SEISMOLOGY

Dynamic triggering of earthquakes

After an earthquake, numerous smaller shocks are triggered over distances comparable to the dimensions of the mainshock fault rupture, although they are rare at larger distances. Here we analyse the scaling of dynamic deformations (the stresses and strains associated with seismic waves) with distance from, and magnitude of, their triggering earthquake, and show that they can cause further earthquakes at any distance if their amplitude exceeds several microstrain, regardless of their frequency content. These triggering requirements are remarkably similar to those measured in the laboratory for inducing dynamic elastic nonlinear behaviour, which suggests that the underlying physics is similar[1,2].

We assume that some aftershocks are dynamically triggered[3] and therefore that, within aftershock zones, mainshock-generated dynamic deformations must be sufficiently large to trigger earthquakes. The square root of the rupture area, $\sqrt{\Sigma}$, provides a useful approximate measure of both rupture dimension and the aftershock zone[4]. We measured the horizontal component of the seismic waves' peak ground-motion velocity (PGV, which approximates the peak shear strain when divided by the shear wave or phase velocity[5]), recorded at distances of 0.1 km to 5,300 km from earthquakes with magnitudes, M, of 4.4 to 7.9 (Fig. 1a).

To define all the aftershock zones by the same scaled distance, for each earthquake we divide the distances corresponding to each PGV measurement by an estimate of $\sqrt{\Sigma}$ (from the database of Finite-Source Rupture Models at www.seismo.ethz.ch/srcmod). The scaling by $\sqrt{\Sigma}$ eliminates any resolvable dependence on earthquake size: this can be explained by a model of seismic radiation from a fault with area Σ. Thus, the PGVs at a scaled distance around unity define an approximate minimum triggering threshold of several to ten microstrain (Fig. 1b).

Remote, spatially widespread triggering has been most clearly observed following the earthquakes at Landers[6] ($M=7.3$) and Denali[7] ($M=7.9$) in California and Alaska, respectively, and (although less pronounced) after the Hector Mine earthquake[6] ($M=7.1$) in California. PGVs for these seem consistent with the inferred threshold (Fig. 1b) at all but one site (the geothermal area of Long Valley, California). Landers data exist at very remote distances only for non-triggered sites, but these provide a lower bound on triggering deformations as they are off the azimuth of expected focusing. The absence of triggering where strain amplitudes exceed the proposed threshold implies that large deformations may be a necessary but not sufficient condition.

These observations indicate that remote triggering may require exceptionally large dynamic deformations, perhaps as a result of strong directivity[6,7], thereby explaining why this occurs only rarely. That a simple amplitude threshold seems to account for both the occurrence and absence of triggering, and the fact that the PGVs come from signals with very different frequency contents (dominant frequencies are roughly proportional to Σ), also implies that the mechanisms of dynamic triggering do not depend strongly on frequency.

We have proposed a model in which dynamic deformations promote earthquake failure by mechanisms involving dynamic nonlinear elasticity and slow dynamics[1]. We base this on the similarities between our seismological and laboratory observations[1] (Fig. 1b, and see supplementary information), field observations[8], and on the modelling[2] of dynamic nonlinear elasticity. Application of dynamic strains of the order of several microstrain seems to be required both for dynamic triggering of earthquakes and for the significant nonlinearity that arises from modulus reduction (softening) in laboratory and field experiments and in models[1,2]. Another similarity is a lack of dependence on loading frequency over bandwidths spanning several orders of magnitude[2]. If a fault is in a critical state near to failure, we suggest that softening leads to failure[1,9].

Although we have not considered the extended durations of triggered earthquake sequences, our model explains them through the recovery that follows dynamic softening by waves (that is, the slow dynamics) from both the mainshock and from creep following subsequent, locally triggered earthquakes[1]. We should be able to validate this model as new earthquake and lab data become available.

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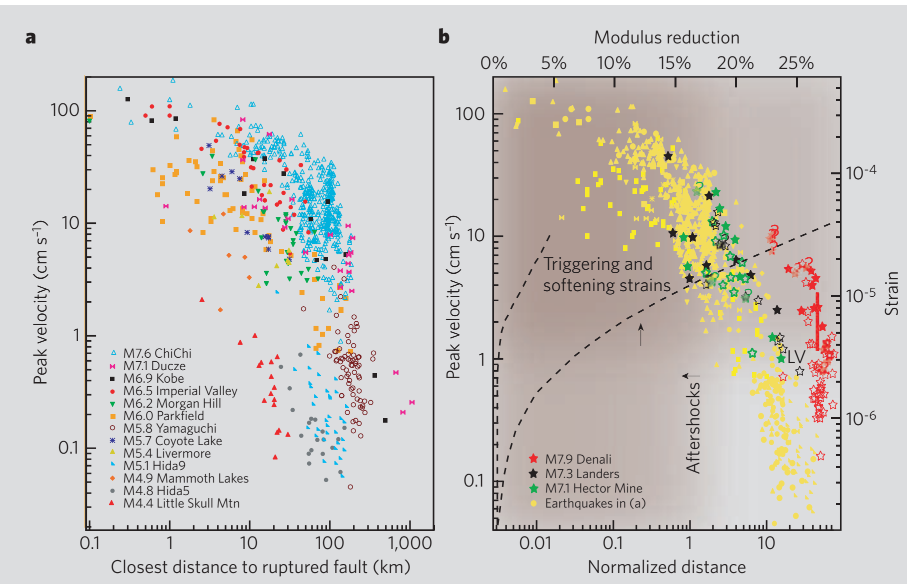

Figure 1 | **Dynamic deformation scaling. a,** Plot of measured peak ground velocities (PGVs) against distance (data from Pacific Earthquake Engineering Research Center; Institutions for Research in Seismology; Japan's National Research Institute for Earth Science and Disaster Prevention K-Net; and Seismology Lab, University of Nevada, Reno). **b,** Plot of measurements in **a** against distance, normalized by rupture dimensions, showing almost identical scaling. Deformations at normalized distances of less than about 1 (left, light brown) must be sufficient to trigger aftershocks; triggering strains lie above the smallest PGV in this normalized-distance range (top, darker brown). This triggering threshold range is consistent with PGVs for three earthquakes that triggered seismicity remotely, measured at sites that did (filled stars) and did not (open stars) experience triggered seismicity (light shading, ambiguous observations). Red bar, additional Denali triggering PGVs[10]; dashed curves bound laboratory measurements of modulus reduction against dynamic loading strain amplitude for various rock types, pressures and saturations; LV, Long Valley.

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