

Correction to "Transient Creep of a Composite Lower Crust 2:
A Polymineralic Basis for Rapidly Evolving
Postseismic Deformation Modes"

by

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The paper "Transient Creep of a Composite Lower Crust 2: A Polyminerale Basis for Rapidly Evolving Postseismic Deformation Modes" by Erik R. Ivins (*Journal of Geophysical Research*, 101 (B12), 28,005-28,028, 1996) contains errors in the calculation of viscosity and Maxwell relaxation time in the section titled "Viscosity of quartzite at lower crustal conditions".

Ivins [1996] used three relations for estimating the effective viscosity of quartzites and crustal rocks having strong quartz affinity. The first corresponds to a Newtonian flow law (equation 1a, therein) for the deformation experiments reported by Wang *et al.* [1994]

$$\eta(Q, f_{\text{H}_2\text{O}}, T) = \frac{1}{2 A f_{\text{H}_2\text{O}}^p} \exp(Q/R T) \quad , \quad (1a)$$

and the second two (equations 1b and 1c, therein) for either constant strain-rate ($\langle \dot{\epsilon} \rangle$), or constant stress ($\langle \sigma \rangle$), forms of the well-known "effective" viscosities for non-Newtonian power law relations;

$$\eta_{\text{eff}}(Q, n, T, \langle \dot{\epsilon} \rangle) = \frac{\langle \dot{\epsilon} \rangle^{(\frac{1}{n}-1)}}{2 A^n} \exp(Q/n R T) \quad , \quad (1b)$$

$$\eta_{\text{eff}}(Q, n, T, \langle \sigma \rangle) = \frac{\langle \sigma \rangle^{1-n}}{2 A} \exp(Q/R T) \quad . \quad (1c)$$

The relations are repeated here for clarity. In (1a) $f_{\text{H}_2\text{O}}$ represents the partial pressure of the lattice impurity H_2O (fugacity) and p an experimentally determined exponent; A , an experimentally determined material constant of units $\text{MPa}^{-(p+1)}$; Q , the activation energy; R , the gas constant; and, T , the temperature in $^\circ\text{K}$. In (1b) and (1c) n in the power law exponent for dislocation creep and A has units MPa^{-n} . Rocks that contain a substantial proportion of quartzite may be highly ductile. Figures 2 - 4 of Ivins [1996] provided quantitative estimates of the effective rheological softening as a function of increasing temperature using laboratory-derived constants (see relations 1a - c) for both quartzites and rock aggregates that contain quartz as a significant constituent. Figure 2 showed the effective non-Newtonian viscosity as a function temperature

ranging from 450 to 750 °C for experimental results obtained by *Paterson and Luan* [1990] using two differently prepared synthetic quartzites. Figure 3 showed the corresponding Maxwell relaxation time constant τ_{MX} that correspond to the same two rock types (Figure 2) over the range 450 to 500 °C. Figure 4 showed the half-space time constants ($2 \times \tau_{MX}$) for four polyphase crustal rocks (from 500 to 800 °C) with comparison to the postseismic exponential time constants that were determined geodetically following the Landers earthquake of June 1992 [*Wyatt et al.*, 1994; *Shen et al.*, 1994]. All three of these figures contained errors.

Figure 4a contains some errors of minor consequence (granite and diorite curves should be transposed, the actual stress values in the calculation are 0.5, 2.5 and 7.5 MPa, respectively, and the diorite flow parameter in the caption was misreported and should be corrected to $A = 8.7 \times 10^{-3} \text{ MPa}^{-n}$). However, a more substantive correction must be applied to Figures 2, 3 and 4 since all three suffer from a blunder in the calculation of viscosity as these lack an appropriate conversion from MPa to Pa. The error is of some consequence. It implies that for significant rheological softening to be operative for all non-Newtonian flow laws under discussion in *Ivins* [1996] the relevant temperature must be elevated to the range of 700 to 950 °C, or more. (All of the results reported therein may be recovered by making the appropriate correction for error in units conversion.) The figures in question, upon correction and shift to the elevated temperature range, retain their primary purpose in the paper. That purpose, quite simply, was to make a quantitative connection between the mechanical strength parameters of crustal materials as measured in laboratory experiments and the weak viscoelastic inclusions that are employed in the analytical model for surface strain and crustal stress evolution in the remainder of the paper. The errors do not directly affect any other section of the paper.

All of the corrected Figures, with some additional captioned information and the elevated temperature ranges on the abscissae, are given in Figures 2, 3 and 4 here. In addition, Figures 2 and 4 are each supplied with new frames (c) wherein the viscosity and Maxwell half-space relaxation times for Heavitree quartzites are examined in greater detail. Figure 4c additionally shows the relaxation time constants that can be estimated using recently published strength relations for

synthetic anorthite feldspar [Dimanov *et al.*, 1998] and partially molten plagioclase [Dimanov *et al.*, 1999]. It should be noted that the Newtonian rheology of Wang *et al.* [1994] is also given in Figure 3. In addition, some of the text in the relevant section needs correction. Following the discussion of the Newtonian flow law (equation 1a) an example calculation was given of viscosity and Maxwell time for Heavitree quartzite [Wang *et al.*, 1994]. Previously, I reported a proxy value for quartzite viscosity at midcrustal temperatures incorrectly and I offer the following corrected example calculation at $T = 700$ °C, such that $\eta = 3.53 \times 10^{14}$ Pa s and a prediction of an intrinsic Maxwell time (η/μ^c) of about 6 hours and 35 minutes with $\mu^c = 14.5$ GPa. For this latter calculation I assume the constants determined by Wang *et al.* [1994] ($A = 1.57 \times 10^{-3}$ MPa $^{-(p+1)}$ s $^{-1}$, $Q = 131.5$ kJ/mol, $p = 0.41$) and a water fugacity of $f_{\text{H}_2\text{O}} = 300$ MPa. The corrected non-Newtonian viscosities and relaxation times now shown in Figure 2 are consistent with a recent paper by Freed and Lin [1998]. Here, however, I also consider values of the strain-rate that are higher than known to be typical of the crustal surface during the interseismic period ($\approx 1/2 \times 10^{-14}$ s $^{-1}$). Note that the value assumed for the constant μ^c in all calculations presented here tend to maximize τ_{MX} . For example the elastic shear modulus of the Moho-asthenosphere environment is, roughly, 27 - 75 MPa (e.g., PREM at 1 sec period). Consequently, I may be underestimating the relaxation times here by as much as a factor of 5.

It may be of importance to include some additional remarks concerning the elevated temperatures that, indeed, seem required in order to establish a causal link between the experimental flow laws derived for deep-seated crustal rocks and postseismic decay times that are of the order of tens of days, or less. Results of deformation experiments with partially molten synthetic plagioclase [Dimanov *et al.*, 1998] and synthetic anorthite feldspar [Dimanov *et al.*, 1999] may be described using the following Newtonian relation:

$$\eta = \frac{d^{-m}}{2A} \exp(Q/RT) \quad (2)$$

The micromechanical rate-controlling mechanism for each of these experiments involved grain boundary diffusion creep. It is instructive to consider the parameters for relation (2) found by

Dimanov et al. [1998; 1999], as in Figure 4c, for partially molten plagioclase (< 1% to 10% melt) and 'wet' ($\approx 8000 H/10^6 Si$, or 0.05 wt.%) anorthite feldspar ($A = 1.09 \times 10^{-15}$, $4.09 \times 10^{-13} \text{meter}^{-m} \text{Pa}^{-1} \text{s}^{-1}$; $Q = 338, 377 \text{ kJ/mol}$; $m = -2.7, -2.5$, for the two materials, respectively). Employing the same value of μ^c as in Figure 3 and 4, at the relatively high temperature $T = 945$ °C and with a grain size parameter $d = 10 \mu\text{m}$ (possibly appropriate to ultramylonites within shear zones of the lower continental crust), the computed half-space relaxation times are: $2 \times \tau_{\text{MX}} = 7.25$ and 9.23 days, respectively. In summary, the corrected and amended Figures 2 - 4 show (with the exception of the parameters for equation (1a) reported by *Wang et al.* [1994]) that the experimental flow laws for crustal rocks require higher crustal temperatures, near those of amphibolite facies, for consistency with the postseismic Landers geodetic data (see Figure 4 a and b). The required temperatures are higher by about 250 °C than indicated in the flawed figures 2 - 4 published by *Ivins* [1996].

Acknowledgments

I wish to thank Elizabeth Hearn who first informed me of errors in the calculation of viscosity and Maxwell time in Figure 4 and elsewhere in *Ivins* [1996]. This research was funded by the Solid Earth and Natural Hazards program under the auspices of NASA. The work reported was performed by the Jet Propulsion Laboratory, California Institute of Technology under contract NAS7-100.

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Figure Captions

Figure 2. Effective viscosities for synthetic quartzite based upon laboratory-derived flow laws. The 'gel' (a) and 'silicic acid' (b) samples are described by *Paterson and Luan* [1990] and *Luan and Paterson* [1992]. For 'gel' $A = 6.5 \times 10^{-8} \text{ MPa}^{-n} \text{ s}^{-1}$ and for 'silicic acid' $A = 4.0 \times 10^{-10} \text{ MPa}^{-n} \text{ s}^{-1}$. The stiffer rheology of the silicic acid samples are, in part, explained by a smaller hydroxyl component and are, thus, more consistent with an 'anhydrous' mineral than is the gel sample. The hydroxyl component (measured as the ratio of water-related species to silicon content defined by the ratio of H to Si). The ratios for gel and silicic acid specimens are $12,600 H / 10^6 Si$ and $2500 H / 10^6 Si$, respectively. Generally, quartz is referred to as 'wet' at about a ratio of $700 H / 10^6 Si$, or larger, and 'dry' for ratios of about $100 H / 10^6 Si$ or lower [*Gerretsen et al.*, 1993]. Also shown in this revised figure is frame (c) for a natural rock quartzite (Heavitree). with water added to the mineral and grain boundary structure. A non-Newtonian dislocation mechanism controlled the deformation of the Heavitree quartzite sample studied by *Jaoul et al.*, [1984]. (For the latter study $A = 5.26 \times 10^{-3} \text{ MPa}^{-n} \text{ s}^{-1}$). The three consecutive unlabeled strain rate ($\langle \dot{\epsilon} \rangle$) and stress ($\langle \sigma \rangle$) values have the same values as in frames (a) and (b).

Figure 3. Maxwell times for both the effective non-Newtonian ($n = 1.4$) and Newtonian ($n = 1$) Heavitree quartz viscosities shown in Figure 2c. Two different partial pressure values (f_{H_2O}) are shown for the law obeying equation (1a). Note that in this corrected figure the "gel" and "silicic acid" results shown in Figure 2 do not plot in this space. Note the quite large differences between the predicted Maxwell time constants for dislocation vs. Harper-Dorn (grain boundary dislocation motions) of Heavitree quartzite. The Harper-Dorn mechanism was the rate-controlling factor in the experiments of *Wang et al.* [1994]. The elastic rigidity is assumed to be $\mu^c = 14.5 \text{ GPa}$ and other parameters for the experiments of *Jaoul et*

al. [1984] are given in the caption to Figure 2. Other parameters assumed for the experiments of *Wang et al.* [1994] (p and A) are identical to those used in the discussion given in the text.

Figure 4. Relaxation times ($2 \times \tau_{MX}$) from steady-state flow laws for polymineralic rocks. In frame (a) granite (heavy line-type) is assumed to be composed of 68% plagioclase, 5% biotite and 27% quartz. Composition of diorite in frame (a) (light line-type) assumes 66% plagioclase, 28% clinopyroxene and 6% quartz. All calculations assume constant stress (as indicated) and μ^c as in Figure 3. The granite and diorite flow parameters are given explicitly in Table 1 of *Ji and Zhao [1993]* wherein: $A = 3.65 \times 10^{-7}$, $8.7 \times 10^{-3} \text{ MPa}^{-n} \cdot \text{s}^{-1}$; $n = 3.9$, 2.9 ; $Q = 206$, 261 kJ/mol , for the polymineralic granite and diorite rock models, respectively. In frame (b) no averaging method applies, but a direct experimental determination is assumed. For Maryland diabase (solid line-type) and Andirondack granulite (light line-type) $A = 6.12 \times 10^{-2}$, $8.0 \times 10^{-3} \text{ MPa}^{-n} \cdot \text{s}^{-1}$; $n = 3.05$, 3.1 ; $Q = 276$, 243 kJ/mol , respectively. (See results reported by *Caristan [1982]* and *Wilks and Carter [1990]*, respectively). Frame (c) shows the Maxwell half-space constants predicted using the flow laws discussed by *Patterson and Luan [1990]* (also see Figure 2) and those using equation (2) for experiments of *Dimanov et al.*, [1998; 1999] for plagioclase and 'wet' feldspar, respectively (see text for values assumed for the parameters of the grain boundary diffusion creep relation).

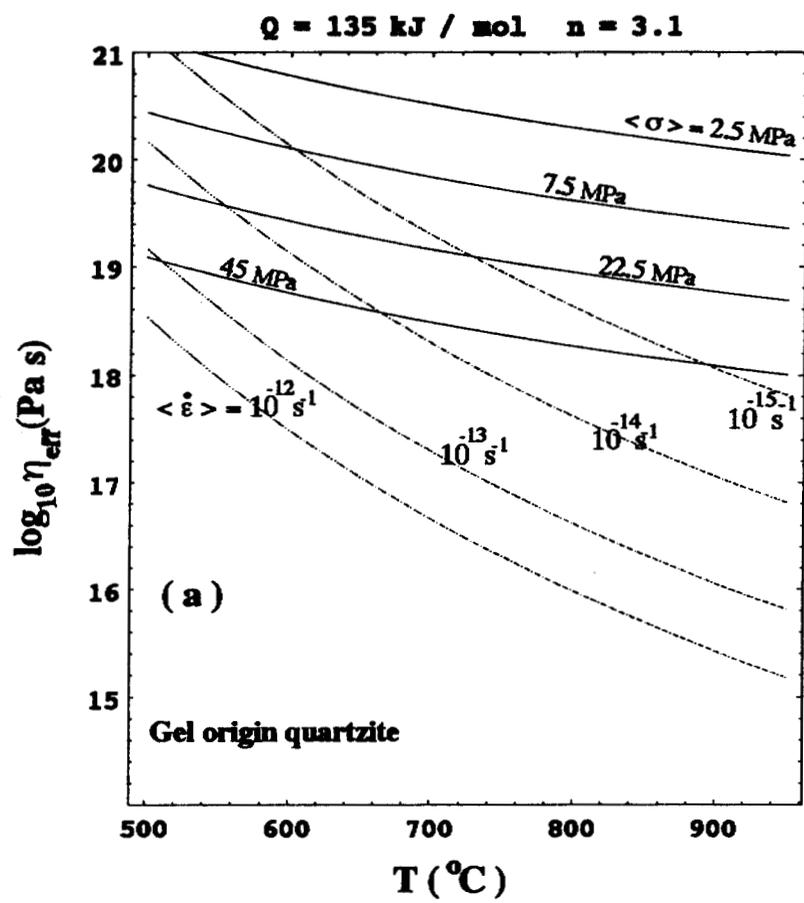


Figure 2a

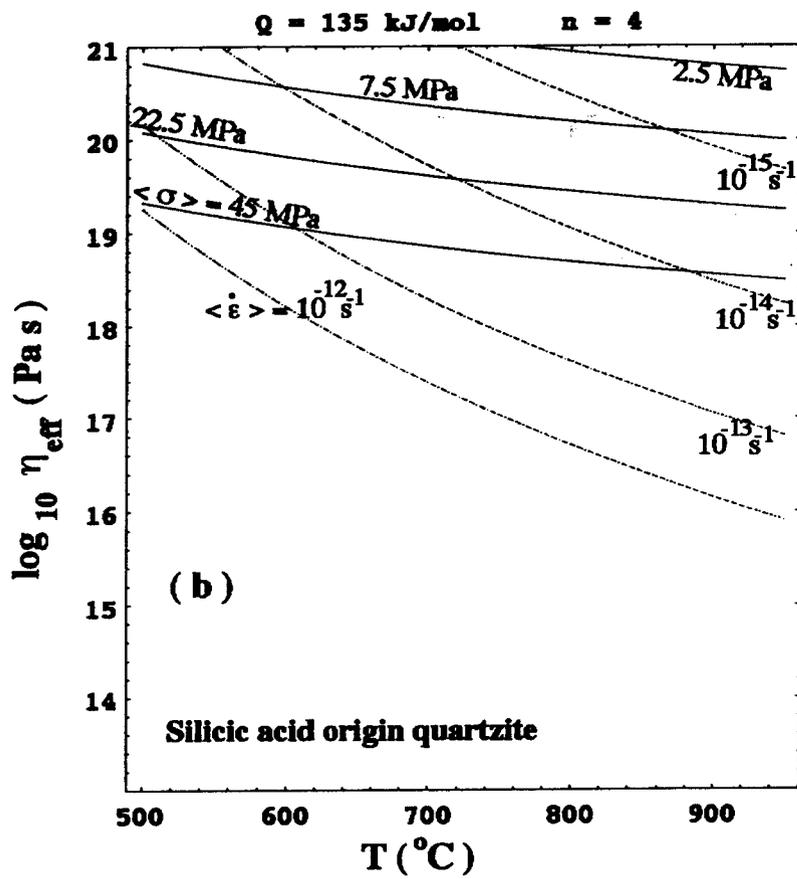


Figure 2b

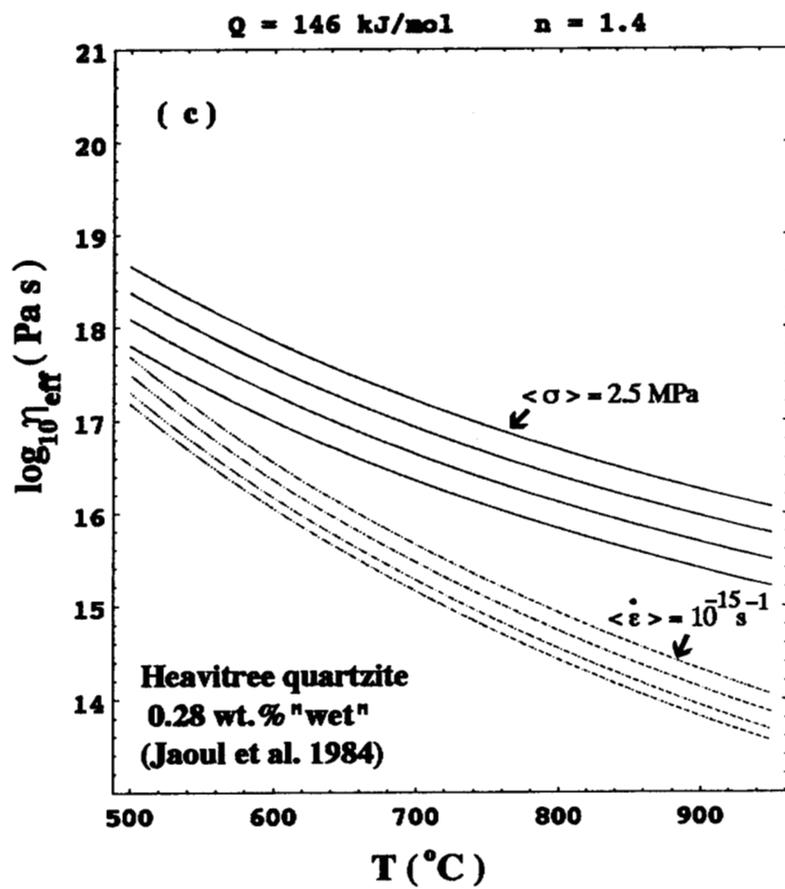


Figure 2c

Harper-Dorn and 0.28 wt.% wet Heavitree quartzite

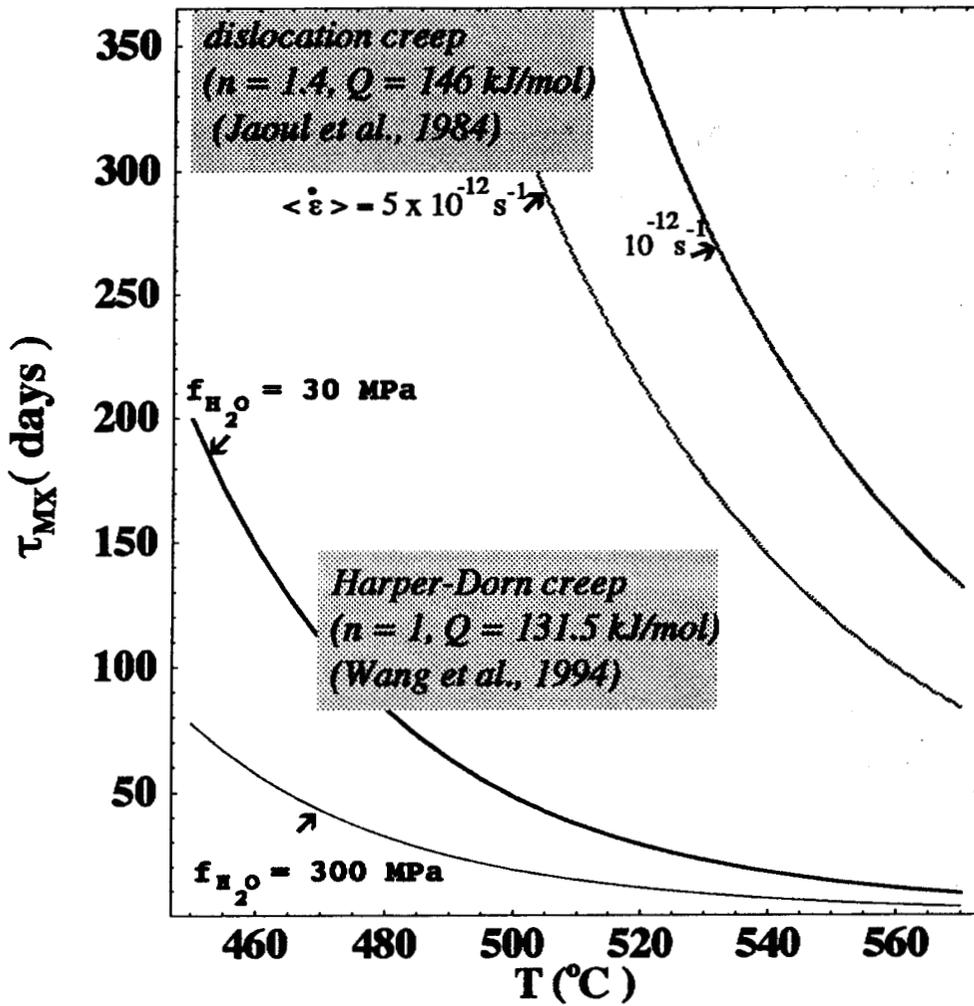


Figure 3

half-space Maxwell times computed from constant stress
polyphase granite & diorite strength laws

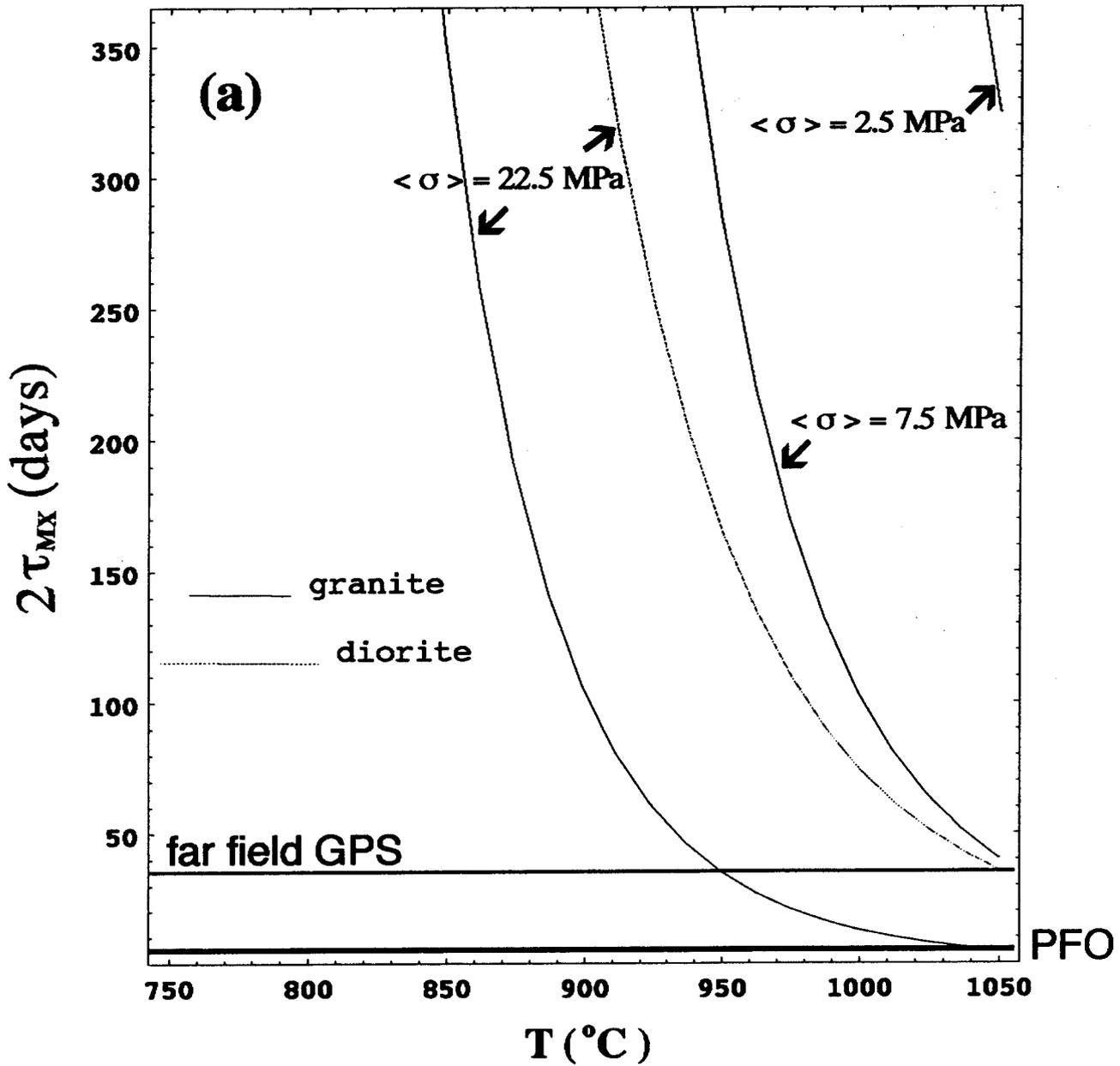


Figure 4a

Maryland diabase & Adirondack granulite

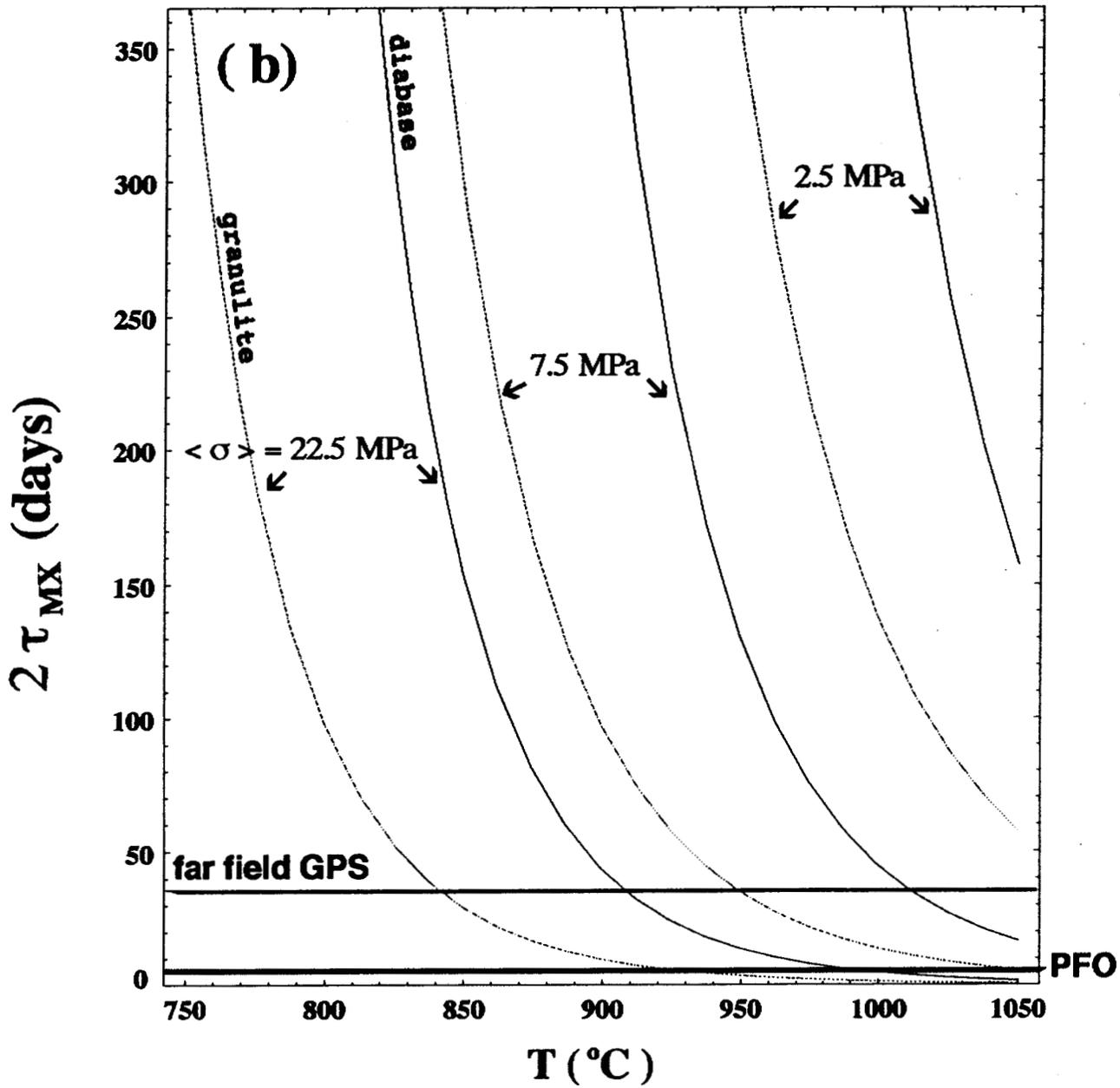


Figure 4b

plagioclase – fine grained feldspar – synthetic quartzites

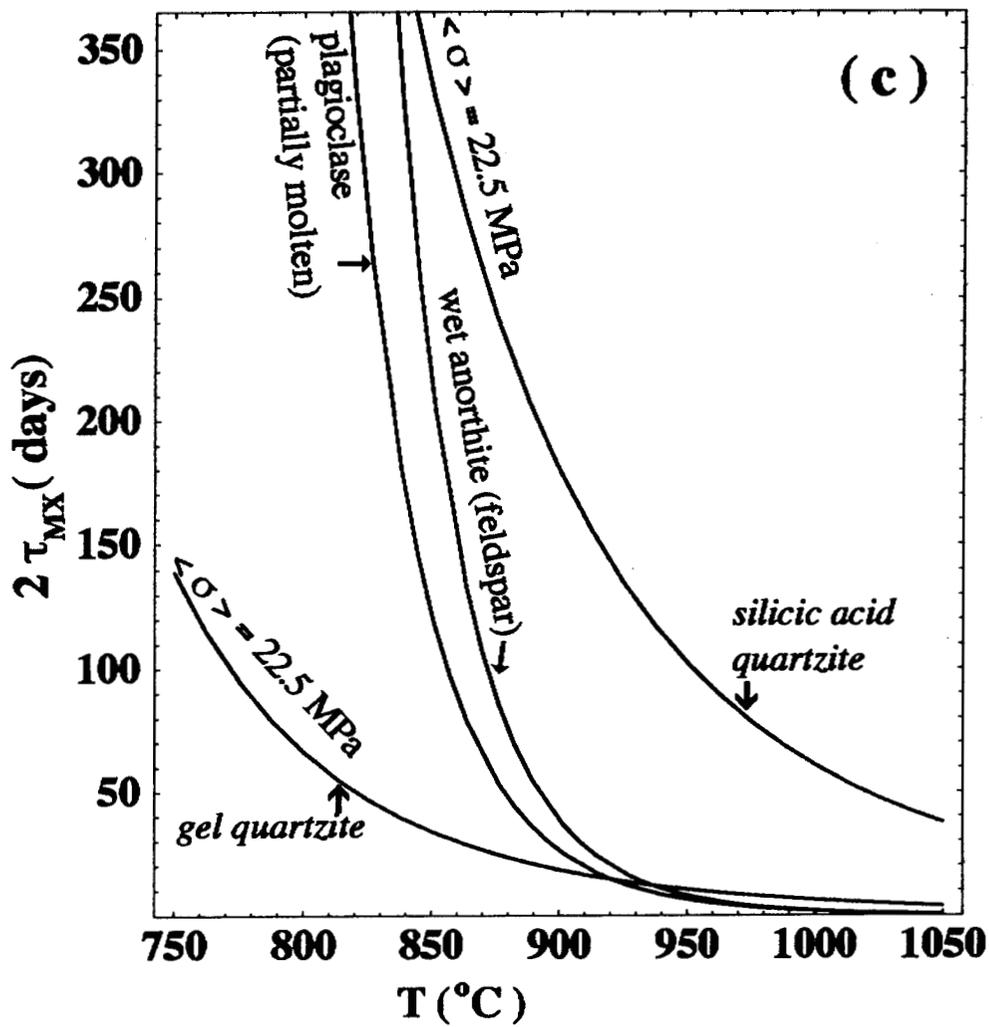


Figure 4c