries. Earthquakes occurring at plate boundaries involve finite segments of the plate boundary which undergo sudden stress drop accompanied by relative slip between the broken fault surfaces.

Observations suggest that consecutive ruptures seem to abut without much overlap and that large ruptures appear to migrate along a plate boundary (e.g., AMBRASEYS, 1979; LI and KISSLINGER, 1985).

Various models have been constructed to study the different phases of the earthquake cycle, including elastic strain accumulation, coseismic release and postseismic readjustment. LEHNER *et al.* (1981) studied the effect of lithosphere/asthe-

[1] University of Patras, Dept. of Geophysics and Applied Geology, Laboratory of Theoretical & Applied Seismology, Patras 261 10, Greece.

[2] University of Athens, Geophysics Dept., Athens 11701, Greece.

nosphere coupling on stress redistribution due to a finite length rupture. The study was based on a modified Elsasser model (ELSASSER, 1969), which includes the viscoelastic response of the asthenosphere. LI and KISSLINGER (1985) extended this work to study the stress transfer along a converging plate boundary.

According to the Li-Kisslinger model, each event produces stress changes in the asthenosphere, which is treated as a viscoelastic substrate, and the latter transfers a time-dependent load back onto the lithosphere. Hence, immediate stress transfer through coseismic elastic stress imposition would be followed by more gradual stress transfer through asthenosphere relaxation. By applying this technique to the Aleutians, LI and KISSLINGER (1985) showed that stress accumulation in the Aleutians is highly nonlinear and that some large earthquakes may have been triggered by nearby ruptures.

The above method of analysis is particularly suitable to model the local stress field processes of the North Anatolian fault in Turkey, because the fault system and the character of fault movements are very well documented and indicate the existence of this triggering effect (ALLEN, 1969; AMBRASEYS, 1979).

2. *Tectonic Features of the North Anatolian Fault System*

Most of the seismic activity in northern Turkey is associated with the North Anatolian fault, a right-lateral strike-slip fault which forms the boundary between the Eurasian and the Anatolian plates (Fig. 1). The origination of the fault system is set during the mid to late Miocene (SENGOR and CANITEZ, 1982), and since then the fault has accumulated an offset of about 85 km.

Fault plane solutions of earthquakes along the fault between 31°E and 41°E give consistently dextral slip with minor thrust component. The fault trace is morphologically well developed for most of its length and can be identified along the 1000 km long central portion, between longitudes 31°W and 41°E. To the west of 31°E longitude the fault appears to break into two or three branches while the earthquake mechanisms in this region show a N-S tension component as well.

To the east, the morphology of the fault changes at its junction with the East Anatolian fault (41°E) and the fault plane solutions indicate an increased component of thrusting consistent with left lateral slip on the East Anatolian fault (ARPAT and SAROGLU, 1972; MCKENZIE, 1976).

Recent attempts have been made to explain the seismotectonic features of this fault in terms of global tectonics and infer that a major part of the thrust between Arabian and Turkish plates is taken up along the North Anatolian fault between two triple junctions as right lateral motions (DEWEY *et al.*, 1986).

Geological as well as geophysical data indicate that the current average slip rate of the fault is about 1.5–2 cm/yr (CANITEZ and EZEN, 1973; DEWEY *et al.*, 1986).

Beginning with the catastrophic Erzincan earthquake of 28/12/1939, a source migration from east to west with a velocity of 50–100 km/yr has been attributed to a deep-seated east to west propagation of high tectonic strain (ALLEN, 1969; AMBRASEYS, 1979). The faulting was accomplished in a series of earthquakes of magnitude 6 to 8, with each earthquake extending the surface rupture westward along the fault from the rupture of the previous large earthquake (Figure 1).

In Figure 1, the distribution of earthquakes with $M > 7$ along the east-west trend of the fault versus time and space is shown. This figure shows that the westward migration of the earthquakes is fastest immediately after the 1939 earthquake and slows after 1945. However, no single velocity of source migration

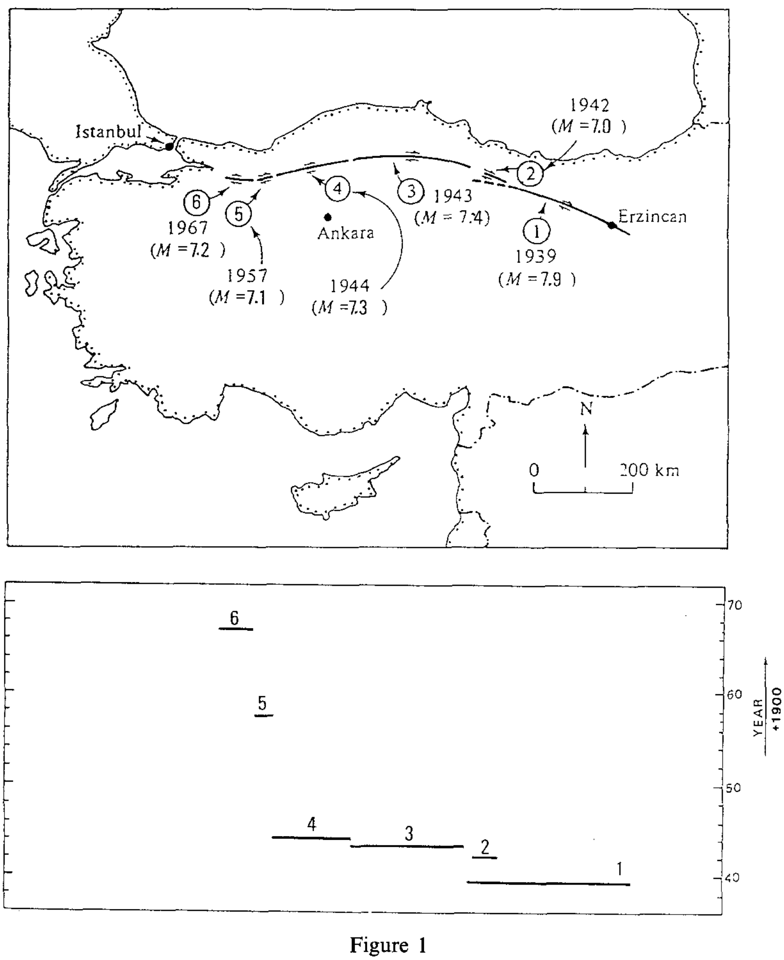


Figure 1

Top Map: the North Anatolian Fault with epicenters of the earthquake series (marked with location numbers), that ruptured the plate boundary segments studied in the present paper. Bottom: Space time plot of the events with corresponding rupture lengths.

can be found to predict the time of occurrence of earthquakes of magnitude 6 or greater on the North Anatolian fault.

Attempts also to interpret events of magnitude 6 or greater after the 1939 event as being triggered by local (coseismic) stress increases resulting from the occurrence of previous large earthquakes have indicated the complexity of the seismic behavior of the fault system (DEWEY, 1976).

In this paper, an alternative approach to investigate the seismic behavior of the North Anatolian fault will be followed by studying stress changes across the fault as the result of both asthenospheric relaxation processes and coseismic stress effects.

3. *Method of Analysis*

In the following we present a brief description of the method of analysis which was applied to the North Anatolian fault. More details of the method can be found elsewhere (LI, 1981; LI and RICE, 1983; LI and KISSLINGER, 1985).

According to the above method the lithosphere is described as a linear elastic plate of uniform thickness H, riding on a viscoelastic asthenosphere of thickness h (Figure 2).

Each earthquake results in instantaneous stress changes in the viscoelastic (Maxwell) substrate which then relax with time, transferring a shear load back onto the lithosphere with a rate of stress relaxation dependent on the time constant as will be explained below.

Let $\sigma_{xx}, \sigma_{xy}, \sigma_{yy}$ be the thickness-averaged in-plane stress components defined by equation (1):

$$\sigma_{ij}(x, y, t) = (1/H) \int_0^H \tau_{ij}(x, y, z, t)\, dz \quad i, j = x, y \tag{1}$$

where τ_{ij} is the three-dimensional stress distribution. Then, obviously equilibrium requires that

$$(\partial \sigma_{ix}/\partial x) + (\partial \sigma_{iy}/\partial y) = \tau_i/H \quad i = x, y \tag{2}$$

where $\tau_i = -\tau_{zi}(x, y, H, t)$ is the resistive shear stress at the base of the lithosphere.

According to the stress-strain relations of an isotropic material, the plane stress and strain components are related as follows

$$\sigma_{ij} = G\left[(\partial u_i/\partial j) + (\partial u_i/\partial i) + (2v/(1-v))\delta_{ij} \sum_{k=x,y} \partial u_k/\partial k)\right] \tag{3}$$

where $i, j = x, y$, δ_{ij} is the Kronecker delta and G, v are the elastic shear modulus and Poisson ratio for the lithosphere.

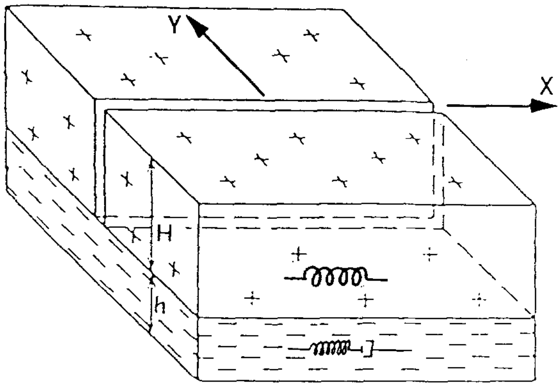


Figure 2
Coupled lithosphere-asthenosphere viscoelastic system.

Since the viscosity of the earth increases abruptly below some depth, it is logical to treat the asthenosphere as a thin channel of thickness h which can be described (for a Maxwell model) by the following stress-displacement relation

$$\partial u_i/\partial t = (b/G)\partial \tau_i/\partial t + (h/\eta)\tau_i. \tag{4}$$

Hence, eq. (3) can be written

$$\left(\alpha + \beta \frac{\partial}{\partial t}\right) \sum_{k=x,y} \left[\frac{\partial^2 u_j}{\partial k^2} + \frac{1+v}{1-v} \frac{\partial^2 u_k}{\partial k\, \partial j}\right] = \frac{\partial u_j}{\partial t} \tag{5}$$

where $a = hHG/\eta$, $\beta = bH$ and $b \simeq (\pi/4)^2 H$ and the characteristic relaxation time is given by the ratio $\beta/a = (\eta/h)/(G/b)$.

For the particular case of strike-slip plate boundaries LEHNER *et al.* (1981) showed that the displacements k, j in (5) can be uncoupled so that the equation takes the following form

$$(\alpha + \beta\, \partial/\partial t)[(1 + v^2)\, \partial^2 u_x/\partial x^2 + \partial^2 u_x/\partial y^2] = \partial u_x/\partial t \tag{6}$$

with an associated shear stress along the fault line, given by

$$\sigma_{xy} = G\, \partial u/\partial y. \tag{7}$$

Modeling an earthquake as a dislocation of length L and a sudden stress drop q, the dislocation magnitude is given by (LEHNER *et al.*, 1981)

$$\begin{aligned} \delta(x, 0^+) &= \frac{29}{G}\left(\frac{\beta}{\pi}\right)^{1/2} \int_0^{\lambda_0|x|} \frac{e^{-\rho}}{\rho^{1/2}} \operatorname{erf}\,[[\rho + \lambda_0(L - |x|)]^{1/2}]\, d\rho, \quad -L \le x \le 0 \\ \delta(x, 0^+) &= \frac{29}{G}\left(\frac{\beta}{\pi}\right)^{1/2} \int_0^{\lambda_0 x} \frac{e^{-\rho}}{\rho^{1/2}} \operatorname{erf}\,[[\rho - \lambda_0(|x| - L)]^{1/2}]\, d\rho, \quad x < -L \end{aligned} \tag{8}$$

where $\lambda = \{2/[\beta(1-v^2)(1+v)]\}^{1/2}$ and assuming the following boundary condition

$$\begin{aligned} u(x, 0^+, t) &= \delta(x)H(t)/2, \quad x < 0 \\ u(x, 0^+, t) &= 0, \quad x > 0 \end{aligned} \tag{9}$$

eq. (6) yields

$$\sigma_{xy}(x, 0, t) = \frac{(1+v)G}{2\pi} \int_{-\infty}^{0} \tau(|x-\xi|, t) \frac{\delta(\xi)\, d\xi}{|x-\xi|} \tag{10}$$

where

$$\tau(|x-\xi|, t) = \lambda_0 e^{-\alpha t/\beta} \left[K_1(|x-\xi|\lambda_0) + \sum_{v=1}^{\infty} \frac{(|x-\xi|\lambda_0 \alpha t/\beta)^v}{2^v (v!)^2} K_{v+1}(|x-\xi|\lambda_0) \right]. \tag{11}$$

This equation provides the thickness averaged stress change at any point x along the fault boundary and at any time t, once the stress drop q and rupture length L are known.

4. Data Analysis

The data used to investigate the above-mentioned stress simulation at various places along the North Anatolian fault are compiled in Table 1. For the events reported in Table 1, the rupture lengths are based primarily on off-sets of the surface of the earth as documented in the literature (e.g., ALLEN, 1969; AMBRASEYS, 1979).

As has been noted from aftershock studies in tectonically similar areas (e.g., San Andreas fault—California, Motagua fault—Guatemala), the fault length estimation from aftershock distributions are usually 25–50% greater than those estimates based on the study of surface faulting (SYKES and QUITTMEYER, 1981). Hence, the rupture lengths reported in Table 1 may underestimate the real rupture lengths.

To estimate the stress drops of the events used in the present analysis, the following equation (CHINNERY, 1969) was used

$$\Delta\sigma = C\mu U/W \tag{12}$$

where $\Delta\sigma$ is the stress drop on a dislocation surface of width W and displacement U, embedded in a medium of shear modulus μ, and where C is a numerical factor related to the shape of the fault which is determined by the solution of the crack problem in which constant stress is applied over the surface of the fault.

Simple fault models suggest that for many data sets the parameter C may be taken to be a constant. For an infinite homogeneous medium with a uniform jump

Table 1

Location, magnitude, fault length, dislocation stress drop of the earthquakes used. Material properties are also listed.

No	Yr	M	D	LAT		LON		M	L(km)	U(cm)	$\Delta\sigma$(bar)
1	1983	12	26	39	80	39	51	7.9	340	370	20.72
2	1942	12	20	40	87	36	47	7.0	50	175	9.80
3	1943	11	26	41	05	33	72	7.4	280	150	8.40
4	1944	02	01	40	08	33	02	7.3	180	350	19.60
5	1957	05	26	40	67	31	00	7.1	40	160	9.00
6	1967	07	22	40	67	30	69	7.2	80	160	9.00

Poisson ratio $v = 0.25$. Shear modulus $G = 5.5 \times 10^{10}$ Pa. Asthenosphere viscosity $= 2 \times 10^{19}$ Pa.s.

in strike-slip displacement over the entire dislocation surface for which $L \geq 2W$, C is about 0.32 (CHINNERY, 1969).

For the events used in the present analysis, a rupture width of 20 km is assumed. This is a reasonable estimate and coincides with the estimates based (whenever available), on the spatial distribution of aftershocks (e.g., CRAMPIN, 1969; NORTH, 1977), and surface wave modeling (KUDO, 1983).

The obtained stress drops are tabulated in Table 1. These values show that the stress drops encountered with the above events are relatively low and considering the dynamical processes of fault motion in the presence of friction (YAMASHITA, 1976), indicate that the residual shear strength on the presheared North Anatolian fault zone is very small.

For the 1967 event, there is also an estimate of stress drop obtained by the inversion of long period P, S waves (HANKS and WYSS, 1972) giving a value of the order of 10 bar which is in good agreement with the value shown in Table 1.

Next we will use the model described in Section 2, to simulate the stress-strain accumulation at 5 locations along the North Anatolian fault, chosen to coincide with the epicenters of the earthquakes tabulated in Table 1.

In the following calculations, we assume a regional tectonic loading rate $\dot{\delta}_0$ of the order of 0.05 bar/yr, which is considered to be a reasonable estimate for the area (DEWEY *et al.*, 1986).

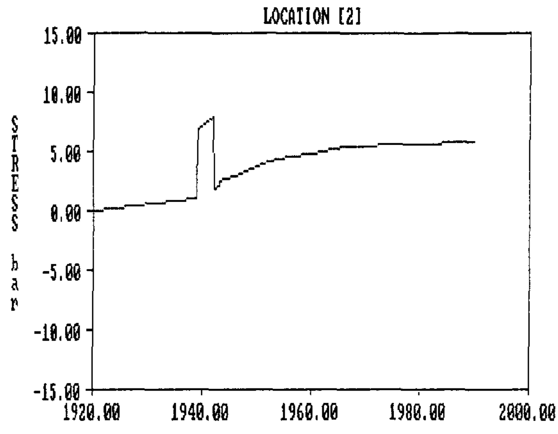
If $\sigma_i(x, t)$ denotes the stress change at the above considered points ($i = 2, 3, 4, 5, 6$), and at time t due to a rupture located between x and $x - L_j$, where L_j is the rupture length, which occurred at time t_j with stress drop q_j, then the overall stress change at the above points can be estimated from the following formula

$$\sigma_i(x, t) = \sum_j \sigma_i(x - x_j, t - t_j, q_j, L_j) + \dot{\sigma}_0 t. \tag{13}$$

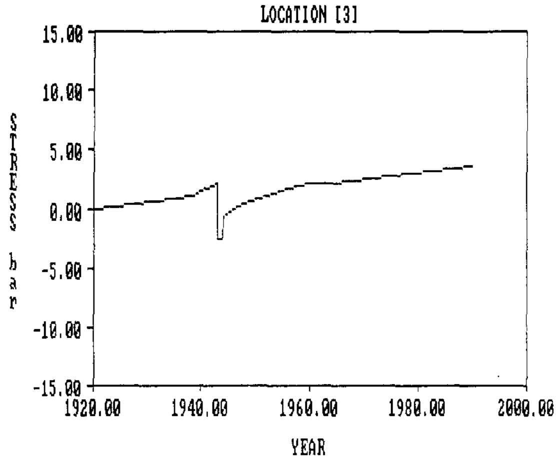
5. Discussion of the Results

The estimated thickness-averaged stresses at different points along the fault, as a function of time, are shown in Figure 3. The year 1920 is chosen as the reference year, when stress is arbitrarily taken to have a zero value at all locations. The influence of any previous events might not affect the accuracy of our calculations since the 20-year period prior to the 1939 event was a period of low activity and the average relaxation time, assuming a lithospheric thickness of 40 km, is of the order of 10 years.

In the interpretation of the current stress state, one must keep in perspective that assuming a different value for the existing tectonic loading does affect considerably the stress level. However, choosing a different value for tectonic loading does not change the general trend of stress accumulation due to lithospheric relaxation.

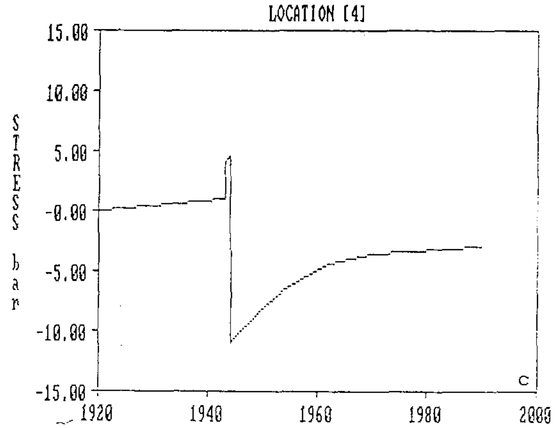


(a)

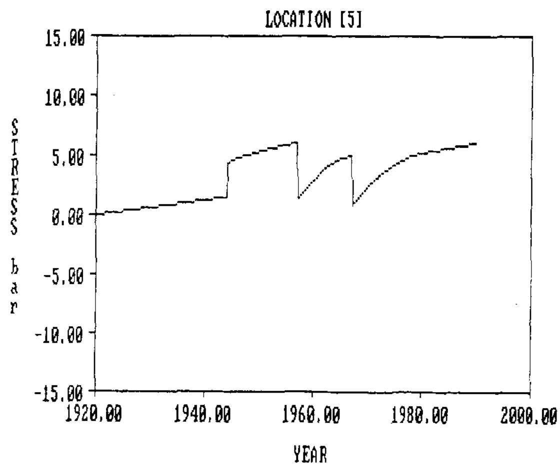


(b)

Figure 3 (a, b)



(c)



(d)

Figure 3
Thickness averaged stresses at the locations marked in Figure 1.

Figure 3a, shows the thickness-averaged-stress alteration at location 2, which coincides with the epicenter of the 1942 earthquake. According to our model, the 1939 rupture to its east produced a coseismic stress jump of about 6 bar at location 2. Assuming a critical stress level at the above site, for which rupture must occur, then the 1942 event would not have happened until about 2070, in the absence of the 1939 event, assuming a tectonic rate of .05 bar/year and until about 1980 assuming a tectonic rate of 0.1 bar/year. Thus, stress transferred from the 1939 earthquake accelerated the 1942 earthquake. The 1943 event, which occurred to the west of site 2, caused a very small stress change at site 2, while the 1944, 1957 and 1967 events did not measurably affect the site, perhaps because of their long distance and their small stress drops. If stress accumulation at site 2 continues at the current rate, it might be speculated that another rupture would occur around 2045, assuming that the location is not once again prematurely stressed by a large earthquake on a neighboring fault segment.

In Figure 3b, the thickness-averaged stresses at location 3 are depicted. The 1942 event seems not to have changed the stress level at the location 3, probably due to its small rupture length, while the 1939 event produced a very small coseismic stress change of about 0.50 bar. This figure shows that the 1943 earthquake might have occurred at about the same time, in the absence of the previous events, releasing the stress accumulated by slow tectonic loading. This area might be regarded as one with high seismic potential, considering the coseismic stress jump due to the 1944 earthquake.

Figure 3c shows the thickness-averaged stress level at location [4], which coincides with the epicenter of the 1944 earthquake. The coseismic stress jump due to the nearby 1943 rupture is evident, indicating that the 1944 earthquake was likely triggered by the 1943 rupture. On the other hand, the 1957 and 1967 events seem to have no effect at location 4 which, considering the current stress rate, might be regarded as an area of low seismic potential.

Figure 3d shows the thickness-averaged stresses at the point centered between locations 5 and 6, where the 1957 and 1967 events occurred. The effect of the 1943 event is very small, resulting in a coseismic stress of about 0.20 bar, while the 1944 event produced a coseismic stress of about 3 bar and probably accelerated the 1957 event.

The 1967 earthquake seems to have occurred when the stress almost recovered to the 1957 failure stress level. Although in the present analysis we have assumed that the 1957 and 1967 earthquakes involved slip on the same fault segment, the 1967 rupture also involved new rupture further to the west. Hence, the 1967 shock cannot be considered as being simply a repeat of the 1955 shock. Also, the seismic history of this region does not suggest that the region experiences such earthquakes every decade or so.

6. Conclusions

A general observation from the above results is that asthenospheric relaxation and coseismic stress jumps seem to modify considerably the stress accumulation process along the North Anatolian fault zone.

The analysis also shows that some of the earthquakes along the fault zone are likely to have been triggered by adjacent ruptures.

An assumption that might affect the results of the present analysis is the assumption that each fault is locked immediately after it produces an earthquake, so that it immediately begins reaccumulating stress (some of which is due to asthenospheric relaxation and to slip on adjacent segments). An alternative might be that a given fault segment might take time to reheal following a large earthquake, so that asthenospheric relaxation would manifest itself as aseismic "afterslip."

The obtained variation of failure stress along the fault zone indicates its inhomogeneity in stress distribution and makes it a favorable object for the use of more advanced "line spring" models (e.g., DMOWSKA and LI, 1982; TSE *et al.*, 1985), by simulating the fault as composed of locked patches and freely slipping parts.

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