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Tectonic Deformation Associated with the 1964 Alaska Earthquake

The earthquake of 27 March 1964 resulted in observable crustal deformation of unprecedented areal extent.

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The epicenter of the earthquake of 27 March 1964 (28 March 1964 Universal Time) and its zone of aftershocks lie within a well-defined belt of shallow and intermediate-depth earthquakes which follows the Aleutian Trench and Volcanic Arc from Kamchatka to south-central Alaska, where the arc enters the North American continent (Fig. 1). The active and dormant volcanoes of the Wrangell Mountains, shown on Fig. 1, may be an eastward extension of the Volcanic Arc. The Aleutian Arc is one of the most seismically active areas of the festoons of volcanic arcs and associated deep ocean trenches which bound the Pacific Ocean from Alaska to New Zealand and from Antarctica to Central America. The seismic and volcanic activity of the arcs of the Pacific, as well as arc structures elsewhere, are manifestations of one type of active diastrophism, or mountain building. The geologic record indicates that comparable arcuate structures in the remote past played a significant role in the formation of mountains and the growth of continents. Consequently, the structure of modern arcs and the nature of the deformations that occur along them are of interest to geologists and geophysicists concerned with the origin of mountains and continents. Most of our knowledge of the deformations that produce large earthquakes along active arcs comes from analysis of the elastic waves they generate. Direct observation of the surface displacements that sometimes accompany these deformations is commonly limited, because all or much of the displacement is submarine.

Perhaps the most notable aspect of the Alaskan earthquake was the great extent and amount of the changes in land level that accompanied it. From the epicenter in northern Prince William Sound, the zone of surface deformation extends for 800 kilometers roughly parallel to the trends of the Aleutian Volcanic Arc and Trench and the coast of the Gulf of Alaska. Where the northeastern end of the arc intersects the continent at an oblique angle, the deformation can be observed in an almost complete section across the arc from Middleton Island, near the seaward edge of the continental shelf, to the west shore of Cook Inlet (Fig. 2).

This article makes available a summary of the basic data acquired during the 1964 field season on the tectonic deformations that accompanied the earthquake. It substantially enlarges upon the information on land-level changes published in a preliminary U.S. Geological Survey report based on a 2-week reconnaissance made immediately after the earthquake (*1*). Some tentative interpretations of the mechanism of the earthquake are advanced here on the basis of the data now available. They may require modification or revision as additional results of continuing and planned investigations into the manifold aspects of this major tectonic event are made known.

Methods

The vertical tectonic movements in coastal areas were determined mainly by making more than 800 measurements of displacement of intertidal sessile marine organisms along the long, intricately embayed coast (Fig. 3). These measurements were supplemented at 16 tidal bench marks by coupled pre- and post-earthquake tide-gauge readings made by the U.S. Coast and Geodetic Survey, and by numerous estimates of relative changes in tide levels by local residents. The amount and distribution of the vertical tectonic movements inland from the coast were defined along the highways connecting the cities of Seward, Anchorage, Valdez, and Fairbanks by the U.S. Coast and Geodetic Survey's prompt releveling of previously surveyed first-order level lines tied to tidal bench marks at Seward, Anchorage, and Valdez.

In measuring the displacement of sessile marine organisms use is made of the zonation of plants and animals between tide marks that has long been recognized in different parts of the world by marine ecologists. The intertidal zone along the predominantly steep and rocky coastline of south-central Alaska is inhabited by certain species of sessile organisms, notably barnacles, mussels, and algae, whose vertical growth limits are commonly well defined. In particular, the common acorn barnacle, *Balanus balanoides* (Linnaeus), is widely distributed on the rocky coast and forms a prominent band with a sharply de-

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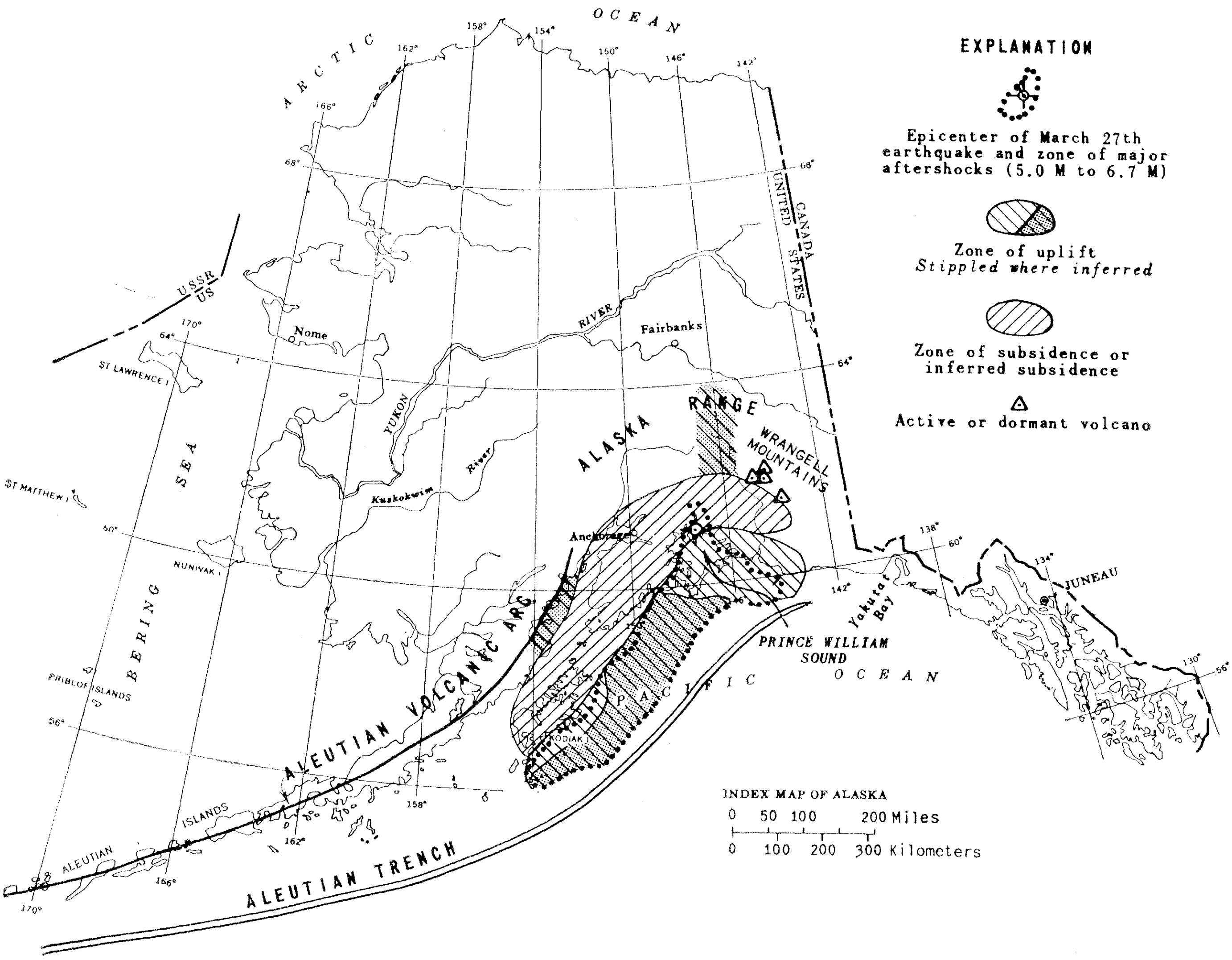


Fig. 1. Index map, showing the relationship of the Aleutian Arc to the earthquake of 27 March 1964, its belt of aftershocks, and zones of change in land level.

fined upper limit which is readily recognizable on most shores. This upper limit corresponds roughly to the top of the Balanoid or Midlittoral Zone of Stephenson and Stephenson; to Zone 2, the High Tide Region, or the Upper Horizon of Ricketts and Calvin; and to the Upper Intertidal Zone of Rigg and Miller (*2–4*). No data on the normal height of barnacle growth relative to tide levels could be found for south-central Alaska. However, experience at places on the Pacific Coast of North America in areas where both the prevailing type of tide and the tidal range are similar to those in parts of Prince William Sound and on the coast of the Gulf of Alaska indicates that this height is close to annual mean high water (*3–5*).

At localities where barnacles do not occur, both the common olive-green rockweed (*Fucus furcatus* Agardh), whose upper growth limit is near that of the barnacles, and the dark gray encrusting alga [*Ralfsia verricosus* (Areschoug) J. Agardh], which commonly occupies the splash zone immediately above the barnacles, served as useful datums for measuring land-level changes. The growth limit of the algae, however, appears to be strongly influenced by exposure to sunlight as well as by the tides; thus their vertical range is generally more variable than that of the barnacles.

The field procedure was to measure the height of the upper limit of barnacle growth (the "barnacle line") above or below water level at any stage of tide (Fig. 3). On steep rocky slopes that are sheltered from heavy surf this line is sharply defined and can be readily determined to within 15 centimeters or less; on sloping shores or shores exposed to heavy surf it tends to be less regular. In the usual case, where the barnacle line was above water, it was measured with a hand level or surveyor's level and stadia rod. Where the barnacle line was visible under water, its depth below the surface was measured directly with the stadia rod. The stage of tide at the time of measurement was then determined from the U.S. Coast and Geodetic Survey table of predicted tides for the closest reference station, and the position of the barnacle line relative to lower low water was calculated. For those stations close to the 16 U.S. Coast and Geodetic Survey tide gauges that were installed in the area following the earthquake, we later made corrections to the actual, rather than the predicted, tides. During the period of field work, tides rarely deviated by as much as ½ meter from predictions. Abrupt changes in the height of the barnacle line in local areas of similar tide range provided a fairly precise method for detecting vertical fault displacements of more than 1 to 2 meters in any given locality.

Absolute uplift or subsidence at any given locality was taken as the vertical difference between the measured elevation of the pre-earthquake barnacle line and the "normal" upper growth limit for the barnacles, as determined empirically at 12 tidal bench marks where the amount of vertical displacement was known from tide-gauge readings (*6*). The upper limit of barnacle growth depends mainly on the ability of yearling barnacles to survive prolonged exposure to air and on the tidal characteristics at any given locality (*7*). To a lesser extent it depends upon a number of other factors which may locally cause the barnacle line to deviate as much as ⅓ meter from its "normal" height. Wave action during the lowest annual neap tides and protection from desiccation in shady locations tend to elevate the upper growth limit; exposure to fresh water, near large streams or tidewater glaciers, tends to depress it. Annual variations in sea level may cause further slight upward or downward shifts of the upper growth limit of the barnacles. It was found that the pre-earthquake barnacle line is at mean high water, or no more than 15 centimeters below it, for the lower tidal ranges of 1.9 to 3.3 meters which prevail along most of the coast of the Gulf of Alaska and in Prince William Sound. For the higher mean ranges of up to 5.1 meters in Cook Inlet and Shelikof Strait, the barnacle line lowers progressively to as much as 45 centimeters below mean high water.

The validity of using the empirically determined upper limit of barnacle growth was confirmed by the agreement found between the observed heights to which new post-earthquake barnacles were growing after the spring neap tides and the predicted heights for those localities. Land-level changes

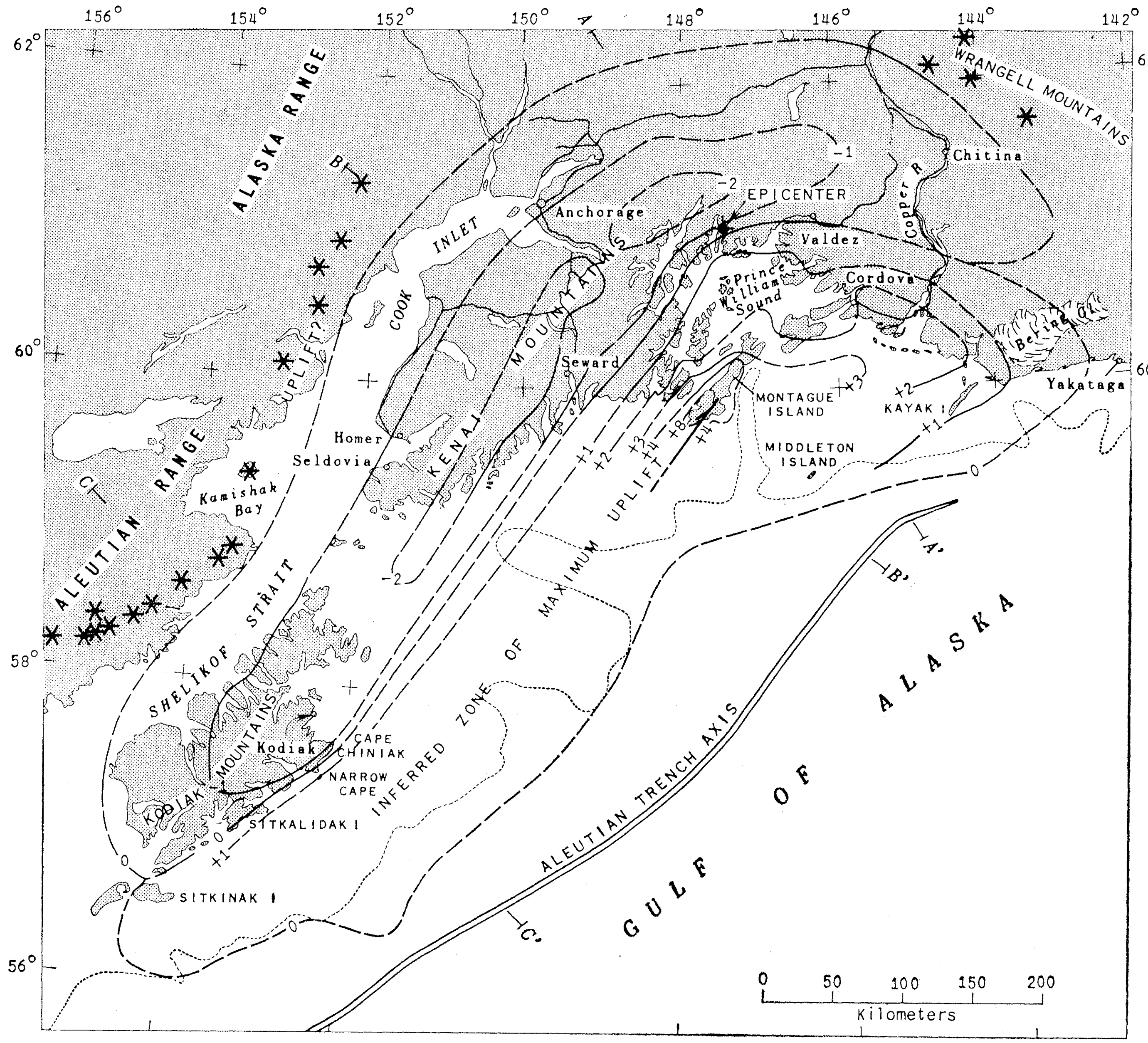
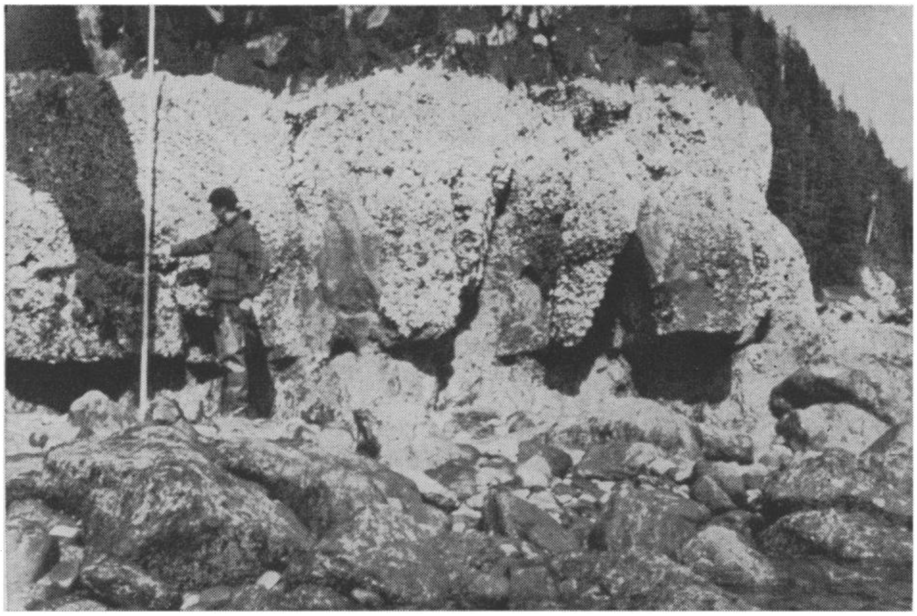
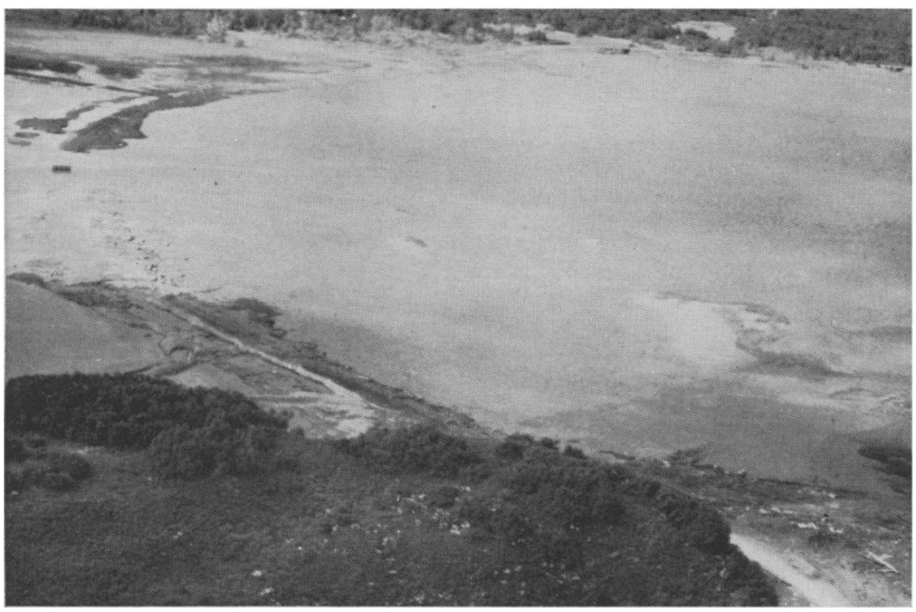


Fig. 2. Tectonic uplift and subsidence in south-central Alaska. The land-level change, in meters, is shown by the contours, which are dashed where approximate or inferred. The outer edge of the continental shelf, at −200 meters, is indicated by a dotted line, and active or dormant volcanoes are shown by stars. A-A′, B-B′, and C-C′ are lines of sections shown in Fig. 6.





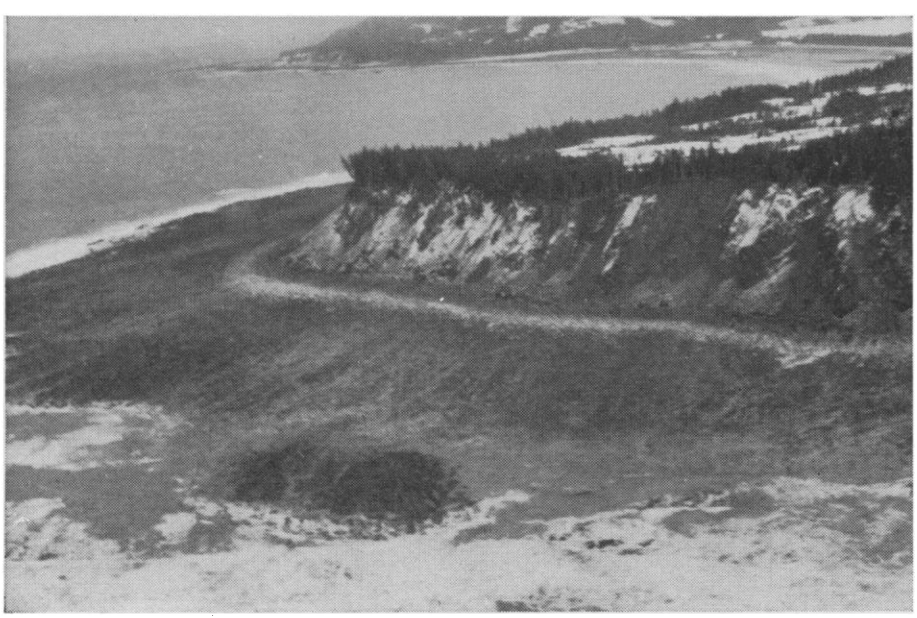


Fig. 3 (top). Measuring the height of the pre-earthquake upper limit of barnacle growth, or barnacle line, above the present water level at Glacier Island in Prince William Sound. The sharply defined upper limit of barnacle growth is typical of much of the Prince William Sound area. Fig. 4 (center). Road along the shore of Middle Bay on Kodiak Island. The road is inundated daily, due to tectonic subsidence of about 1⅔ meters and to an unknown, but probably substantial, amount of local settling of unconsolidated deposits. The photograph was taken 20 July 1964 at 1.2-meter tide. Fig. 5 (bottom). The southwest tip of Montague Island at half tide, 3 days after the earthquake. The surf-cut platform shown here has been exposed by about 8 meters of uplift. Prior to the earthquake the approximate half-tide line was close to the base of the prominent white zone of barnacles below the sea cliff. The entire reef later turned white because of desiccation of calcareous organisms in the former interidal zone.

determined by the barnacle-line method are generally within ⅓ meter of changes estimated by local residents or found by means of other techniques; even under the least favorable combination of circumstances the error of the barnacle-line method is probably less than ⅔ meter.

Throughout the area of land-level change a new post-earthquake line of yearling barnacles was well established above or below the normal growth limit by the end of the field season. The height of this new barnacle line below the pre-earthquake limit of barnacle growth furnished a direct measure of the maximum amount of uplift; its height above the older barnacles indicated the minimum amount of subsidence.

This same technique of measuring elevation differences between the upper limits of dead and living barnacles was used by Tarr and Martin (*8*), 8 years after the Yakutat Bay earthquake of 1899, as a criterion of the uplift that occurred during that quake.

There are many other indications of subsidence and uplift along the affected shorelines. In subsided areas vegetation has been killed by immersion in salt water, and beach berms have been shifted landward and are built up to the new, relatively higher sea levels. Stream deltas are also building up to the higher sea levels, and some former beach-barred lakes have become tidal lagoons. Inundation of coastal roads and shoreline installations and property on the Kenai Peninsula and the Kodiak Islands is a direct result of the regional subsidence (Fig. 4).

The amount of uplift could be determined directly in some areas by measuring the height of the zone of dead algae and barnacles which were raised above the reach of the tides. Qualitative indications of uplift include new reefs and islands, raised sea cliffs and sea caves, and drained lagoons. Figure 5 is a photograph, taken 3 days after the earthquake, of the surf-cut platform that was exposed at the southern end of Montague Island by about 8 meters of uplift. Uplift resulted in mass extermination of sessile intertidal fauna and flora. The effects of changes in land level on intertidal organisms, as well as other biological effects of the earthquake, have been summarized by Hanna (*9*). Navigability of waterways and harbors throughout the eastern part of Prince William Sound and the coastal area to the east of Cordova was impaired by the uplift. At many localities, canneries, docks, waterfront homes, and piers are now inaccessible by boat except at extreme high tides.

Regional Uplift and Subsidence

The areal distribution and approximate amounts of tectonic changes in land level that accompanied the earthquake are shown in Fig. 2 and the profile lines of Fig. 6. The coastal area south of a line extending along the southeast coast of Kodiak Island through the western part of Prince William Sound to the vicinity of Valdez has been elevated, and the area north and west of this line has been lowered. To the east the zone of deformation appears to die out between the Bering Glacier and Yakataga. The seaward limits are not known, although parts of the continental shelf as far south as Middleton Island and southwest to Sitkinak Island have been elevated. The northwestern limit of deformation extends at least to the west side of Shelikof Strait and Cook Inlet. Its inland limit is known only along the highway between Valdez and Fairbanks, where it extends northward at least to the latitude of the Wrangell Mountains, and possibly into the Alaska Range.

The major area of uplift is about 800 kilometers long and trends northeast from southern Kodiak Island to Prince William Sound, and east-west to the east of Prince William Sound. It includes the southern and eastern parts of Prince William Sound, the coastal area as far east as Bering Glacier, and part of the contiguous continental shelf, as shown by uplift on Kayak and Middleton islands. The average uplift within the zone in Prince William Sound, where the most detailed study was made, is about 2 meters. The maximum measured uplift on land is 10 meters at the southwest end of Montague Island and more than 15 meters offshore from Montague Island (*10*). The uplift of Middleton Island demonstrates that the tectonic deformation extended at least 195 kilometers south-southeast of the epicenter almost to the margin of the continental shelf. The exact amount of uplift on the island, however, is uncertain; available estimates range from 1 to 3 meters. Uplift also occurred along the extreme southeastern coasts of Kodiak Island, Sitkalidak Island, and part or all of Sitkinak Island (Fig. 2). The maximum measured uplift on Sitkalidak Island, as determined from displacement of barnacles, was ⅖ meter. The estimated uplift on Sitkinak Island is ⅓ to ⅔ meters. The maximum known uplift on Kodiak Island probably occurs at Narrow Cape, 50 kilometers due south of Kodiak; it is estimated to be at least ⅔ meter, and possibly as much as 1½ meters.

The areal distribution and initial direction of water motion of the seismic sea waves strongly suggest that the belt of uplift also embraces a large segment of the continental shelf and slope.

Changes in land level were measured in detail between the zones of uplift and subsidence in Prince William Sound and, where possible, along the southeast coast of Kodiak Island, to determine whether the uplifted and lowered areas are separated by a fault. These measurements showed no abrupt changes of level indicative of vertical fault displacement but, rather, a north and northwestward tilting about the axis of zero change in elevation (Fig. 6). The slope on the tilted surface in the vicinity of the zero line in Prince William Sound is northward to northwestward at a maximum rate of 11 centimeters per kilometer. Data on the configuration of the warping in the vicinity of the zero line on Kodiak Island, although less conclusive, suggest that here, too, warping results from pronounced northwestward tilting and possible local flexure without detectable surface faulting.

There are two areas north and northwest of the zone of subsidence where minor amounts of uplift may have occurred during the earthquake (Fig. 1). One is in the Alaska Range, where post-earthquake releveling along the Richardson Highway indicates as much as ⅓ meter of uplift relative to the determinations of earlier surveys made in 1944 and 1952 (*11*). It is not possible to determine how much, if any, of the change occurred during the earthquake of 27 March. The second area of possible uplift is along the northwest shore of Cook Inlet, where residents of Iliamna and Tuxedni bays report anomalously low tide levels following the earthquake which suggest uplift on the order of ⅓ to ⅔ meter.

The main area of known uplift includes at least 60,000 square kilometers. However, the trend of the contours of uplift in the northeastern part of the area, the presence of a fringe of uplift along the southeast coast of Kodiak Island, and the distribution of seismic sea waves and aftershocks suggest that the zone of uplift also includes the continental shelf and part of the continental slope within the belt of major aftershocks, or a total area of about 90,000 square kilometers. Furthermore, it is possible that part, or all, of the continental slope between the edge of the continental shelf and the axis of the Aleutian Trench may also have been included in the uplift.

The zone of subsidence includes the northern and western parts of Prince William Sound, the western segment of the Chugach Mountains and portions of the lowlands north of the mountains, most of the Kenai Peninsula, and almost all of the Kodiak Islands group. It forms an asymmetrical downwarp, 800 kilometers long and approximately 150 kilometers wide, whose axis is roughly along the crest of the coastal mountains. The axis of subsidence plunges gently, northeastward from the Kodiak Mountains and southwestward from the Chugach Mountains, to a low of 2⅓ meters on the south coast of the Kenai Peninsula. The total area of probable subsidence is about 110,000 square kilometers, and the average amount of subsidence is roughly 1 meter. The volume of crust that has been depressed below its pre-earthquake level is about 115 cubic kilometers.

From profiles A–A′ and B–B′ of Fig. 6 it may be seen that the vertical displacements about the zero axis of tilting are strongly asymmetrical. Along these lines, the volume of crust elevated above its pre-earthquake position is at least double the volume of the subsidence in the northern part of the deformed area.

The area of observable crustal deformation, or probable deformation, that accompanied the Good Friday earthquake—between 170,000 and 200,000 square kilometers—is larger than any such area known to be associated with a single earthquake in historic times. Comparable tectonic deformations probably have occurred during other great earthquakes, but where they occurred beneath the sea, along linear coastlines, or inland, it generally has not been possible to determine their areal extent with any degree of confidence. The 10 meters of absolute vertical displacement measured on Montague Island is known to have been exceeded only by the 14.3 meters (47 feet, 4 inches) of uplift that occurred during the earthquake of 1899 at Yakutat Bay, 320 kilometers to the east (Fig. 1). Substantially larger subsidences are known to have accompanied earthquakes in the past, although in many instances determination of the absolute amount of tectonic subsidence is complicated by surficial effects.

Submarine Uplift Indicated by Seismic Sea Waves

That the belt of uplift embraces a large segment of the continental shelf and slope, as shown in Fig. 2, is inferred from the areal distribution and initial direction of water motion of the seismic sea waves that accompanied the earthquake. The term "seismic sea waves" (also known as tsunamis or "tidal" waves) refers to the train of long-period waves generated in the Gulf of Alaska which caused extensive damage to the outer coast of the Kodiak Islands group, the Kenai Peninsula, and the coasts of British Columbia and the Pacific Northwest (*12*).

Seismic sea waves are gravity waves set up in the ocean by vertical disturbances of the sea bottom. The relative displacement of the sea bottom in the generative area can be determined from the initial water motion at suitably situated tide stations. Tide gauge records of the seismic sea waves outside the immediate area affected by the earthquake show an initial rise, indicating a positive wave resulting from upward motion of the sea bottom (*13, 14*). The initial direction of water movement along the coast of the Gulf of Alaska within the area affected by the earthquake is less clear, because there were no operative tide gauges and in many localities the water movements were complicated by (i) uplift and subsidence of the coast, (ii) local waves generated by numerous submarine and subaerial landslides, and (iii) seiches. However, most observers along the coast of the Kenaî Peninsula and the Kodiak Islands group report that the first strong motion of the waves that came in from the Gulf of Alaska was upward. An approximation to the amount of submarine uplift that generated the waves is suggested by half-wave amplitudes for the highest waves of about 5 meters at Kodiak, 7 to 8 meters at Seward, and 6 meters at Cordova (*1*). Along segments of the coast exposed to the open sea, run-up of the seismic sea waves was substantially higher.

The shape of the source area within which the train of seismic sea waves was generated can be approximated from an envelope of imaginary wave fronts projected back toward the wave source from observation stations along the shore at which arrival times are known. The distance traveled by the wave to any shore station is calculated from the velocity of propagation, which conforms closely to Lagrange's equation $V = (gh)^{1/2}$ (where g is the gravitational constant and h is the depth of water as determined from nautical charts), and from the elapsed time between the main shock and the reported arrival of the wave. As computed independently by Van Dorn of the Scripps Institution of Oceanography and Spaeth of the U.S. Coast and Geodetic Survey, from data of observation stations outside the area of deformation, the wave source lies in a broad area, between the Aleutian Trench axis and the coast, that extends from the northeastern limits of the zone of uplift on land southwestward to about the latitude of Kodiak (*13, 14*). The direction of travel and reported arrival times of the initial wave crest, which struck the shores of the Kenai Peninsula within 19 minutes and Kodiak Island within 34 minutes after the start of the earthquake, indicate that the wave crest was generated along one or more line sources within an elongate belt that extends southwestward from the axis of maximum uplift on Montague Island (Fig. 2).

It is assumed in the computations of travel distance that the waves were generated at the time the earthquake began or shortly thereafter. This assumption appears to be justified by the numerous reports of immediate withdrawals of water from uplifted coastal areas, which suggest that much, if not all, of the deformation occurred during the most violent tremors, estimated to have lasted 1½ to 4 minutes. Two apparently reliable accounts further suggest that, under conditions favorable for close observation, the vertical displacements were perceptible as distinct upward accelerations in the area of uplift or downward accelerations in the subsided area. The vertical displacements occurred at least fast enough to generate atmospheric waves that were recorded shortly after the earthquake on microbarographs at both the University of California at Berkeley and the Scripps Institution of Oceanography at La Jolla (*14, 15*).

From long experience with seismic sea waves, Japanese seismologists have found that the generative area of seismic sea waves for a given earthquake broadly corresponds to the distribution of major aftershocks (*16*). In the absence of direct information, the seaward extent of uplift indicated in Figs. 1 and 2 is arbitrarily inferred to coincide with the belt of major aftershocks. The actual area probably is no smaller, but could be somewhat larger, than the area outlined.

Surface Faults

Faulting associated with the earthquake of 27 March 1964 was found at the two localities shown in Fig. 7, through combined air reconnaissance and measurement of barnacle-line displacements. The fault along the southeast side of the island has been informally named the Patton Bay Fault; that on the northwest side of the island, the Hanning Bay Fault.

The longer of the two faults, the Patton Bay Fault, can be traced on the ground for 16 kilometers from near Patton Bay westward to the place where it strikes out to sea at Neck Point. Inland it is marked by a discontinuous line of landslides along the base of the rugged main ridge that forms the axis of the island (Fig. 8). Where the fault crosses level or gently sloping ground overlain by unconsolidated surficial deposits, there is commonly a line of open fissures and stepped topography along the fault trace. Vertical offset could be measured only where the fault intersects the coastline west of Neck Point. Here the displacement of barnacles indicated a net vertical movement of 5.2 meters, with the northwest side relatively upthrown and both sides of the fault uplifted relative to sea level. About 2½ meters of the displacement occurred along a fault which cuts beach gravels and the reef at the shore (Fig. 9). The remainder of the offset results from a pronounced seaward bending of the upthrown block within 300 meters of the fault, as illustrated in section A–A′ of Fig. 7. The fault trace is vertical where it can be seen in the sea cliff (Fig. 9). It is not known whether there is any component of strike-slip displacement. No evidence was found, during our brief reconnaissance investigation, of lateral displacement of roots or fallen trees

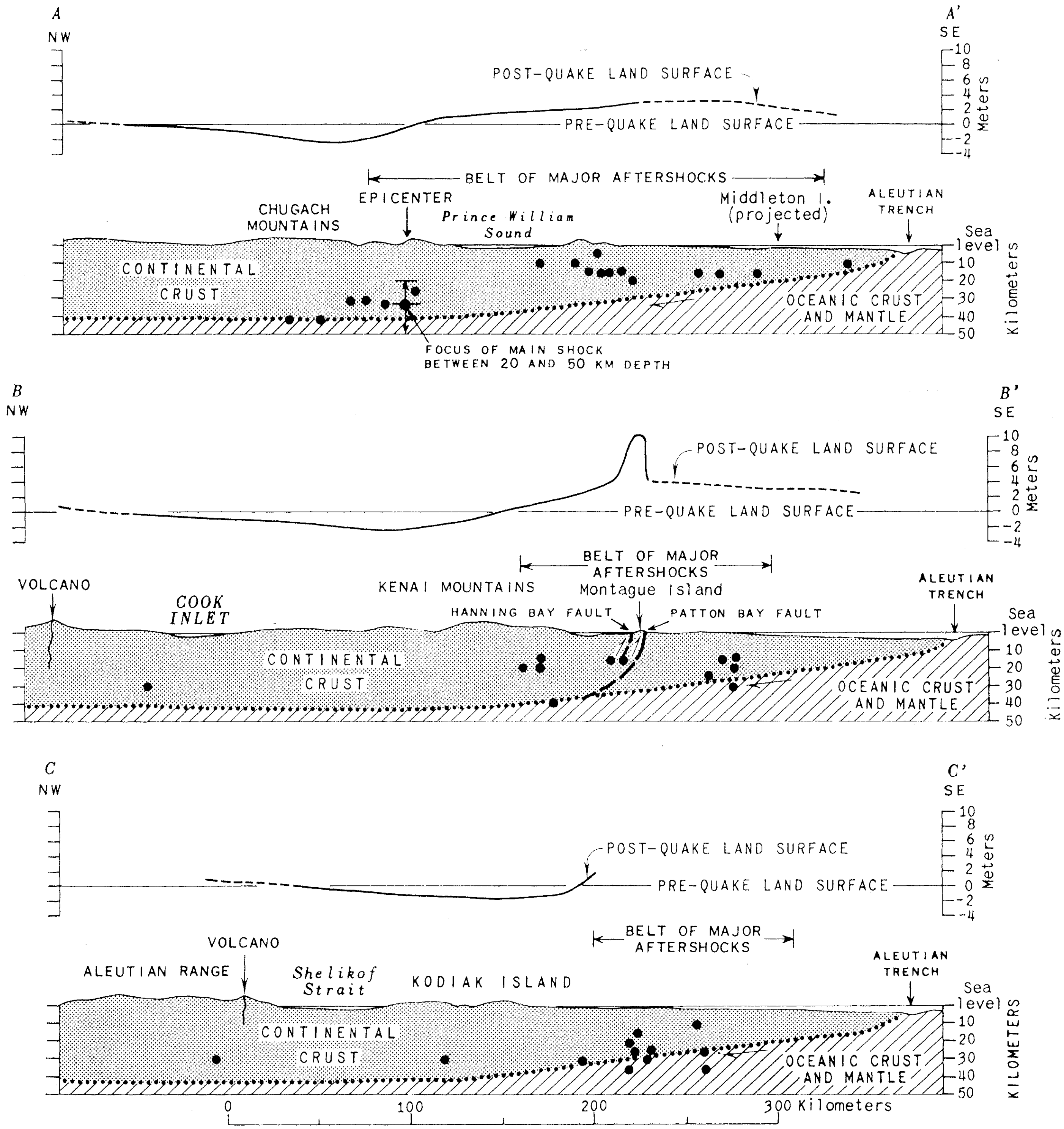


Fig. 6. Schematic sections across the Aleutian Arc, showing approximate (solid) and inferred (dashed) land-level changes, aftershock distribution, and geologic interpretations along the section lines. Heavy dots are earthquake hypocenters (U.S. Coast and Geodetic Survey) within 30 kilometers of the section lines. Section locations are shown in Fig. 2.

across numerous surface fissures in unconsolidated deposits that mark the fault trace. Displacement along the fault dies out to the northeast in the vicinity of Patton Bay. Fathometer profiles taken southwest of Montague Island are interpreted as indicating that the fault may extend southwestward as a zone of prominent scarps for at least 28 kilometers. Comparisons made by Malloy of bathymetric data obtained before and after the earthquake in the area 15 to 28 kilometers southwest of Neck Point show that displacement along the fault increases seaward to a maximum of about 10 meters (*10*).

The shorter of the two faults found on Montague Island is exceptionally well exposed for 4.8 kilometers, extending from the south shore of Hanning Bay almost to MacLeod Harbor. As with the Patton Bay Fault, the northwest side is upthrown relative to the southeast side, but both blocks are uplifted relative to sea level (Fig. 7, section A–A′). The fault plane dips at an angle of 70 to 80 degrees toward the northwest, and the movement is essentially dip-slip, with a maximum-measured left-lateral strike-slip component of 15 centimeters in surficial deposits near the southern limit of exposure. Vertical displacement varies rapidly along the fault strike; it reaches a maximum of 4 meters in bedrock and 5 meters in a beach ridge at the cove between Hanning Bay and MacLeod Harbor (Figs. 10 and 11). The fault trace inland from the coast is marked by a continuous line of toppled trees, ponded streams, landslides, and fissures. The fault dies out to the south before reaching MacLeod Harbor, and it could not be traced northward beyond the south shore of Hanning Bay.

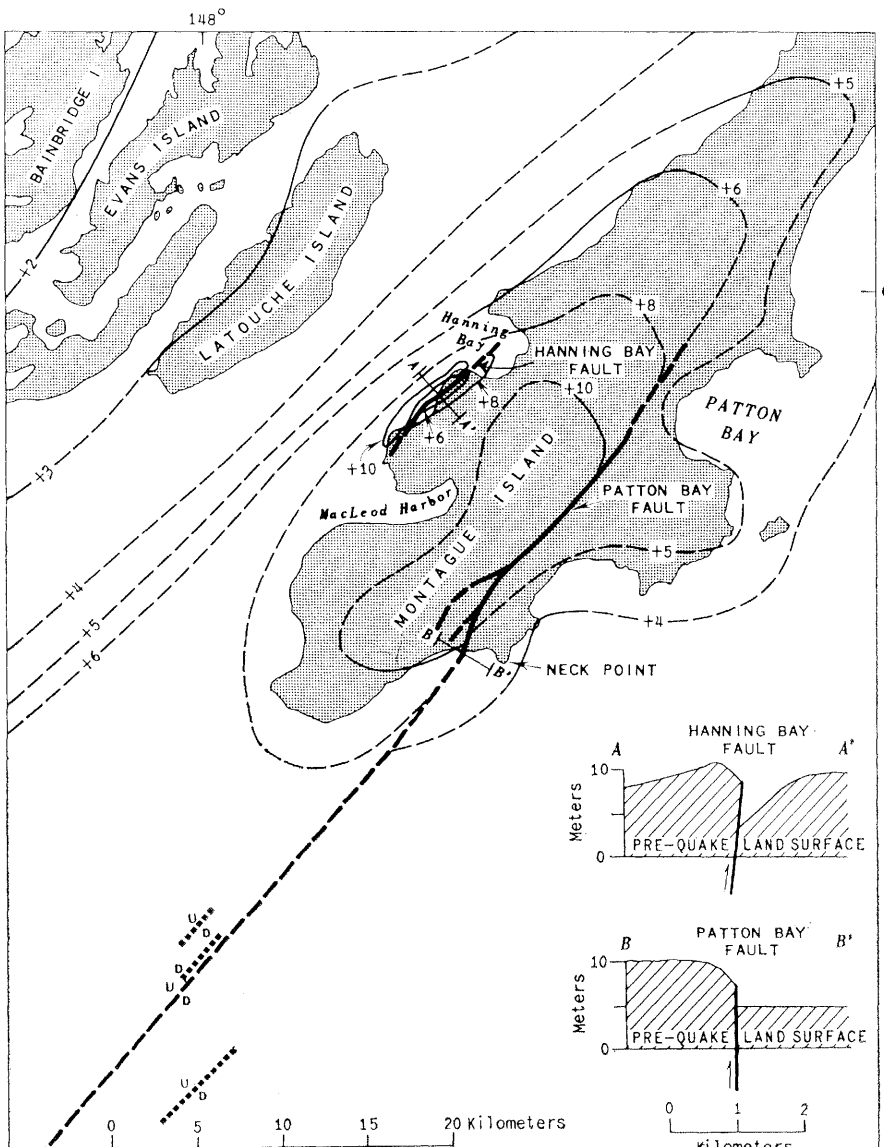


Fig. 7. Uplift and faulting on Montague Island and in the vicinity. The contours indicate approximate uplift, in meters; they are dashed where inferred.

The pattern of near-vertical and reverse faulting on southern Montague Island suggests that the displacements resulted from compressive stress oriented approximately normal to the trend of the Arc. Preliminary results of the U.S. Coast and Geodetic Survey's re-triangulation of part of the network of primary horizontal control stations (triangulation net) in the vicinity of Montague Island indicate that the shortening resulting from this compression may be substantially greater than that indicated by the surface ruptures. Survey points less than 15 kilometers apart on southern Montague Island and the islands immediately to the north and northwest have apparently been displaced horizontally toward one another by as much as 5 meters (*11*).

Both the Patton Bay and the Hanning Bay faults on Montague Island occurred along prominent linear breaks in slope or linear stream valleys clearly visible on aerial photographs taken before the earthquake. One of these, the Patton Bay Fault, had been delineated on a photogeological map of part of Prince William Sound made prior to the earthquake (*17*). Although the vertical displacements that occurred along these two faults during this earthquake are large, our reconnaissance observations did not reveal any significant lithologic differences in the rock sequences on the two sides of the faults, or wide zones of brecciated rock along them, such as commonly occur along faults that form major tectonic boundaries.

There are a number of abrupt submarine scarps in the vicinity of Montague Island and Hinchinbrook Island, which lies immediately northeast of Montague Island, and on the continental shelf between Montague and Kodiak islands (*18*). Some of these may be the surface expression of faults along which dip-slip displacement occurred during the earthquake.

Seismological Data

Large earthquakes at shallow and intermediate depths are thought by most North American geologists and seismologists to result from the sudden rupture of strained rocks (*19*). According to the elastic rebound theory, the energy released in seismic waves and in other ways is derived from accumulated elastic-strain energy in deformed blocks of rock as they snap back toward equilibrium on either side of a fracture or fault (*20*). The instrumental epicenter of the main shock marks the surface projection of the point at which the rupture begins.

The epicenter of the Alaskan earthquake, which has a Richter magnitude variously estimated as 8.4 (Pasadena seismograph station) to 8.5 (U.S. Coast and Geodetic Survey), is located (with an uncertainty of 0.2 degree, or a radius of error of 12 km) on the east shore of Unakwik Inlet in northern Prince William Sound at latitude 66.1°N, longitude 147.7°W (*21*). The hypocenter (focal depth) could not be determined more closely than in the range between 20 and 50 kilometers.

Aftershocks which follow large earthquakes are thought to be generated by continuous adjustments of the strained volume of rock, or focal region, within which the rupture occurred (*22, 23*). Thus, the spatial distribution of the aftershocks approximately delimit the focal region of the main shock in all three dimensions, even where there is no surface breakage.

The horizontal extent of the focal region is roughly delineated by the areal distribution of 83 percent of the first 132 large aftershocks (shocks of magnitude greater than 5.0, as determined by the Coast and Geodetic Survey), which account for most of the release of elastic-strain energy during the aftershock sequence. The aftershocks lie in a well-defined belt, 100 to 200 kilometers wide and about 800 kilometers long, that roughly parallels the trend of the Aleutian Trench and includes the known area of uplift and the adjacent continental shelf and parts of the continental slope. The most intense aftershock activity is concentrated toward the northeastern and southeastern ends. The northwestern limits of this belt are close to the boundary between the major areas of uplift and subsidence. The only part of the belt of major aftershocks that lies within the zone of subsidence is a small area immediately north of the epicenter; seven aftershocks occurred in this area. Major aftershocks that lie outside the belt shown in Fig. 1 were widely distributed beneath the Kenai-Kodiak Mountains, along the west shore of Cook Inlet, and seaward from the Aleutian Trench. The more numerous smaller aftershocks follow the same general pattern of distribution, although they are spread over a somewhat larger area.

Fig. 8. Aerial photograph, looking northeastward, showing the line of landslides along the trace of the Patton Bay Fault. Patton Bay is in the background.

The vertical extent of the inferred focal region is less perfectly defined because of inherent errors in the determination of focal depths in areas of uncertain crustal structure and seismic velocity. Hypocenters of the major aftershocks occur at depths between 5 and 40 kilometers, averaging about 20 kilometers (*21*). They show a general tendency toward deepening of their lower limit beneath the continent; the deeper large aftershocks occur at depths of 30 to 40 kilometers and are situated, in general, approximately along the axis of the Chugach-Kenai Mountains (Fig. 2). There is no clear indication of a regular increase in depth of the upper limit of aftershocks within the zone of uplift; this presumably means that fractures were occurring throughout the thickness of the continental crust.

According to a theory advanced by Benioff, based on both experimental and observational data, the characteristics of the strain-release pattern of the aftershock sequence provide a means of distinguishing shear from compressional strain (*23, 24*). The strain-release curve derived from the aftershock sequence of the earthquake of 27 March is of the form that indicates dominantly compressional deformation within the focal region (*21*), in accord with the observed pattern of surface warping and faulting.

The initial direction of rupture at an earthquake focus may be derived from the world-wide distribution of initial compressions and dilatations of seismic waves recorded at seismograph stations, an elastic-rebound fault source being assumed (*25*). Solutions based upon compressional waves, or P-waves, define a pair of orthogonal planes at the focus, one of which presumably contains the active fault surface. Theoretically, an unambiguous solution for the fault plane may be obtained by analysis of shear waves and surface waves in addition to the P-waves, although the records of these phases may be considerably more difficult to interpret than those of the P-waves. Inherent in the fault-plane solutions is the basic assumption made by earthquake seismologists that the initial displacement, at least in the larger earthquakes, reflects the regional stress field.

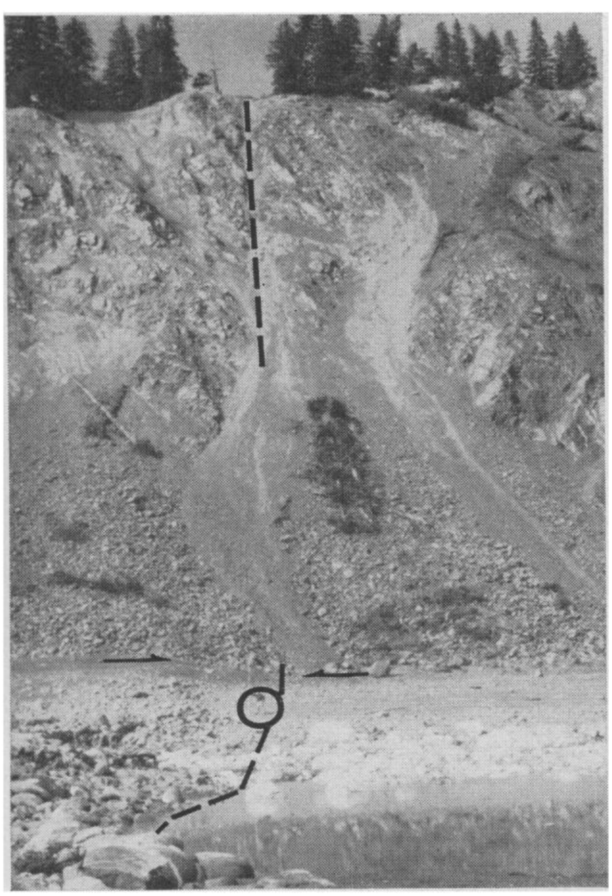
A fault-plane solution based on P-waves, made by the Seismological Division of the U.S. Coast and Geodetic Survey, yields two planes that strike N64°E and dip about 82° SE and 8°NW through the focus (*21*). Another solution, by Berg, of the seismological laboratory of the University of Alaska, yields one well-defined plane that strikes N72°E, with an almost vertical dip, and a poorly controlled second plane with a low-angle dip

(*26*). If the steep plane is taken as the fault plane, the solutions indicate predominantly dip-slip movement, with the southeast side relatively upthrown and a left-lateral strike-slip component. The alternative low-angle plane would yield a thrust fault with the northwest block upthrown. The near-vertical fault is indicated by a study that was made of the surface-wave spectra (*27*). As Berg points out, however (*26*), "the other possibility of a low angle thrust can not be disregarded, and in fact, was strongly suggested in one of the aftershock solutions, which in other respects was very similar to the main shock." It is significant that, regardless of the orientation of the primary faulting, the preliminary solutions indicate that the major stress axis is oriented normal to the structural trend of the Aleutian Arc, in agreement with the general trend suggested by mechanism studies of previous large earthquakes elsewhere along the Arc (*28*).

Geologic Record of Pre-Earthquake Deformation

The tectonic movements that occurred on 27 March were the most recent pulse in an episode of deformation that began in south-central Alaska during late Cenozoic time and has continued intermittently ever since. The available geologic evidence reveals a history of complex postglacial vertical displacements relative to sea level, in which areas of net uplift or subsidence appear to correspond in general with areas in which uplift and subsidence occurred during the 27 March earthquake.

The structure of strata of Tertiary age along the coast of the Gulf of Alaska within the zone of uplift and the adjacent area to the east is dominated by asymmetric folds and north-dipping thrust faults that strike roughly parallel to the coastline (*29*). Late Cenozoic diastrophism along this same trend is evidenced by 30-degree dips in marine strata of late Pliocene and early Pleistocene age on Middleton Island, well out on the continental shelf. Postglacial deformation has left multiple raised beaches and terraces along the southeast coast of Kodiak Island, in Prince William Sound, along the mainland to the east of the sound, on Kayak Island and adjacent islands, and on Middleton Island. Recent spasmodic vertical displacements with net uplift of the continental-shelf margin relative to sea level are recorded in five steplike marine terraces to an elevation of 30 meters on Middleton Island (*30*). Radiocarbon dating of driftwood from the highest terrace (*31*) demonstrates that 30 meters of net uplift relative to sea level has occurred in the past 4470 ± 250 years. The actual amount of uplift is slightly greater than 30 meters because sea level was also rising to its present level until about 3500 years ago (*32*). Within most of Prince William Sound and the Controller Bay area to the



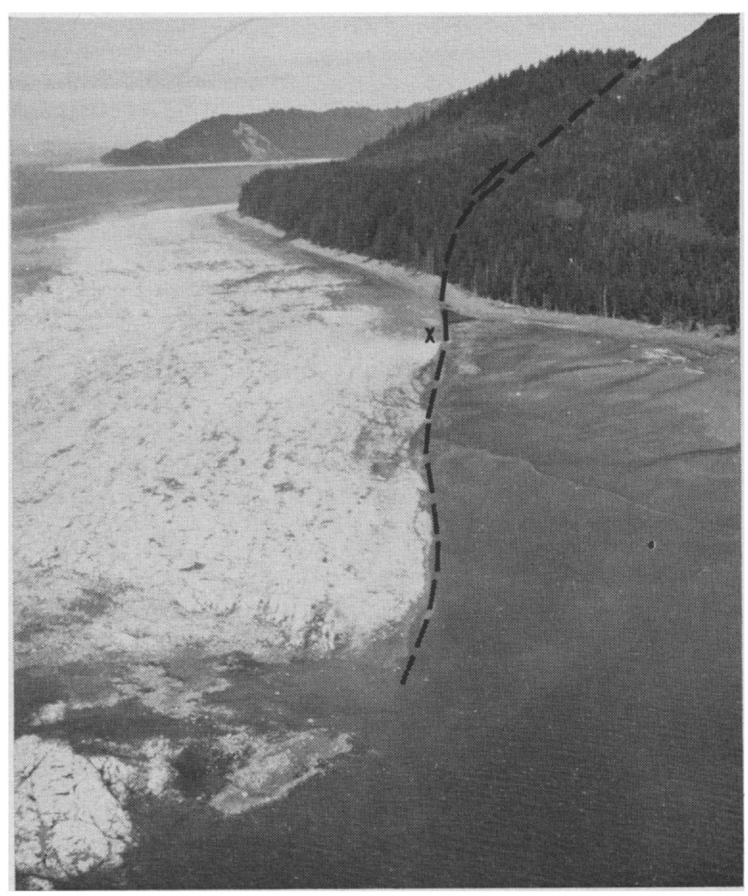


Fig. 9 (left). Vertical shear zone in sea cliff, and offset along the beach where the Patton Bay Fault intersects the shoreline near Jeanie Point. The approximate trace of the principal fault break is shown by the dashed line, and the vertical offset in talus at the base of the cliff is indicated by arrows. The figure of a man standing at the base of a 3-meter-high scarp is circled. Dip of bedding is 45 degrees to the southwest (right). Fig. 10 (right). View to the northeast along the Hanning Bay Fault. The northwest block (left) has been relatively upthrown 4 to 5 meters along a high-angle reverse fault. The white coating on the reef rock of the upthrown block is the bleached remains of calcareous algae and bryozoans that lived below mean tide level. The *X* marks the location of Fig. 11.

east, deformation immediately prior to the earthquake was a general subsidence relative to sea level, as shown by a fringe of drowned forests and intertidal peat bogs along the shores of the mainland and the islands of eastern and southern Prince William Sound (*33*).

In contrast to the complex structural deformation and uplift of the Cenozoic strata along the coast, Tertiary sediments in the Cook Inlet region within the zone of subsidence and along the Aleutian Volcanic Arc to the southwest are flat-lying or only mildly deformed (*29, 34*). During postglacial time that part of the zone of subsidence which is within the Cook Inlet area has been a region of relative crustal stability (*35*). Postglacial high-angle reverse-fault movement is reported at one point along the Castle Mountain–Lake Clark fault system, which extends for nearly 320 kilometers along the northwestern and northern boundary of the Cook Inlet basin (*36*). Local subsidence of the land relative to sea level along the axis of the Kenai and Kodiak mountains is indicated by Wisconsin-age cirques that are now well below sea level along the south shore of the Kenai Peninsula.

Surface deformation has been recorded in Alaska after three previous major earthquakes that occurred along the coast of the Gulf of Alaska. In 1889 and 1958 displacement occurred along segments of the Chugach-St. Elias-Fairweather fault system, which roughly parallels the south coast of Alaska (*37*). The maximum known vertical displacement associated with the 1899 earthquake at Yakutat Bay (Fig. 1) was 14.3 meters of uplift on bedrock and as much as 2 meters of subsidence in unconsolidated deposits. There was also an unknown amount of oblique-slip movement during this earthquake at one locality in Nunatak Fiord (*8*). Measured displacement along the Fairweather Fault at a point 200 kilometers southeast of Yakutat Bay after the 1958 earthquake showed a predominant right-lateral strike-slip offset of as much as 6½ meters, with a slight upward movement of the southern block (*38*). A large earthquake in 1880 on Chirikof Island, roughly 160 kilometers southwest of Kodiak Island (Fig. 1), was accompanied by vertical fault displacement of 1.8 meters (*39*). In my opinion the limited historic record of faulting on Chirikof Island, on Montague Island, and possibly at Yakutat Bay is consistent with the geologic record, which suggests late Cenozoic movement that is predominantly dip-slip along the coast of south-central Alaska. Significant strike-slip displacement along faults parallel to the coast of the Gulf of Alaska during late Cenozoic time has not yet been recognized west of Yakutat Bay, although it is possible that such evidence may have been overlooked or underestimated in the geological mapping of these areas.

Fig. 11. The Hanning Bay Fault scarp, looking northeast. Vertical displacement in the foreground, in rock, is about 4 meters; the maximum measured displacement of 5 meters is at the beach ridge in the trees.

Speculation on Origin of the Earthquake

A characteristic of the earthquake belt associated with the Aleutian Arc, and of others associated with island arcs throughout the world, is that the earthquake epicenters lie on the concave, generally continental side of the associated oceanic trench and the hypocenters deepen toward the adjacent continent (*40*). As stated elsewhere (*1*), "this distribution of earthquakes in the Aleutian Arc and Trench is believed by many geologists and geophysicists to indicate that the earthquakes originate in a fault or perhaps a zone of movement which extends with a moderate northward dip from the Aleutian Trench northward beneath the Aleutian Arc." That the earthquake of 27 March occurred along the postulated zone of faulting is strongly suggested by the occurrence of its belt of seismic activity and surface deformation within, and parallel to, the Aleutian Arc and Trench. The orientation and sense of displacement on this postulated fault or zone of faulting, however, can only be deduced indirectly from the seismological data, the residual surface displacements, and the geologic record of past deformation.

If it is assumed that the earthquake originated by rupture along one or more faults, and that the uplift and subsidence resulted from elastic rebound, it is important to consider the possible orientation and sense of movement that might best explain the observed pattern of surface displacement. The two most plausible alternatives consistent with the available fault-plane solutions are (i) dip-slip movement on a near-vertical fault between the major zones of uplift and subsidence, and (ii) thrusting along a fault or zone of faulting that dips northwestward from the Aleutian Trench beneath the Aleutian Arc.

According to the first hypothesis, a near-vertical fault strikes along the line of zero change in land level, with the southeast block up and the northwest block down relative to sea level. The greatest attraction of this hypothesis is that it provides a simple model to account for the distribution of surface uplift and subsidence in the two major zones. It is supported by the occurrence of the earthquake epicenter close to the zero line between the two zones, and by a preliminary unambiguous fault-plane solution based on surface waves (*27*).

Upon detailed examination, however, the hypothesis appears to pose more problems than it answers. Most serious among these are the absence of surface fault displacement at or in the vicinity of the zero line and the lack of evidence that the line corresponds to major geologic boundaries, as might be expected if it marked the trace of a major fault along which vertical movement has occurred in the past. An alternative possibility is that the displacement represents flexure above a fault at depth. In an elegant analysis of the displacements by application of dislocation theory, Press and Jackson (*27*) have shown that the observed vertical surface displacements of less than 6 meters could be accounted for by a steeply dipping subsurface fault extending from 15 kilometers below the surface to the considerable depth of 100 to 200 kilometers. There are four serious objections to this conclusion, however. (i) It seems improbable that a near-vertical fault 800 kilometers long and from 85 to 185 kilometers deep, which generated one of the greatest earthquakes in history, should fail to reach the surface anywhere along its length or show detectable surface offset, especially since 5 meters of vertical offset occurred at the surface along faults on Montague Island which are presumed to be subsidiary ruptures. (ii) The postulated fault is more than twice the depth of the hypocenters of the initial shock or of the deepest large aftershocks of this earthquake or of previously recorded major earthquakes in the same general region (*40*). (iii) The focal region, as inferred from aftershock distribution, is a belt 160 to 320 kilometers wide that lies mainly to the south of the postulated fault rather than being more symmetrically disposed with respect to its surface trace, as might be expected for a steeply dipping fault (Fig. 1). (iv) The sense of displacement on the postulated fault is opposite to the up-to-the-north movement of known faults that strike parallel to the regional structural trend and the surface faulting that occurred on Montague Island during the earthquake.

The second hypothesis, which proposes that the earthquake originated along a fault or zone of movement that extends northward from the Aleutian Trench, seems to be more promising, perhaps because it postulates a primary fault that is safely concealed from view beneath the sea. According to this model, the zone of fault slippage is within the belt of major aftershocks (Figs. 1 and 6). Dominantly compressive stress, oriented approximately normal to the Arc, is indicated by the pattern of folds and faults in rocks of late Cenozoic age as well as by the pattern of surface deformation that accompanied the earthquake and the strain-release characteristics of the aftershock sequence. A downward vertical component of pre-earthquake strain is also suggested by the fringe of drowned forest along much of the coast in the northeastern part of the uplifted zone. The postulated stress pattern could result from progressive underthrusting of the oceanic crust and mantle beneath the continental margin, as illustrated diagrammatically in Fig. 6. The crustal configuration shown in Fig. 6 is largely speculative, for crustal thicknesses in this part of Alaska have been determined only along short seismic-refraction lines in the Kodiak Island area and in northern Prince William Sound (*41*). Elastic rebound during the earthquake resulted in relative seaward thrusting of the continental margin along one or more primary northward-dipping faults (not shown in Fig. 6) with accompanying uplift and warping of the upper continental block. The surface faults on Montague Island and their seaward extensions are situated in the zone of maximum uplift, where renewed movement occurred along preexisting vertical or high-angle reverse faults. Comparable subsidiary faults could well occur elsewhere on the continental shelf.

The most serious limitation of the thrust-fault hypothesis lies in its attempts to account for the observed subsidence to the north of the zone of uplift. The relative scarcity of large aftershocks within the zone of subsidence, except in the immediate vicinity of the epicenter of the initial shock, is interpreted as indicating that this zone was largely outside the area of the primary fault rupture along which the earthquake occurred. In the absence of surface-fault displacement, it is tentatively suggested that the subsidence may be a secondary effect resulting from elastic deformation immediately adjacent to the postulated zone of thrusting. Preliminary results of resurveys of small portions of the triangulation net within the zone of subsidence near Anchorage show as much as 2⅔ meters of horizontal elongation within a north-south distance of 48 kilometers (*11*), indicating that crustal extension could have been a significant factor in producing the observed subsidence.

The hypothesis outlined above is generally consistent with most modern theories which relate arc structures in the circum-Pacific region and elsewhere to master thrust faults along the unstable interface between the oceanic and continental crusts. It can also account for the following observed features: (i) the areal distribution of the major zones of uplift and subsidence; (ii) the marked asymmetry in the volumes of uplift and subsidence in the two major zones; (iii) vertical or reverse surface faults in the zone of uplift with up-to-the-north displacement; (iv) occurrence of the belt of major aftershocks mainly within the zone of uplift and its inferred offshore extension; (v) the shallow depths of the initial shock and aftershocks and the tendency toward deepening of aftershock hypocenters beneath the continent; (vi) the geologic record of late Cenozoic folding, reverse faulting, and net uplift of the continental margin and shelf, in contrast to the history of relative stability or slight subsidence in the adjacent area to the north.

Neither of the two hypotheses outlined above attempts to account for the possibility that there may be a second zone of slight uplift adjacent to the zone of subsidence. Additional field investigations planned for the 1965 season by the U.S. Geological Survey and the U.S. Coast and Geodetic Survey are aimed at resolving this problem, and at filling gaps in the existing picture of horizontal and vertical surface deformations that accompanied the earthquake.

Summary

Alaska's Good Friday earthquake of 27 March 1964 was accompanied by vertical tectonic deformation over an area of 170,000 to 200,000 square kilometers in south-central Alaska. The deformation included two major northeast-trending zones of uplift and subsidence situated between the Aleu-

tian Trench and the Aleutian Volcanic Arc; together they are 700 to 800 kilometers long and from 150 to 250 kilometers wide. The seaward zone is one in which uplift of as much as 10 meters on land and 15 meters on the sea floor has occurred as a result of both crustal warping and local faulting. Submarine uplift within this zone generated a train of seismic sea waves with half-wave amplitudes of more than 7 meters along the coast near the source. The adjacent zone to the northwest is one of subsidence that averages about 1 meter and attains a measured maximum of 2.3 meters. A second zone of slight uplift may exist along all or part of the Aleutian and Alaska ranges northwest of the zone of subsidence.

The studies made to date demonstrate that great earthquakes such as the earthquake of 27 March may be accompanied by regional deformation on a larger scale than has been generally recognized. Perhaps better than any previous seismic event, this earthquake indicates that vertical displacement of the sea bottom can generate destructive seismic sea waves, even where the epicenter of the main shock is as much as 100 kilometers inland from the coast.

The focal region of the earthquake, as inferred from the spatial distribution of the major aftershocks, lies almost entirely within the seaward zone of uplift and extends from close to the surface to a depth of about 50 kilometers. The primary fault or zone of faulting along which the earthquake presumably occurred is not exposed at the surface on land. The only known surface breakage is along two preexisting faults on Montague Island, within the area of maximum uplift, that trend northeast and are near-vertical or dip steeply northwest. The displacement, which was subsidiary to the regional uplift, was essentially dip-slip, with the northwest blocks relatively upthrown. The maximum measured vertical displacement on land is 5 meters, and along the submarine extension of one of these faults the vertical displacement may exceed 10 meters.

It is postulated that the earthquake is genetically related to the Aleutian Arc and probably resulted from regional compressive stress oriented roughly normal to the Arc. Neither the orientation nor the sense of movement on the primary fault along which the earthquake occurred is known. Available fault-plane solutions based on P-waves indicate that the primary fault could be either a northwest-dipping thrust or a northeast-striking near-vertical fault with the southeast side upthrown. Which, if either, of these alternatives represents the primary fault cannot be determined without additional field data on the horizontal and vertical displacements that accompanied the earthquake and detailed analyses of the innumerable seismographic records written by the earthquake and its aftershocks.

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