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Alaskan Earthquake, 27 March 1964: Vertical Extent of Faulting and Elastic Strain Energy Release

Abstract. *The residual displacement field indicates that the primary fault responsible for the great Alaskan earthquake extended to depths of 100 to 200 kilometers and came to within 15 kilometers of the surface. The vertical extent is an order of magnitude greater than reported for all other earthquakes. Approximately 10^{25} ergs of elastic strain energy was released. About 12,000 aftershocks ($M_L \geqq 3.5$) probably occurred in a 69-day period after the main shock. One-half the strain rebound occurred the first day with the main shock contributing one-fourth of the total.*

It has become possible to determine several important properties of faults associated with major earthquakes, such as initial slip direction, orientation and horizontal extent of the fault plane, and rupture velocity. The vertical extent of faulting and the elastic strain energy release are parameters which have been estimated, but dilemmas arise in connection with the results. The vertical extent can be deduced from the rate of decrease of the residual displacement or the seismic energy with distance. When the data from the San Francisco earthquake (1906) were applied to the San Andreas fault, fault depths of 2 to 10 km were obtained (*1, 2*). It is difficult to reconcile these small values with the 436-km length of the fault break. Estimates of strain energy release depend on the extent of faulting, and they also are in doubt.

The Alaskan earthquake (27 March 1964) produced residual vertical displacements which could be measured to distances of 200 km on the basis of tide gauge records and shoreline changes (*3*). These data are unmatched in extent and precision and offer an excellent opportunity for estimating the vertical extent of faulting and the elastic strain energy release.

The length of the primary fault is approximately 800 km as estimated from the extent of the belt of aftershocks. The distribution of polarity of first motion is consistent with a nearly vertical fault plane (*4*). Fault length determined from the spectra of surface waves is about 650 km, and a near-vertical fault plane is uniquely indicated (*5*). The residual vertical changes projected on a section normal to the strike of the fault are shown in Fig. 1. Locally on Montague Island, uplift exceeded 10 m. These points are not plotted since they are apparently secondary features, subsidiary to the regional uplift and associated with the zone of maximum flexure. Striking features of the residual displacement profile are (i) the occurrence of a zone of zero elevation change rather than a scarp separating the region of uplift and subsidence; (ii) zones of maximum elevation change and greatest flexure which roughly define the width of the belt of epicenters; (iii) gradual decrease in residual displacement to distances of 150 to 250 km; and (iv) asymmetry in the curves of uplift and subsidence.

The last feature may be due to slight dip in the fault plane. An approximate interpretation of the first three features can be made representing the fault as a vertical, rectangular dislocation sheet in a half-space. The displacement fields for such a source can readily be computed (*2, 6*) and fitted to the observed displacements with the vertical extent of faulting as an adjustment parameter. Two types of theoretical curves are drawn through the data in Fig. 1. The curves with $d = 0$ correspond to a fault reaching the surface and show the expected scarp. The curves with $d = .04\,L \sim 16$ km (L being the fault half-length) show zero displacement above the fault and a zone of flexure separating the uplifted and down-dropped blocks. The two curves are indistinguishable at distances greater than 30 km in the direction normal to the fault. At these larger distances the curves are sensitive

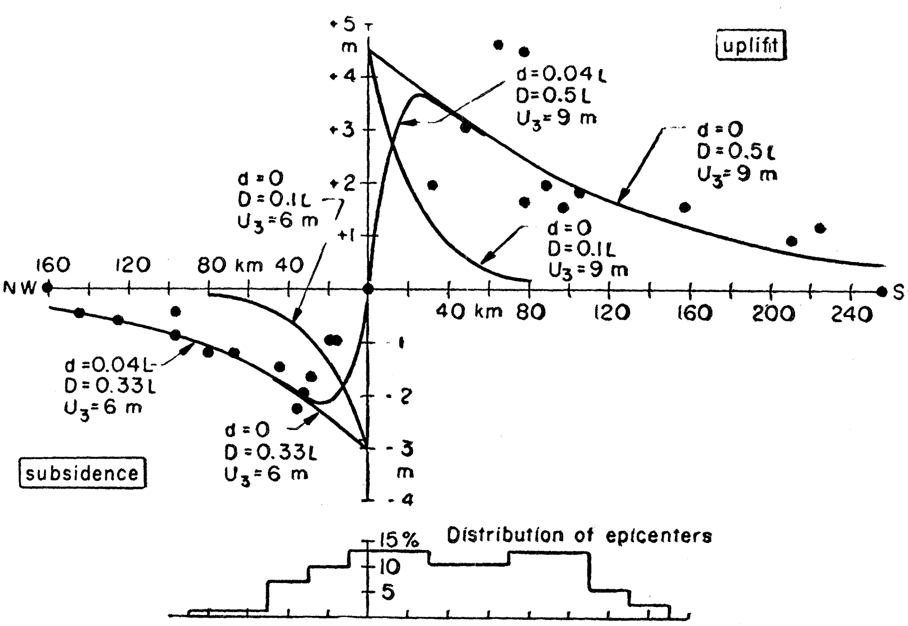

Fig. 1. Residual vertical displacements projected on a section normal to and bisecting the fault. Parameters for theoretical curves are d, depth to top of fault; D, depth to bottom of fault; L, half-length of fault (400 km); U_3 vertical slip; index of seismicity, shown at bottom, is percentage of aftershocks in zone 20 km wide in the 3 days after main shock.

to the depth of the lower boundary of the fault, as can be seen by comparing the cases $D = 0.1\,L \sim 40$ km with $D = 0.33\,L \sim 130$ km and $D = 0.5\,L \sim 200$ km. The last two cases fit the data in the down-dropped and uplifted blocks respectively, the first case showing much too rapid a decrease in displacement with distance. We conclude that the primary fault came to within 15 to 20 km of the surface and extend to depths of 100 to 200 km.

A large depth of faulting is also required to explain the residual strain of 10^{-8} observed in Hawaii (Fig. 2). If the dislocation representation is used and the strike-slip is 5 to 6 m (*7*), strains in the range 10^{-9} to 10^{-8} are computed for Hawaii when $D \sim 1.0\,L$, whereas 10^{-10} to 10^{-9} are found when $D = 0.1\,L$ (*8*).

An estimate of the elastic strain energy released by the Alaskan earthquake can be made by evaluating the volume integral of the strain energy density function or computing the energy required to produce the dislocation. If the rigidity is 5×10^{11} dynes/cm^2, fault dimensions are 800 by 200 km, and horizontal and vertical slips are 6 m, we find stresses of 10^7 dynes/cm^2 adjacent to the fault and an energy release of 10^{25} ergs. Presumably this is an upper bound for the energy released as seismic waves. With the Gutenberg-Richter relation between magnitude and energy, $\log E = 11.8 + 1.5\,M$, and the value $M = 8.4$ found from surface waves, a value of 3×10^{24} ergs is indicated for the energy in seismic waves. A line distribution of 100 underground nuclear explosions totaling 100 megatons each would correspond in seismic energy to such a source. It may be compared to the value 10^{28} ergs which represents the energy reaching the earth's surface annually as heat from the interior.

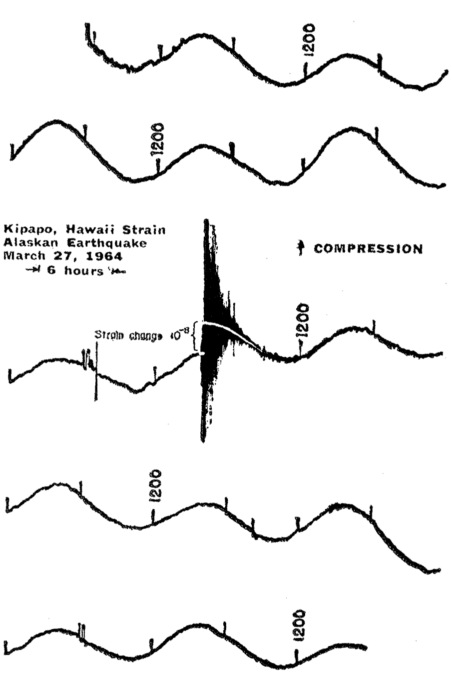

Fig. 2. Residual strain of 10^{-8} recorded at Kipapa, Oahu, on strain seismograph. Several days of recording before and after earthquake show strains associated with tidal loading only.

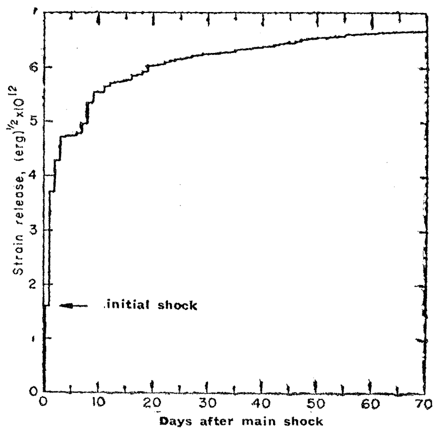

Fig. 3. Cumulative strain release for 69-day period for Alaskan earthquake and aftershock sequence. Ordinate is square root of energy computed from magnitudes, and is proportional to strain change. Except for main shock, daily values represent sums of strain release.

Thirteen days after the main shock we installed standard Wood-Anderson seismographs in Seward and on Middleton Island (*9*). We have used these instruments to estimate some aftershock statistics as follows: In a 57-day period beginning 9 April, N earthquakes were indicated per 0.1 unit in magnitude M_L, where $\log N = a - bM_L$, $a = 6.6$, and $b = 1.1$. A correction for radius of perceptibility was used on the assumption that the seismic release was constant over the fault. Approximately 800 shocks were actually recorded in the magnitude range $M_L \geqq 3.5$. The coefficient b was found to be independent of time. For the period 28 March to 9 April, the U.S. Coast and Geodetic Survey reported 103 earthquakes with $M_L \geqq 5.3$ (*10*). With these data to determine an appropriate value of a, the number of shocks in the range $3.5 \leqq M_L \leqq 5.3$ could be inferred. For the 69-day period following the earthquake, 12,000 shocks are estimated to have occurred with $M_L \geqq 3.5$.

The characteristics of the cumulative strain release for these 12,000 shocks are summarized in Fig. 3. The ordinate is $E^{\frac{1}{2}}$, which is proportional to strain (*11*) and is computed from the magnitude-energy relationship $\log E = 9.9 + 1.9M_L - 0.024M_L{}^2$. We see that the total seismic energy released in the 69-day period was about 5×10^{25} ergs. Approximately half the strain rebound occurred in the first day, with the main shock contributing one-fourth the total. It is unusual that no aftershock with $M > 7$ occurred, the aftershock strain release occurring mainly from the large number of smaller shocks.

The vertical extent found for the Alaskan earthquake fault exceeds by one or two orders of magnitude the values found for other earthquakes. This difference may be associated with the larger magnitude reported for the Alaskan fault. In the case of the San Francisco earthquake, this explanation does not suffice since the magnitude difference between the two events is small. Most displacement observations fell within 5 km of the San Andreas fault, and anomalously low rigidities in the fault zone (*12*) could have resulted in a misleading displacement-distance relationship.

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