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Abstract

Postseismic surface deformations are attributed to the inelastic flow of the subcrustal regions of the earth following an earthquake. A multilayer representation of the earth's rheological properties is used in conjunction with a finite element computational scheme to calculate time-dependent displacements and strains subsequent to a strike-slip earthquake. The deviatoric stress-strain relation for the uppermost layer is assumed elastic. Lower layers are assumed to be, in order of increasing depth, a standard linear, three-element, viscoelastic solid; a linear, viscoelastic fluid; and another elastic solid. Physically these layers correspond to the upper lithosphere, lower lithosphere, asthenosphere, and lower mantle, respectively. Elastic dilatational properties are assumed throughout. Appreciable postseismic displacements, possibly approaching meters, for large earthquakes, arise from the viscoelastic relaxation following the sudden coseismic slip. Furthermore, compared to the simpler case of an elastic lithosphere over a viscoelastic asthenosphere the near-fault postseismic shear strain is increased, by a factor of two or more in some cases, by the presence of a viscoelastic lower lithosphere. Also the duration of postseismic straining is increased if the viscosity of the lower lithosphere is greater than that of the underlying asthenosphere.

The time dependent deformation of the earth following an earthquake has frequently been studied in terms of the rheological properties of the crust and upper mantle. In particular, several studies have attributed surface deformation to viscoelastic relaxation of the asthenosphere following the sudden stress change associated with the coseismic slip. The most commonly employed model has been that of an elastic layer (lithsophere) lying over a viscoelastic asthenosphere. The deviatoric stress-strain relation has been taken to be that of a linear Maxwell body (Nur and Marko, 1974; Rundle and Jackson, 1977; Thatcher and Rundle, 1979) or a Maxwell body with strain rate proportional to a non-unity power of stress (Melosh, 1976). Other models have examined the role of fault slip at depth (Thatcher, 1974), flow of low viscosity magma regions (Wahr and Wyss, 1980), and anelastic relaxation of the lithosphere (Cohen, 1980) on postseismic rebound. Other factors that might influence coseismic and/or postseismic deformation include, but are not limited to, fault geometry, spatial variations in rigidity (Mahrer and Nur, 1979), and spatial variations in coseismic slip.

The conceptual understanding of the influence of the earth's rheological properties on postseismic rebound has been influenced by studies of the creep properties of rocks (see e.g., reviews by Weertman and Weertman (1975) and Kirby (1977)). These studies suggest that the creep mechanism and the duration and magnitude of the creep can be a sensitive function of temperature, stress, and to a lesser extent pressure. There seems to be little doubt that flow of the asthenosphere is an important process in a variety of geologic processes including tectonic plate motions and the accumulation of strain at plate boundaries as part of the earthquake cycle. Therefore it is not surprising that asthenospheric flow has been suggested as a mechanism for postseismic deformation. There is also evidence to suggest that the conditions in the lower lithosphere may also be appropriate for some form of creept. Temperatures may well exceed 1000°C and stress may exceed 1 kilobar (Goetze and Evans, 1979);

conditions at which creep processes may be active.

With these thoughts in mind I have investigated, in a theoretical and numerical model, the postseismic deformations to be expected from visco-elastic relaxation of the lower lithosphere and asthenosphere following an earthquake. The models I have considered are three- and four-layer representations of the earth with a strike-slip fault in the uppermost layer. As will be shown below, the surface deformations derived from this model are, under appropriate conditions, appreciably larger than those computed by ignoring the lower lithosphere viscoelasticity. On the other hand if the effective rigidity of the lower lithosphere shows little time dependence, no effect of this layer on postseismic rebound is expected. Similarly the surface deformations are enhanced when the depth of slip extends to the vicinity of the flow region and they are suppressed when the slip zone is shallow compared to these viscoelastic regions.

The multilayer viscoelastic model is shown in Figure 1. A strike-slip fault is embedded in the upper elastic layer (upper lithosphere with rigidity, μ_1) from the surface to a depth, D. The elastic layer has thickness H_1. Below this layer, to a depth H_2, is a viscoelastic solid layer consisting of an elastic element (rigidity, μ_{2a}) in series with a parallel combination of elastic (rigidity, μ_{2b}) and viscous elements (viscosity, η_2). Below this lower lithosphere layer is a Maxwell body consisting of a series combination of elastic (rigidity, μ_3) and viscous elements(viscosity, η_3) extending to a depth H_3 and repesenting the asthenosphere. In some of our calculations we have included a fourth layer to a depth H_4 which consists of an elastic body and represents the high viscosity region of the mantle lying under the fluid upper asthenosphere. The presence or absence of this layer does not effect the conclusions of this paper. The rheological elements are meant to describe the deviatoric or shear properties of the earth, the dilatational or volumetric properties are assumed elastic. A description of the creep properties of these linear elements can be found in standard texts and are summarized in the review article by Cohen (1979). In particular, for a suddenly applied and maintained constant shear stress, the three-element solid has an initial rigidity, μ_a, undergoes transient creep with a time constant $\tau = \eta/\mu_b$, and has a long term rigidity $\frac{\mu_a \mu_b}{\mu_a + \mu_b}$. Similarly the Maxwell substance has an initial rigidity, μ, undergoes steady state creep and has no long term rigidity. Under conditions of constant strain this body relaxed an applied stress with a time constant $\tau=\eta/\mu$.

The results I will discuss in the following paragraphs are derived using a two-dimensional finite element scheme which employs a version of the computer program developed by Melosh and Raefsky (1980a) which I have modified to accomodate the aforementioned rheological model for the lower lithosphere. The split node techniques of the aforementioned authors (Melosh and Raefsky, 1980b) has been employed to specify the fault slip. The results are derived using an explicit time integration algorithm. The parameters used in most of the calculations are $\mu_1 = \mu_{2a} = \mu_{2b} = \mu_3 = \mu_4 = 5 \cdot 10^{11}$ dyne/cm^2, $\eta_2 = 1 \cdot 10^{21}$ poise, $\eta_3 = 5 \cdot 10^{19}$ poise, D = 20 km, H_1 = 40 km, H_2 = 75 km, H_3 = 400 km H_4 = 800 km, slip = 1m. Sample calculations using an approximate elastic half-space, two rheologically different viscoelastic half-spaces, and a two-layer model of an elastic layer over a viscoelastic half space were compared to analytic and previously published numerical solutions to check for accuracy in the numerical procedure. As expected the uniform elastic and viscoelastic half-space models showed only coseismic and not time-dependent deformations, while the two-layer model showed the expected relaxation due to asthenospheric flow.

Turning to numerical results, the postseismic displacements, $\Delta W(t) = W(t) - W(0)$, where W is the surface displacement along the fault strike direction and time t=0 is the time of the earthquake, is shown as a function of the distance from the fault, in Figure 2. At the time chosen for the figure $t = 5 \times 10^9$ sec $\sim$ 159 years nearly all the postseismic relaxation due to both lower lithosphere and asthenosphere creep has been completed, at least within 500 km of the fault. The figure reveals that the postseismic displacement may exceed several tens-of-centimeters for major earthquakes involving several meters of coseismic slip. There is also the suggestion that postseismic displacements extend over a much broader distance from the fault than do the coseismic ones. The point of maximum postseismic displacement (and zero postseismic shear strain, see below) is about 200 km from the fault, although it must be remembered that this two-dimensional model assumes essentially infinitely long faults. The postseismic shear strain $\Delta\varepsilon_{13} = \frac{\partial W}{\partial X}$ (engineering strain) is shown as a function of distance from the fault in Figure 3. It can be shown that the coseismic strain drop $\varepsilon_{13}(0)$ is reduced from one-half its peak value at a distance X = D where it is assumed that the slip is uniform and the fault ruptures the surface and extends to the depth D (= 20 km in the present case). By contrast the postseismic strain (an increase) initially decreases more slowly with distance from the fault and reaches one-half its peak value at $X \doteq 60$ km. This broader region of significant postseismic strain is of course a reflection of the deeper position of the source of the postseismic motion, in this case the creeping portions of the lithosphere and asthenosphere.

While the preceding figures have examined the cumulative surface deformation versus distance from the fault, it is also interesting to consider the time dependence of the deformation at a fixed location. Figure 4 shows the strain versus time at a position close to the fault (averaged over an element of the finite element grid at distances 1-10 km from the fault). Also shown for comparison is a similar result for the case where lower lithsophere visco-elasticity is ignored, i.e., computed with the elastic layer extending to the top of the asthenosphere (at a depth of 75 km). It is clear that the effect of the shallow zone of partial flow is to double the ultimate post-seismic strain. Moreover, the results suggest that appreciable postseismic deformaton can occur even if the asthenosphere is deep compared to the depth of the coseismic slip zone provided the region of lithsopheric viscoelasticity begins at a sufficiently shallow depth. Figure 4 also reveals that the duration of surface straining following an earthquake is longer in the multilayer model. This is, of course, a direct consequence of the assumption of a higher viscosity (and hence longer response time) of the lower lithosphere layer compared to that of the asthenosphere. The magnitude of the postseismic deformation due to lithospheric relaxation can be increased by reducing this layer's effective longer rigidity (i.e., decrease μ_b) and by bringing the layer close to the seismogenic zone. ($H_1 \to D$). A detailed analysis of the postseismic displacements, strains, stresses and the time dependence and variation with depth, distance from the fault, and model parameters will be published elsewhere (Cohen, 1981).

It would be desirable to compare the model predictions to geodetic data. Unfortunately, since high precision surveys have been available only in recent years and over limited networks (in geographical location and size) and since major strike-slip events are infrequent, albeit potentially important, there is virtually no data for making such a comparison. Thatcher (1975) has reported postseismic strains in the range 41.9 ± 9.1 to 72.1 ± 15.5 μstrain over a period of 24 years in the immediate vicinity of the San Andreas fault subsequent to the 1906 San Francisco earthquake. While it is not clear that this model is applicable for such a shallow event (D = 10 km) we find that with a slip of 3-6 meters and H_1 in the range 10-20 km postseismic strains of 7-36 μstrain are predicted with the time scale depending on the chosen viscosities. Neither the observations nor the model predictions have been adjusted for the possible effects of straining due to plate motion although this effect may be significant (i.e., with $\dot{\varepsilon} = 10^{-6}$/yr, $\Delta\varepsilon$ = 24 μstrain after 24 years).

In summary, a multilayer model of earth has been used to examine post-seismic deformation following a major strike-slip earthquake. The analysis has suggested the viscoelastic relaxation of the earth's subsurface layers including the lower lithosphere and asthenosphere may be responsible for significant deformations, reloading of the fault slip region, and broadening of the initial coseismic deformation region. These viscoelastic effects are of potentially greater significance for some major thrust earthquakes which occur at greater depths than strike-slip events and hence in regions where creep processes may be expected to have a higher degree of activation. I hope to report on calculations of the postseismic deformations for the thrust case at a later date.

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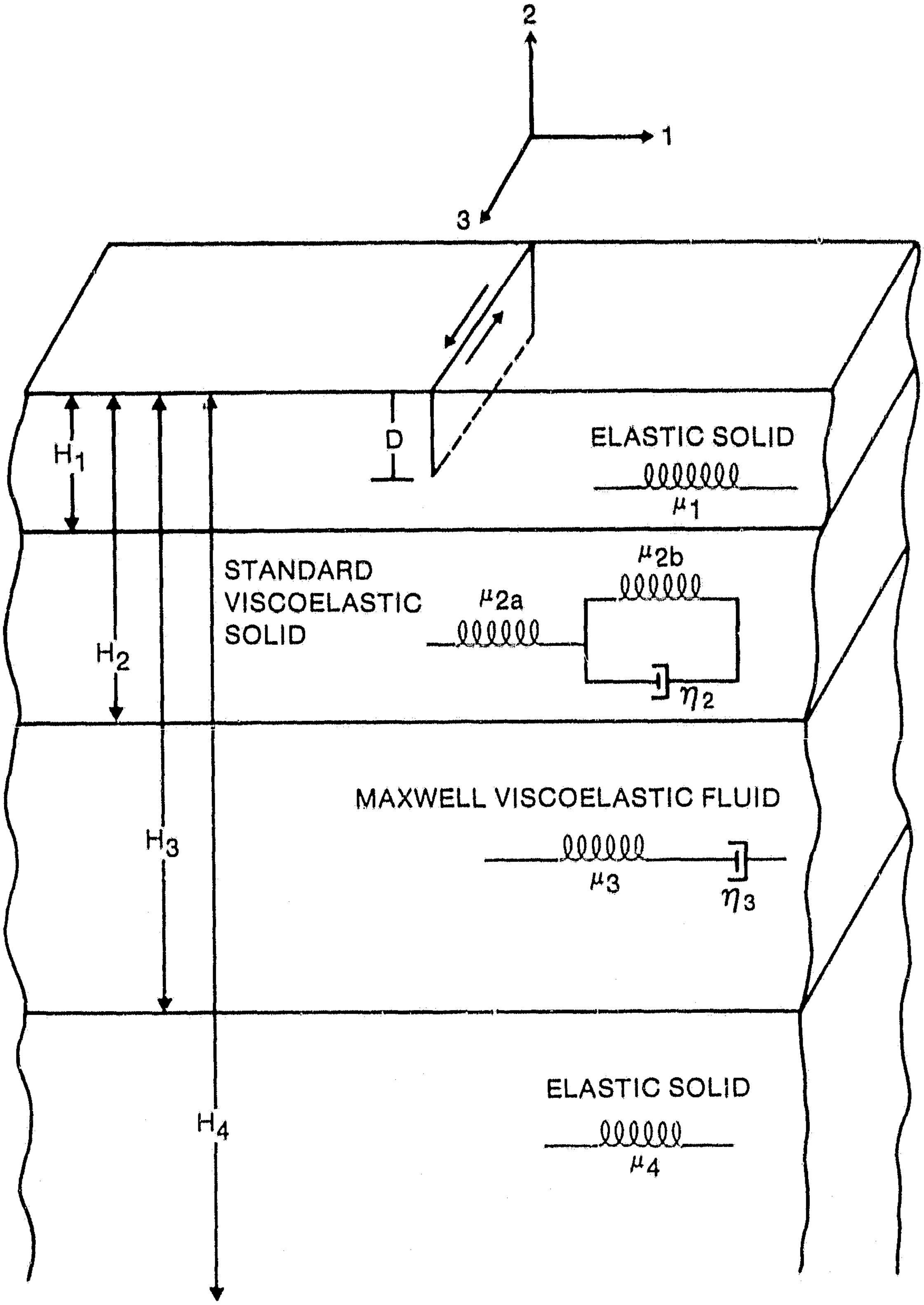

Figure 1. Rheological properties and parameters of multilayer postseimic deformation model.

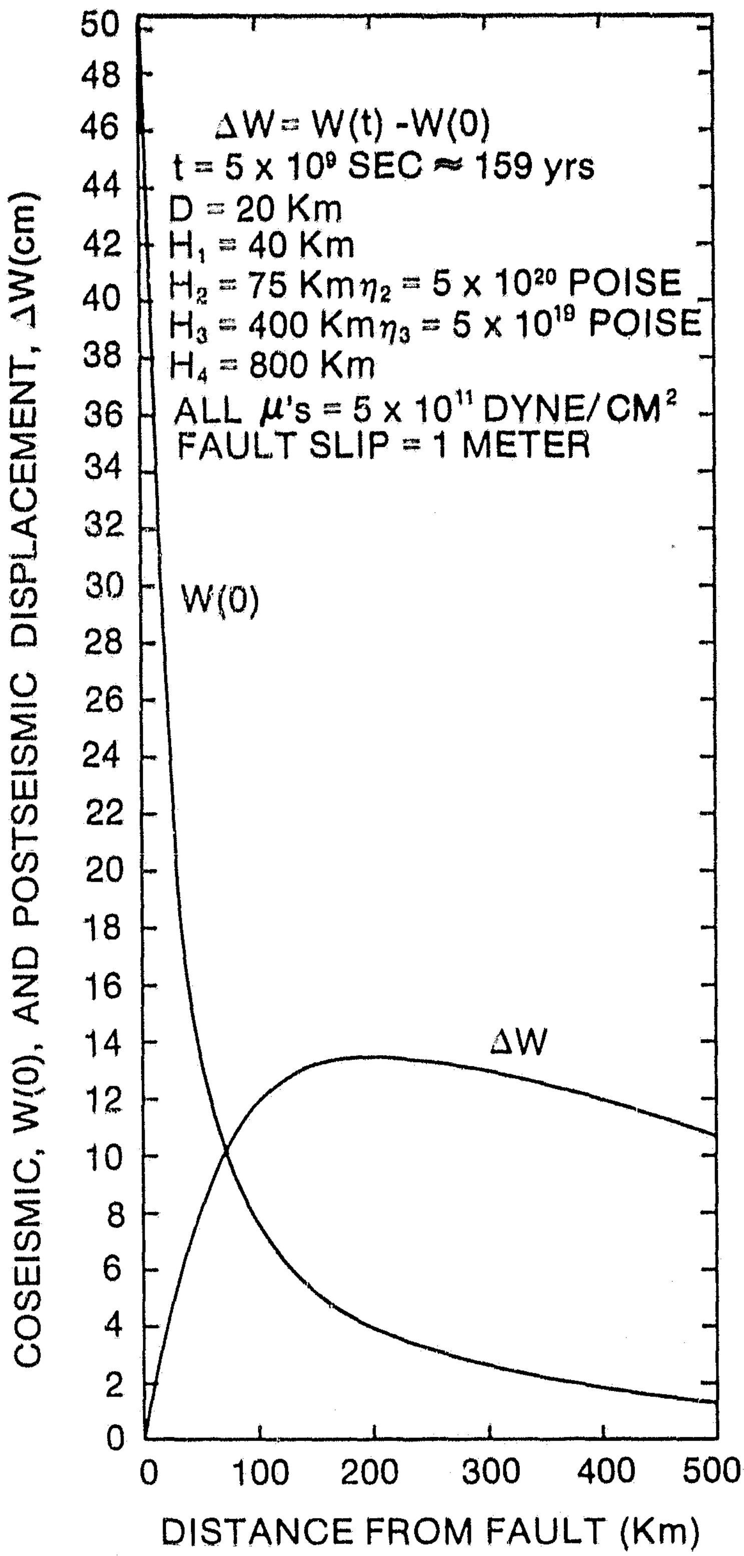

Figure 2. Computed surface displacements parallel to fault strike versus distance from fault. For postseismic displacements, ΔW, curve shows total displacements after $5 \cdot 10^9$ seconds ~ 159 yrs.

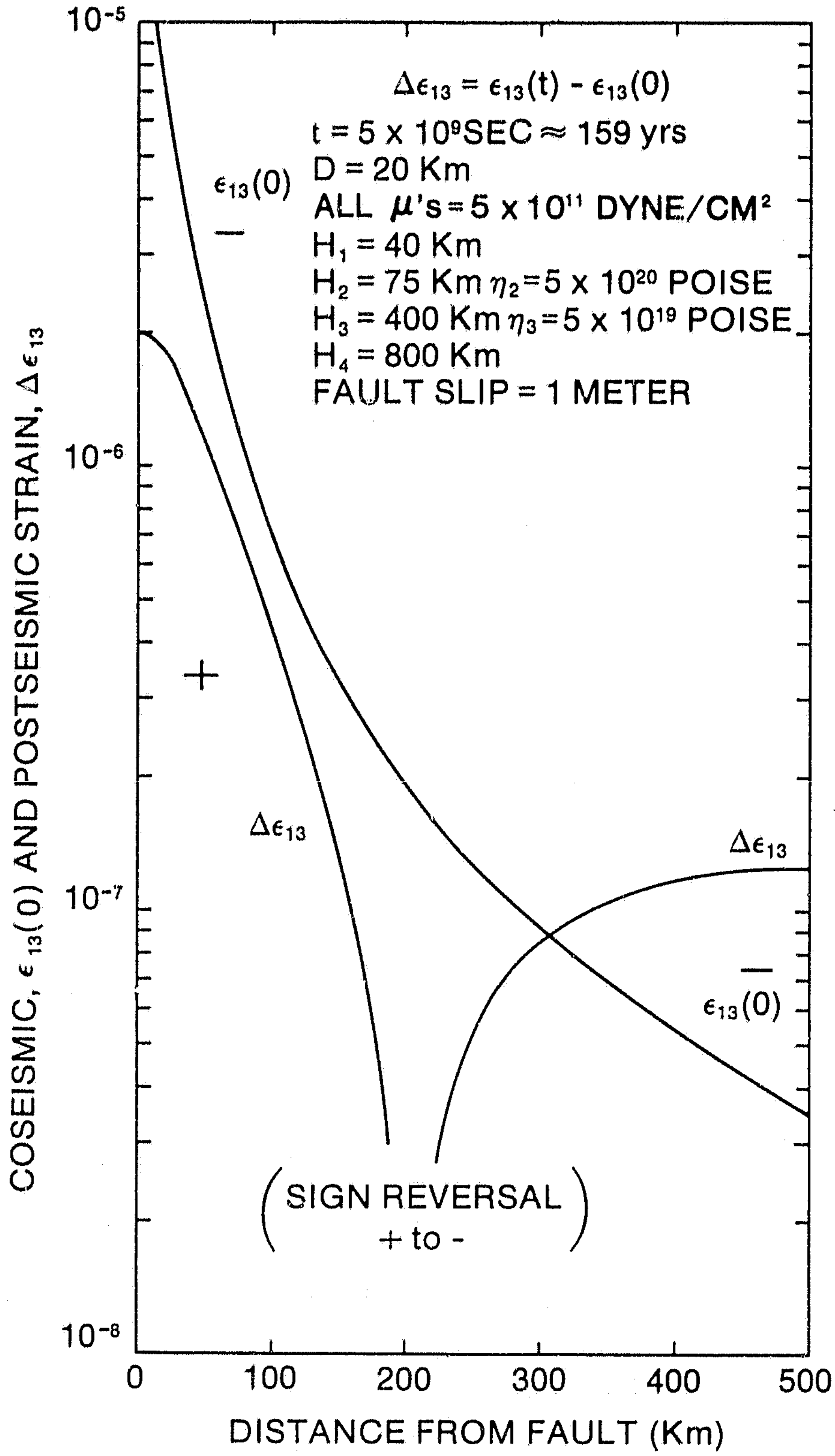

Figure 3. Computed strain in horizontal surface plane versus distances from fault. For postseismic displacements, $\Delta\epsilon_{13}$, curve shows total strain after $5 \cdot 10^9$ seconds ~ 159 yrs. Engineering strain $\Delta\epsilon_{13} = \frac{\partial W}{\partial X}$ is plotted.

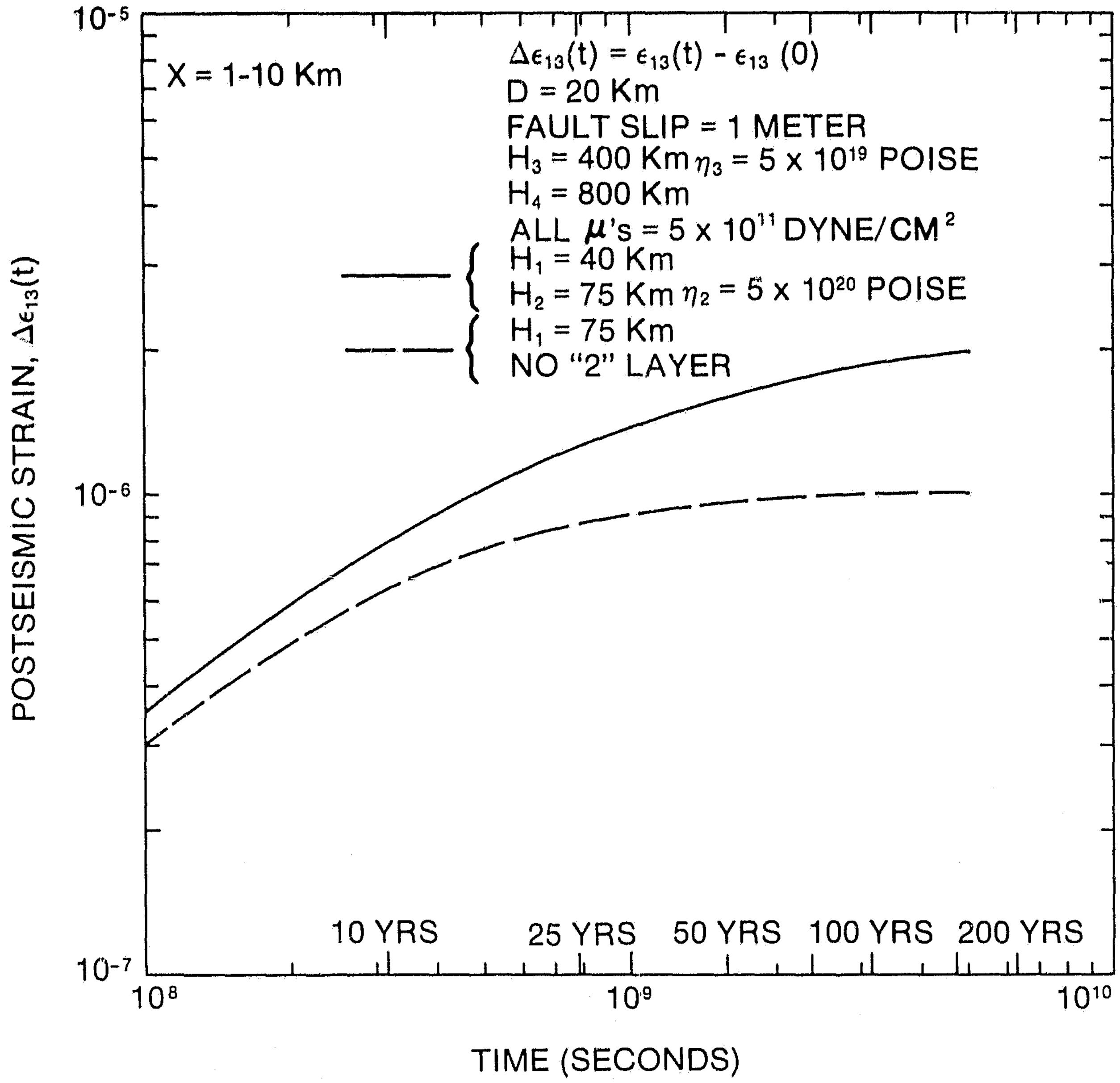

Figure 4. Computed postseismic strain versus time following earthquake. Dashed line shows corresponding result for elastic lithosphere–viscoelastic asthenosphere model with no creep in lower lithosphere.