resolvable Raman bands in this spectral region at 653, 600, 530, and 330 cm^{-1}. The weak bands at 653 and 600 cm^{-1} are similar in frequency to strong bands at 648 and 608 cm^{-1} in crystalline $Na_4Ge_9O_{20}$ and have been assigned to symmetric stretching of Ge-O-Ge linkages between octahedra. The strong band at 530 cm^{-1} has been generally assigned to symmetric stretching or deformation of Ge-O-Ge linkages between germanium tetrahedra. The weak band at 330 cm^{-1} compares favorably in frequency with a strong band at 309 cm^{-1} in crystalline $Na_4Ge_9O_{20}$, and the intensity of a feature near 300 cm^{-1} in glasses in the K_2O-GeO_2 system is clearly correlated with the intensities of bands at 645 and 603 cm^{-1} (Verweij and Buster, cited in this note).

20. The effect of the added H_2O in our samples is easily discerned. In the zero-pressure spectrum of anhydrous $Na_2Ge_2O_5$ glass, there are strong bands at 867 and 790 cm^{-1}. Both of these bands are associated with vibrations of relatively polymerized units of tetrahedrally coordinated germanium [Verweij and Buster; Walrafen; Furukawa and White; all in (*18*)]. In our lowest pressure spectra of the hydrated liquid, these bands are reduced in amplitude to the degree that they are unresolvable. Thus, added water produces depolymerization of the liquid relative to the anhydrous glass, in accord with the solution mechanisms of H_2O in silicate liquids [E. M. Stolper, *Contrib. Mineral. Petrol.* **81**, 1 (1982)].
21. We utilized a range of abundances of Ge^{VI} {between 3 and 8% [Ueno *et al.* and Fukushima *et al.*, respectively, in (*18*)]} and the areas under the ~500 and ~300 cm^{-1} bands in the zero-pressure Raman spectrum of anhydrous $Na_2Ge_2O_5$ glass to calculate relative scattering cross sections of the two species in the glass. Estimated relative Raman scattering cross sections of octahedral to tetrahedral species range from 0.39 to 2.6. Our data are plotted with an average value of 1.5, but the stippled region of Fig. 2 incorporates the entire range of values.
22. Other possible mechanisms for densification include both simple bond compaction and more complex changes in polyhedral packing involving changes in second or more distant neighbor distances. The sequence with increasing pressure of the liquidus phases of ultrabasic melts of olivine, pyrope-majorite solid solutions (up to 25% Si^{VI}), and $MgSiO_3$ perovskite (100% Si^{VI}) suggests that the dominant structural change over the pressure range of the transition zone is the conversion of fourfold silicon to sixfold silicon [E. Takahashi, *J. Geophys. Res.* **91**, 9367 (1986); K. Wei, R. G. Tronnes, C. M. Scarfe, *ibid.* **95**, 15817 (1990)].
23. The shoulder present at ~425 cm^{-1} in the spectrum at 1.4 GPa grew with increasing pressure to become the dominant component in this spectral region (400 to 600 cm^{-1}) at the highest pressures of this study (Fig. 1). This shift in frequency suggests a change in the local tetrahedral environment upon compression. We believe that the decrease in the frequency of this vibration is produced by the tetrahedra in the high-pressure liquid sharing oxygens with a greater proportion of relatively weakly bound GeO_6 octahedra. Such a weakening of the tetrahedral environment would be expected to lower the vibrational frequency of this band, as is observed. However, we cannot preclude that the peak at 425 cm^{-1} is not associated with more highly coordinated germanate units. If this is the case, then our estimates of the amount of GeO_6 octahedra within the liquid are conservative.
24. The amplitude and the FWHM of the ~500 cm^{-1} peak decreased with increasing pressure, but neither peak broadens by a factor of ~3 to 4 with compression, as is seen in spectra of compressed germanate glasses (*7*). Thus, the local environment of the germanium may be less distorted in the liquid than in the compressed glass.
25. The zero-pressure bulk moduli of $Na_2Ge_2O_5$ glass and $Na_2Si_2O_5$ glass are 33 and 40 GPa, respectively [R. R. Shaw, *J. Am. Ceram. Soc.* **54**, 170 (1970); G. K. White, J. A. Birch, M. H. Manghnani, *J. Non-Cryst. Solids* **23**, 99 (1977)]. This similarity indicates that the elasticity of $Na_2Si_2O_5$ liquid may be used to conservatively estimate the compression of $Na_2Ge_2O_5$ liquid. The bulk modulus of metastable $Na_2Si_2O_5$ liquid at 940 K is 20 GPa [V. C. Kress, Q. Williams, I. S. E. Carmichael, *Geochim. Cosmochim. Acta* **52**, 283 (1988)]: over the pressure range of our experiments, this would produce a volume compression of ~8.5% (note that the presence of water will probably decrease the bulk modulus of the liquid, again rendering this estimate conservative). The volume compression in silicate liquids in a pressure range over which they are inferred to undergo coordination changes (8 to 25 GPa) is comparable with the compression occurring in this germanate melt under lower pressure conditions (*10, 11*).
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27. Our observations of coexisting and interconverting Ge^{IV} and Ge^{VI} preclude some models for high-pressure coordination changes in liquids. Models that involve uniform distortion and deformation of tetrahedra until octahedral coordination is generated are inconsistent with our observation of extensive coexistence of two distinct coordination states {for example, Stolper and Ahrens [in (*26*)]}. However, our results are consistent with structural models of compressed alkali halide liquids that rely on coexisting coordination states [J. L. Tallon, *Phys. Lett.* **72**, 150 (1979)]. Furthermore, on the basis of our observations of a mixed-coordination liquid, we concur with Rustad *et al.* (*4*) that molecular dynamic simulations utilizing potentials in which the coordination number of silicon is fixed at four are unlikely to produce geophysically relevant results [for example, (*14*)].
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Small Earthquakes, Tectonic Forces

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Earthquake scaling and frequency-of-occurrence relations require that small earthquakes be just as important as larger ones in redistributing the forces that drive relative displacements across active faults of any dimension, including plate boundaries.

According to the Gutenberg-Richter frequency-of-occurrence relation (*1*), small earthquakes (*2*) occur in vastly greater numbers than large ones:

$$\log N = a - bM \quad (1a)$$

In Eq. 1a, N is the cumulative number of earthquakes that exceed or are equal to a given magnitude M in a chosen area for a chosen time interval. The constant a depends on the overall seismicity rate, which can vary greatly from one region to the next. The constant b, however, is almost always $\simeq 1$, although recent studies have suggested that $b < 1$ at very small magnitudes ($M \lesssim 2$) (*3, 4*) and have shown that $b \simeq 1.5$ at large magnitudes (*5*). The distribution function corresponding to Eq. 1a is of the same form: the number of earthquakes ΔN within a magnitude increment ΔM centered on M is

$$\log \Delta N\left(M \pm \frac{\Delta M}{2}\right) = a' - bM \quad (1b)$$

where the constant a' now depends on a, ΔM, and b (if different from 1).

Because of their greater numbers, microearthquakes have played an important role in illuminating faintly seismogenic structures (*6*). Otherwise, the physical significance of small earthquakes has been debated for decades, at least since the time of the first energy-magnitude studies (*7*) of the 1940s and 1950s, which showed that small earthquakes contributed negligibly to seismogenic energy budgets, either globally or regionally. By the early 1970s, seismologists had also learned that small earthquakes contribute neither to seismic moment sums nor to long-term displacement rates along active faults (*8*).

Numerous studies of the past two decades, however, have revealed that earthquake stress drops, in the midst of considerable scatter, are sensibly constant for small earthquakes and large (*9*), for interplate earthquakes and intraplate (*10*), and for shallow earthquakes and deep (*11*). For equidimensional sources of circular radius r, the stress-drop $\Delta\sigma$ relation is

$$\Delta\sigma = (7/16)M_0/r^3 \quad (2a)$$

whereas for large earthquakes with fault length l much greater than fault width w,

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the stress-drop relation is

$$\Delta\sigma = CM_0/lw^2 \quad (2b)$$

where M_0 is seismic moment in both cases and C is a constant that depends on whether the earthquake is strike-slip or dip-slip but in either case is near 1.5 (*12*). The moment-magnitude relation (*13*) is

$$\log M_0 = 1.5M + 16.0 \quad (3)$$

where M may be M_L (local magnitude), M_s (surface-wave magnitude), M_w (energy magnitude), or, by definition, **M** (moment magnitude).

Equations 1a, 1b, 2a, and 3 may be combined to obtain for $b = 1$ (small earthquakes)

$$N,\ \Delta N \sim r^{-2} \quad (4a)$$

whereas Eqs. 1a, 1b, 2b, and 3 yield for $b = 1.5$ (large earthquakes)

$$N,\ \Delta N \sim (lw)^{-1} \quad (4b)$$

In either case, Eqs. 4a and 4b say that both small and large earthquakes occur with a frequency that scales as reciprocal faulting area (*14*). Small earthquakes "cover" the fault plane, then, to the same extent as the larger earthquakes. The larger earthquakes, however, account for virtually all of the strain energy release and virtually all of the seismogenic slip on faults. Matters are different with respect to the redistribution of forces that drive seismogenesis.

Small earthquakes redistribute the forces that exist along active faults, including plate boundaries; only the largest earthquakes, presumably, can interact with distant loading conditions (*15*) to reduce, temporarily, the net force on these faults. The force redistribution ΔF induced by all ΔN earthquakes of dimension r or area $A = lw$ is

$$\Delta F \sim \Delta\sigma(r^2,A)\Delta N \quad (5)$$

For constant stress drop and ΔN given by Eqs. 4a and 4b, ΔF is constant. Thus, smaller earthquakes are just as important as larger ones in redistributing the driving forces along active faults; they just do their business on shorter wavelengths, making up for their smaller sizes with their greater numbers.

The Earth is visibly heterogeneous over 11 orders of magnitude in dimension, from the scale of continents and oceans to individual grains in rocks. Seismogenic (brittle) failure, as known from laboratory studies (*16*), investigations of mining-induced seismicity (*17*), and results from naturally occurring earthquakes (*9–11*), occurs through most of this range, although the rupture zones of even the greatest earthquakes fall short of continental dimensions. That Eqs. 1 through 4 say that seismogenic failure is a self-similar, scale-invariant process with a natural area-scaling for frequency-of-occurrence would seem to be the straightforward consequence of stochastically well-behaved heterogeneity within the Earth and of the brittle failure process (*18*). At first blush, so would Eq. 5, the simplicity of the result in keeping with its derivation from simple dimensional analysis. Hindsight of this sort, however, is probably closer to 20-02 than to 20-20. Had the results of Eqs. 4a and 4b and Eq. 5 been anticipated 20 years ago [and those of Eqs. 4a and 4b had been anticipated (*9*)], when work on the magnitude dependence of seismic moment and stress drop began in earnest, we could have immediately inferred that earthquake stress drops would be constant and, given that Eqs. 1a and 1b were known at the time, that the moment-magnitude relation coefficient would be 1.5, findings that required hundreds of studies and scientist-millennia to establish empirically.

The nondispersive character of Eq. 5 means that a collection of small earthquakes, however big, will not in general provide a load distribution favorable to a large earthquake, even though such a sequence can occur by chance alone. Conversely, the occurrence of a large earthquake is not necessary to "roughen" a fault in order to provide for small earthquakes, even though this is, to all appearances, a common occurrence in the case of crustal aftershock sequences. Rather, the abundance of aftershocks and paucity of foreshocks are more likely attributed to rapid load redistribution in the first instance and very slow load accumulation in the second. This is to say that earthquake occurrence is a strain-rate–dependent process, given that driving stresses are near the failure stress. This strain-rate dependence, when properly formulated, will materialize only in the a values of Eq. 1a, not in the b value, as observational evidence alone suggested more than a decade ago (*19*).

Because the forces that act on crustal faults must be redistributed evenly on all scales according to Eq. 5, reports that Eq. 1a (*3, 4*) and the constancy of earthquake stress drops (*3*) fail at $M \lesssim 2$ means that active faults are less brittle at small scale than at large scale. This implication defies intuition, but in any event there is abundant evidence for brittle failure on dimensions much smaller than $M \simeq 2$ earthquakes (*16, 17*), for which $r \simeq 50$ m, and other explanations for these effects exist (*20*). Because the loading of plate boundaries and other active crustal faults is apparently smooth on scales of hundreds, and perhaps thousands, of kilometers (*21*), the occurrence of small earthquakes within the framework of Eqs. 1 through 5 can be taken as certain evidence of the potential for larger earthquakes, as large as permitted by coherent segment lengths of the seismogenic structure (*22*). Because the framework here is that of earthquake occurrence on throughgoing, essentially planar faults, the nature of volumetrically distributed seismicity, such as occurs in the eastern United States, requires further thought. With their California cousins, these earthquakes share b values near 1 (*23*), constant earthquake stress drops (*10*), and the same moment-magnitude relation.

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1. B. Gutenberg and C. F. Richter, *Seismicity of the Earth and Associated Phenomena* (Princeton Univ. Press, Princeton, NJ, 1949).
2. The phrase "small earthquakes" is used here to mean earthquakes with source radius less than one-half of the fault width; the fault width is typically 15 to 20 km in continental crust but can be 50 to 100 km in the case of major subduction zones. In terms of moment magnitude **M** (*13*), this corresponds to $\mathbf{M} \leq 6.5$ for the continental crust and $\mathbf{M} \leq 7.5$ for major subduction zones. Consistent with (*5*), a large earthquake will mean $M \geq$ **7.5**. {B. Romanowicz [*Geophys. Res. Lett.* **19**, 481 (1992)] has recently compiled seismic moment–fault length data for strike-slip earthquakes and places the transition from small to large strike-slip earthquakes at $M_0 = 6 \times 10^{26}$ dyne cm or, equivalently, $\mathbf{M} = 7.2$.} "Microearthquakes" or "microseismicity" typically means $\mathbf{M} \leq 3$. In areas where the seismicity rate is very low, Washington, DC, for example, some confusion can arise when the largest earthquake known to the area is a microearthquake. The largest earthquake of a set will usually dominate its radiated energy or seismic moment sum.
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10. L. C. Haar *et al.*, *Bull. Seismol. Soc. Am.* **74**, 2463 (1984); P. G. Somerville *et al.*, *ibid.* **77**, 322 (1987); T. C. Hanks and A. C. Johnston, *ibid.* **82**, 1 (1992).
11. P. Molnar and M. Wyss, *Phys. Earth Planet. Inter.* **6**, 263 (1972); M. Wyss and P. Molnar, *ibid.*, p. 279.
12. Seismic moment is the product of the average fault slip $\bar{u}$ over the faulted area A times the shear modulus μ of the source region, $M_0 = \mu\bar{u}A$. This fundamental measure of source strength [K. Aki, *Bull. Earthquake Res. Inst.* **44**, 73 (1966)] is a development more recent than Eqs. 2a and 2b, as originally formulated. C. H. Scholz [*The Mechanics of Earthquakes and Faulting* (Cambridge Univ. Press, Cambridge, 1990)] connects Eqs. 2a and 2b to their proper parentage.
13. T. C. Hanks and H. Kanamori, *J. Geophys. Res.* **84**, 2348 (1979).
14. Hanks [in (*9*)] derived Eq. 4a for small earthquakes ($b = 1$) from Eqs. 1a, 2a, and 3, although Kanamori and Anderson [in (*9*)], by assuming that N would scale as reciprocal faulting area (Eqs. 4a and 4b), showed that b should be equal

to 1 with respect to M_s. Similarly, J. B. Rundle [*J. Geophys. Res.* **94**, 12377 (1989)], assuming the validity of Eq. 4b, used Eqs. 1a, 1b, 2b, and 3 to derive $b = 1.5$ for large earthquakes, which has been recently confirmed observationally (*5*). Other studies that have developed Eqs. 1 through 4 to serve one purpose or another include T. C. Hanks, *ibid.* **84**, 2235 (1979); D. J. Andrews, *ibid.* **85**, 3867 (1980); K. Aki, in *Earthquake Prediction: An International Review*, D. W. Simpson and P. G. Richards, Eds. (American Geophysical Union, Washington, DC, 1981), pp. 566–574; G. King, *Pure Appl. Geophys.* **121**, 761 (1983); D. Turcotte, *ibid.* **131**, 171 (1989); A. Frankel, *J. Geophys. Res.* **96**, 6291 (1991). Readers who execute the algebra to reach Eq. 4b will note that $N, \Delta N \sim (\ell w)^{-1} w^{-1}$. Because w is customarily taken as a constant, this large-earthquake scaling reduces in fact to reciprocal fault-length scaling.

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19. A. McGarr, *Bull. Seismol. Soc. Am.* **66**, 33 (1976); _____ and R. W. E. Green, *ibid.* **68**, 1679 (1978).
20. The failure of Eq. 1a at small M is commonly attributed to sampling problems: a region large enough to have one $M = 5$ earthquake per year, say, would require a large number as well as downhole deployment of seismic systems to record and locate all the $M = 1$ earthquakes per year, about 10,000 of them. Obtaining uniform and consistent magnitudes for all these small earthquakes constitutes another formidable problem. The breakdown of the constancy of earthquake stress drops at small M can be restated as the inevitable consequence of f_{max}, the high-frequency band limitation of the radiated field of earthquakes [T. C. Hanks, *ibid.* **72**, 1867 (1982)]. Whether f_{max} arises from source or site, however, has not yet been decided. Aki (*3*) has championed the source-effect explanation; I am inclined to the influence of recording-site phenomena.
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Effects of Aerosol from Biomass Burning on the Global Radiation Budget

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An analysis is made of the likely contribution of smoke particles from biomass burning to the global radiation balance. These particles act to reflect solar radiation directly; they also can act as cloud condensation nuclei, increasing the reflectivity of clouds. Together these effects, although uncertain, may add up globally to a cooling effect as large as 2 watts per square meter, comparable to the estimated contribution of sulfate aerosols. Anthropogenic increases of smoke emission thus may have helped weaken the net greenhouse warming from anthropogenic trace gases.

Atmospheric aerosol particles have a noticeable effect on solar radiation and are sufficiently widespread for this effect to have implications for climate. However, their relatively short and irregular lifetimes have made it difficult to quantify their contribution to regional and global radiation budgets. Tropospheric aerosols (which have lifetimes of less than a week and an extremely patchy horizontal spatial distribution) may, in total, cool climate through their direct radiative effect by an amount comparable to the warming expected from a doubling of CO_2 (*1*).

Sulfate has been recognized as the dominant contributor to tropospheric aerosols over and near industrial continental areas (*2*). Smoke aerosol may have a similar role over and near tropical continental areas. The production of both sulfate and smoke aerosol from human activities is expected to have greatly intensified during the last century.

Recent estimates of the contribution of anthropogenic emissions of sulfate aerosol to climate change have been based on the combined approaches of biogeochemical cycling, atmospheric radiative transfer theory, and climate modeling (*2*). The direct and indirect contributions of the anthropogenic sulfate aerosol to the atmospheric radiation balance appear to be comparable. They are similar in magnitude and opposite in sign to the effects of the increase in atmospheric greenhouse gases over the last century. The global effect of aerosols from smoke adds further to this effect. Thus, the increase of aerosols over the last century may explain why the warming observed over this period is at the lower limit of that indicated by climate models (*3*) from the increased levels of greenhouse gases.

The global mass of smoke, M, in the atmosphere depends on the total rate, R, of biomass burning, the fraction, f, of burned material that goes into smoke, and the lifetime, t_s, of the smoke in the atmosphere

$$M = fRt_s \quad (1)$$

Most (more than 70%) of the burned biomass is in the tropics, with contributions from burning of forests for shifting agriculture and burning of virgin forest cleared for colonization, savannas, fuel wood, and agricultural wastes (*4*). Recent estimates of tropical biomass burning from the Food and Agriculture Organization statistics for the period 1975 to 1980 (*5*) suggest that burning in savannas dominates that in forests, in contrast to earlier estimates (*4*). The annual carbon burning from deforestation, savanna burning, and shifting agriculture is estimated to be 1300 to 3300 Tg of C (*5*) (1 Tg = 10^{12} g). An estimate of the contributions of fuel wood, agricultural wastes, and charcoal is 1000 to 2000 Tg of C per year (*6*), or, in sum, the C burned per year globally is likely 2600 to 5000 Tg.

Measurements of the fraction of smoke produced (*7–9*) suggest that

$$f = 0.03 \text{ g of smoke per gram of C} \quad (2)$$

Thus, for $R = 3800$ Tg of C per year, 114 Tg of smoke is produced per year. This value is comparable to the global annual anthropogenic sulfate production of 110 Tg per year (*2*). The smoke has approximately the same distribution of particle sizes as sulfate aerosols. In particular, the bulk of the mass and effectively all of the optical effects are associated with or determined by particles of this size in the aerosol accumulation mode with a mass-averaged radius of around 0.3 μm. Because these particles include hygroscopic components, as do sulfate particles, they should have a comparable lifetime. For a lifetime of 6 days (*2*), Eq. 1 implies that the global smoke loading is $M = 3.7 \times 10^{-3}$ g m^{-2}. The optical thickness of an aerosol layer is

$$\tau = \tau_a + \tau_s \quad (3)$$

where τ_a is the absorption optical depth and τ_s is the scattering optical depth. Optical

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