The Alaska Earthquake
March 27, 1964

Regional Effects

Tectonics

GEOLOGICAL SURVEY PROFESSIONAL PAPER 543-I

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TECTONICS OF THE
MARCH 27, 1964
ALASKA EARTHQUAKE

Former sea floor at Cape Cleare, Montague Island, Prince William Sound, exposed by 26 feet of tectonic uplift. The surf-cut surface, which slopes gently from the base of the sea cliffs to the water, is about a quarter of a mile wide. The white coating on the rocky surface consists mainly of the desiccated remains of calcareous algae and bryozoans. Photograph taken at about zero tide stage, May 30, 1964.

THE ALASKA EARTHQUAKE, MARCH 27, 1964:

REGIONAL EFFECTS

Tectonics of the March 27, 1964 Alaska Earthquake

By GEORGE PLAFKER

"But it is time that the geologist should, in some degree, overcome those first and natural impressions which induced the poets of old to select the rock as the emblem of firmness—the sea as the image of inconstancy."

(*Lyell, 1874, p. 179*)

GEOLOGICAL SURVEY PROFESSIONAL PAPER 543-I

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THE ALASKA EARTHQUAKE SERIES

The U.S. Geological Survey is publishing the results of its investigations of the Alaska earthquake of March 27, 1964, in a series of six Professional Papers. Professional Paper 543 describes the regional effects of the earthquake. Eight chapters in this volume have already been published. Other Professional Papers in the series describe field investigations and reconstruction and the effects of the earthquake on communities, on the hydrologic regimen, and on transportation, utilities, and communications. A selected bibliography and an index for the entire series will be published in the concluding volume, Profession Paper 546.

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TECTONICS OF THE MARCH 27, 1964, ALASKA EARTHQUAKE

By George Plafker

ABSTRACT

The March 27, 1964, earthquake was accompanied by crustal deformation—including warping, horizontal distortion, and faulting—over probably more than 110,000 square miles of land and sea bottom in south-central Alaska. Regional uplift and subsidence occurred mainly in two nearly parallel elongate zones, together about 600 miles long and as much as 250 miles wide, that lie along the continental margin. From the earthquake epicenter in northern Prince William Sound, the deformation extends eastward 190 miles almost to long 142° and southwestward slightly more than 400 miles to about long 155°. It extends across the two zones from the chain of active volcanoes in the Aleutian Range and Wrangell Mountains probably to the Aleutian Trench axis.

Uplift that averages 6 feet over broad areas occurred mainly along the coast of the Gulf of Alaska, on the adjacent Continental Shelf, and probably on the continental slope. This uplift attained a measured maximum on land of 38 feet in a northwest-trending narrow belt less than 10 miles wide that is exposed on Montague Island in southwestern Prince William Sound. Two earthquake faults exposed on Montague Island are subsidiary northwest-dipping reverse faults along which the northwest blocks were relatively displaced a maximum of 26 feet, and both blocks were upthrown relative to sea level. From Montague Island, the faults and related belt of maximum uplift may extend southwestward on the Continental Shelf to the vicinity of the Kodiak group of islands. To the north and northwest of the zone of uplift, subsidence forms a broad asymmetrical downwarp centered over the Kodiak-Kenai-Chugach Mountains that averages 2½ feet and attains a measured maximum of 7½ feet along the southwest coast of the Kenai Peninsula. Maximum indicated uplift in the Alaska and Aleutian Ranges to the north of the zone of subsidence was 1½ feet. Retriangulation over roughly 25,000 square miles of the deformed region in and around Prince William Sound shows that vertical movements there were accompanied by horizontal distortion, involving systematic shifts of about 64 feet in a relative seaward direction. Comparable horizontal movements are presumed to have affected those parts of the major zones of uplift and subsidence for which retriangulation data are unavailable.

Regional vertical deformation generated a train of destructive long-period seismic sea waves in the Gulf of Alaska as well as unique atmospheric and ionospheric disturbances that were recorded at points far distant from Alaska. Warping resulted in permanent tilt of larger lake basins and temporary reductions in discharge of some major rivers. Uplift and subsidence relative to sea level caused profound modifications in shoreline morphology with attendant catastrophic effects on the nearshore biota and costly damage to coastal installations. Systematic horizontal movements of the land relative to bodies of confined or semiconfined water may have caused unexplained short-period waves—some of which were highly destructive—observed during or immediately after the earthquake at certain coastal localities and in Kenai Lake. Porosity increases, probably related to horizontal displacements in the zone of subsidence, were reflected in lowered well-water levels and in losses of surface water.

The primary fault, or zone of faults, along which the earthquake occurred is not exposed at the surface on land. Focal-mechanism studies, when considered in conjunction with the pattern of deformation and seismicity, suggest that it was a complex thrust fault (megathrust) dipping at a gentle angle beneath the continental margin from the vicinity of the Aleutian Trench. Movement on the megathrust was accompanied by subsidiary reverse faulting, and perhaps wrench faulting, within the upper plate. Aftershock distribution suggests movement on a segment of the megathrust, some 550–600 miles long and 110–180 miles wide, that underlies most of the major zone of uplift and the seaward part of the major zone of subsidence.

According to the postulated model, the observed and inferred tectonic displacements that accompanied the earthquake resulted primarily from (1) relative seaward displacement and uplift of the seaward part of the block by movement along the dipping megathrust and subsidiary faults that break through the upper plate to the surface, and (2) simultaneous elastic horizontal extension and vertical attenuation (subsidence) of the crustal slab behind the upper plate. Slight uplift inland from the major zones of deformation presumably was related to elastic strain changes resulting from the overthrusting; however, the data are insufficient to permit conclusions regarding its cause.

The belt of seismic activity and major zones of tectonic deformation associated with the 1964 earthquake, to a large extent, lie between and parallel to the Aleutian Volcanic Arc and the Aleutian Trench, and are probably genetically related to the arc. Geologic data indicate that the earthquake-related tectonic movements were but the most recent pulse in an episode of deformation that probably began in late Pliocene time and has continued intermittently to the present. Evidence for progressive coastal submergence in the deformed region for several centuries preceding the earthquake, in combination with transverse horizontal shortening indicated by the retriangulation data, suggests pre-earthquake strain directed at a gentle angle downward beneath the arc. The duration of strain accumulation in the epicentral region, as interpreted from the time interval during which the coastal submergence occurred, probably is 930–1,360 years.

INTRODUCTION

Among the most notable aspects of the 1964 Alaska earthquake was the great areal extent and amount of the tectonic movements that accompanied it. From the epicenter in northern Prince William Sound, the region affected by tectonic deformation parallels the trends of the Aleutian Volcanic Arc, the Aleutian Trench, and the Gulf of Alaska coast for about 600 miles (fig. 1). In south-central Alaska, where the northeastern end of the arc intersects the continent at an oblique angle, the pattern of deformation can be observed in an exceptionally complete profile extending more than 200 miles northward from the seaward edge of the Continental Shelf across the northeastern end of the volcanic arc. Within this region, tectonic displacements on land include absolute vertical movements ranging from as much as 38 feet of uplift to 7½ feet of subsidence, relative horizontal movements of about 64 feet, and dip-slip offset on reverse faults of as much as 26 feet. Furthermore, the available data indicate that these movements extended over a large segment of the adjacent offshore area where they may have been as large, or larger, than those measured on land.

This report substantially enlarges upon a preliminary summary and interpretation of the tectonic movements that accompanied the 1964 earthquake (Plafker, 1965). It presents the available data on the distribution and nature of the earthquake-related displacements and of the manifold, often disastrous, effects of these movements. Geologic, geodetic, and seismologic data pertinent to the tectonics that were available to the writer prior to completion of this report in July 1967 are summarized. Implications of the data for the earthquake mechanism are reviewed, and a tentative qualitative model is outlined, which attempts to explain most of the observations.

ACKNOWLEDGMENTS

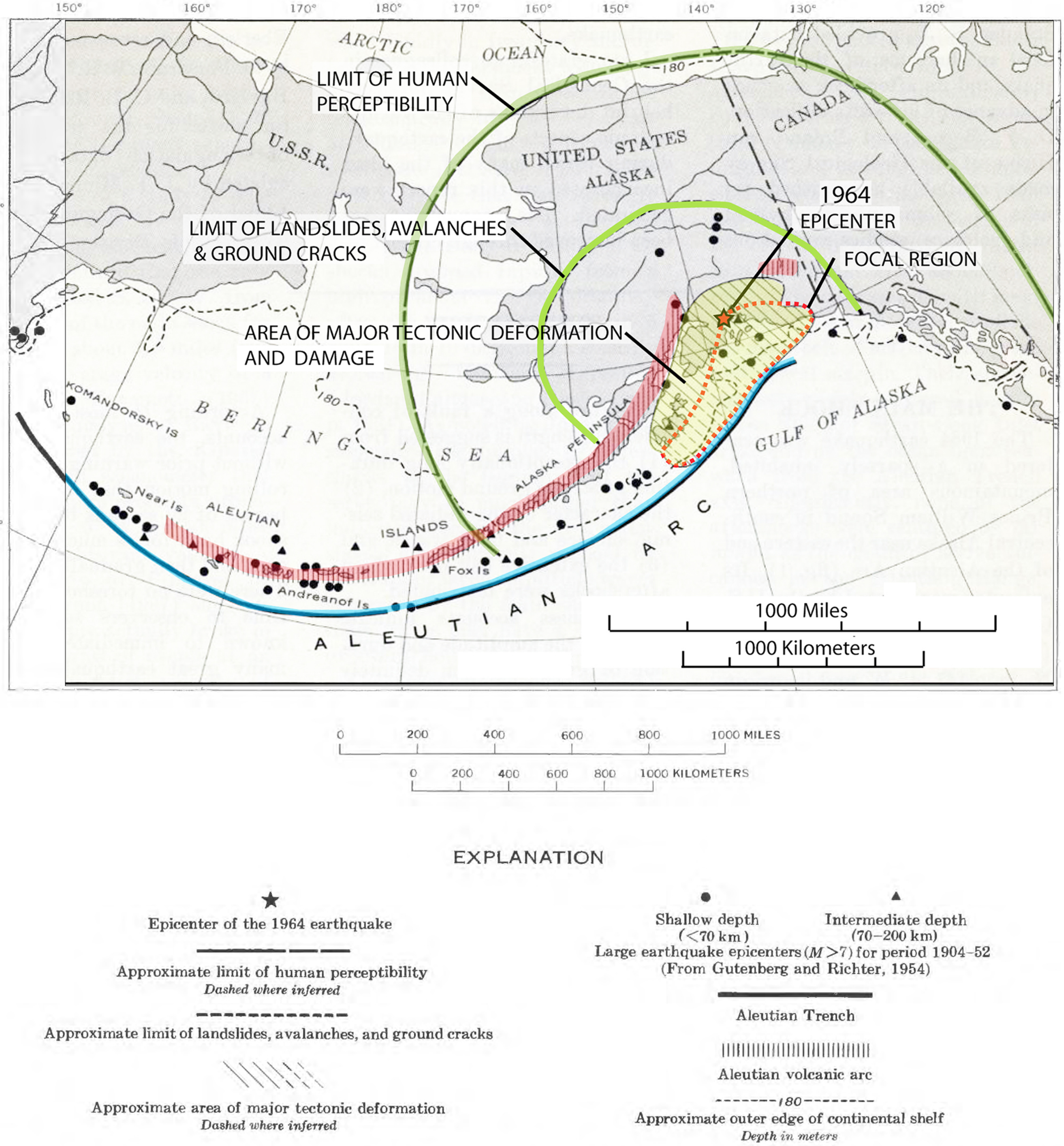
Field mapping of the vertical shoreline displacements and surface faults was accomplished by a party headed by the writer from mid-May through August 1964 and during 1 month in 1965. Geological Survey personnel included L. R. Mayo, J. B. Case, S. L. Robbins, and William Bastian during 1964, and Mayo and M. G. Bonilla during 1965. An especially large part of the fieldwork and data compilation were carried out by L. R. Mayo during both field seasons. G Dallas Hanna, marine biologist of the California Academy of Sciences, spent 3 weeks with the party in Prince William Sound in 1964, during which time he studied effects of the earthquake on the ecology of the intertidal fauna and flora and provided invaluable advice on the use of sessile marine organisms for determining changes in land level.

The U.S. Geological Survey research vessel, *Don J. Miller*, was used as a base of operations for work in Prince William Sound and Resurrection Bay during the 1964 season. Helicopters and fixed-wing aircraft were used during both field seasons for work along shorelines elsewhere. Outstanding logistical support of the field investigations was provided—often under difficult circumstances—by the crew of the *Don J. Miller*, which consisted of Capt. John Stacey and cook-seaman John Muttart of the U.S. Geological Survey, and by bush pilots Jim Osborne, Glenn Wheeler, "Stinky" Myers, Al Cratty, Bob Leonard, Oren Hudson, and Bob Barnett.

R. L. Detterman, Reuben Kachadoorian, T. N. V. Karlstrom, G. W. Moore, and B. L. Reed, all of the Geological Survey, contributed data on changes in land level as determined by coastal residents of Cook Inlet and the Kodiak group of islands. Many residents of Alaska gave helpful information on earthquake effects, including tectonic changes, in about 150 interviews with the writer and on numerous form questionnaires.

Virtually all the seismologic, geodetic, and marine data incorporated in this report were obtained by the U.S. Coast and Geodetic Survey as part of their massive investigations of this major seismic event. I am especially indebted to the numerous individuals in the U.S. Coast and Geodetic Survey who freely made available unpublished data and have discussed their interpretations of these data with the writer on many occasions. Among these, special thanks for assistance are due S. T. Algermissen of the Seismology Division, W. D. Barbee, R. J. Malloy, and E. W. Richards of the Office of Oceanography, and C. A. Whitten, E. J. Parkin and J. B. Small of the Geodesy Division.

Postearthquake vertical aerial photographs were provided through the courtesy of Col. M. L. Falwell, U.S. Army, Fort Richardson, and H. R. Cravat of the Photogrammetry Division, U.S. Coast and Geodetic Survey. R. A. Page, Jr., of the Lamont Geological Observatory furnished a computer plot of the larger magnitude



1.—Map of Alaska and adjacent areas showing the location of the 1964 earthquake, the area affected by the earthquake, epicenters of previous major earthquakes, belts of active volcanism, and the Aleutian Trench.

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aftershock epicenters. William Stauder, S. J., provided data on focal mechanisms of the earthquake and its aftershock sequence in advance of his own publication. D. F. Barnes and Roland von Huene of the Geological Survey made available unpublished results of submarine geophysical and geological studies carried out in south-central Alaska after the earthquake.

I am grateful to colleagues in the Geological Survey for very helpful discussions of the various tectonic aspects of the earthquake during which many of the ideas incorporated in this report were generated. The manuscript has been improved through critical reviews by E. B. Eckel and G. D. Eberlein and reviews of portions by D. F. Barnes, R. R. Coats, R. O. Burford, and C. B. Raleigh. Vector shifts for the readjustment of triangulation data were calculated by J. T. Minta and C. R. Lloyd of the Geological Survey's Topographic Division.

SEISMICITY

THE MAIN SHOCK

The 1964 earthquake was centered in a sparsely inhabited, mountainous area of northern Prince William Sound in south-central Alaska near the eastern end of the Aleutian Arc (fig. 1). Its epicenter was located by the U.S. Coast and Geodetic Survey (Wood, 1966, p. 62) at lat 61.06° N., long 147.44° W. and its origin time was at about 5:36 p.m., Friday, March 27, 1964, A.s.t. (03:36:13.5, Saturday, March 28, 1964, G.m.t.). The hypocenter, or focus, could not be determined more closely than between 12 and 31 miles (20–50 km) in depth.

Magnitude of the earthquake, based upon surface-wave amplitudes (M_s) is estimated to have been about 8.4 (Pasadena seismograph station). Earthquake vibrations were felt over an area in excess of a million square miles in Alaska and adjacent parts of Canada (fig. 1), and they caused widespread damage throughout an area of about 50,000 square miles in south-central Alaska. The manifold effects of the shaking, which have been described in numerous reports are concisely summarized by Hansen and Eckel (1966) in the introductory volume of this series of reports.

Rupture along a fault of considerable length is suggested from (1) the exceptionally long duration of strong ground motion, (2) the character of the radiated seismic surface and body waves, and (3) the extensive belt over which aftershocks were distributed.

Eyewitness accounts indicate that both the amplitude and duration of ground motion definitely tended to be largest in areas of relatively unconsolidated saturated deposits and least in areas of crystalline or metamorphic rocks. Within the immediate area affected by the earthquake, the only known instrumental records of the duration of shaking were made on several automatic recording charts in a steam powerplant built on bedrock at Whittier. On the clearest of these records, trace vibrations lasted for nearly 4 minutes. At other bedrock sites on Kodiak Island and on the Kenai Peninsula, where the motion was timed by observers with pocket watches or clocks, it was $2\frac{1}{2}$–5 minutes. Anomalously short durations, ranging from only 15 seconds to $1\frac{1}{2}$ minutes, were reported by residents at Seldovia and in two nearby localities in areas underlain by metamorphic rock at the southwestern tip of the Kenai Mountains (Plafker, Kachadoorian, Eckel, and Mayo, 1969).

According to most eyewitness accounts, the earthquake started without prior warning as a gentle rolling motion that lasted for a period of 20 seconds to 1 minute, shook hard for as much as 4 minutes, and then gradually subsided. There were no foreshocks perceptible to observers such as are known to immediately precede many great earthquakes. A few observers heard premonitory low rumbling sounds several seconds before the earthquake was felt. The ground motion was variously described as a rolling wavelike motion, a strong horizontal acceleration, or a hard jarring motion. A few individuals noted that the ground motion during the earthquake periodically eased up for a short period and then resumed with increased violence.

Reports of the directions of vibrations vary widely, and at some places, prevailing vibration directions reportedly changed during the earthquake. The majority of the reports indicate a tendency for ground oscillation to be grouped in the quadrants between northwest-southeast and northeast-southwest although it also was reported from all other quadrants. A westward propagation of vibrations between Anchorage and Kenai is indicated by the fact that power failure due to shaking in the Anchorage area caused an overload of the circuits at the interconnected Bernice Lake powerplant near Kenai, 53 miles farther from the epicenter, some 15–20 seconds before the plant superintendent felt the tremors.

On the basis of a study of seismic surface waves, Toksöz, Ben-Menahem, and Harkrider (1965, p. 154) determined that the rupture propagated S. 50° W. from the epicenter of the main shock for a distance of about 370 miles (600 km) at an average velocity of 3 km per sec. Furumoto (1967) found from analysis of Rayleigh waves recorded by the Kipapa, Hawaii, strain seismometer that the rupture more probably propagated S. 30 ± 5° W. for 500 miles (800 km) at this same velocity. A detailed study of *P*-phases by Wyss and Brune (1967) suggests that the rupture actually broke in a complex series of events at an average propagation velocity of 3.5 km per sec and that each event was characterized by more or less distinct high-amplitude bursts or events. Their data further indicate that, although the rupture propagated initially in various azimuthal directions, after an elapsed time of 44 seconds it continued only in a southwesterward direction.

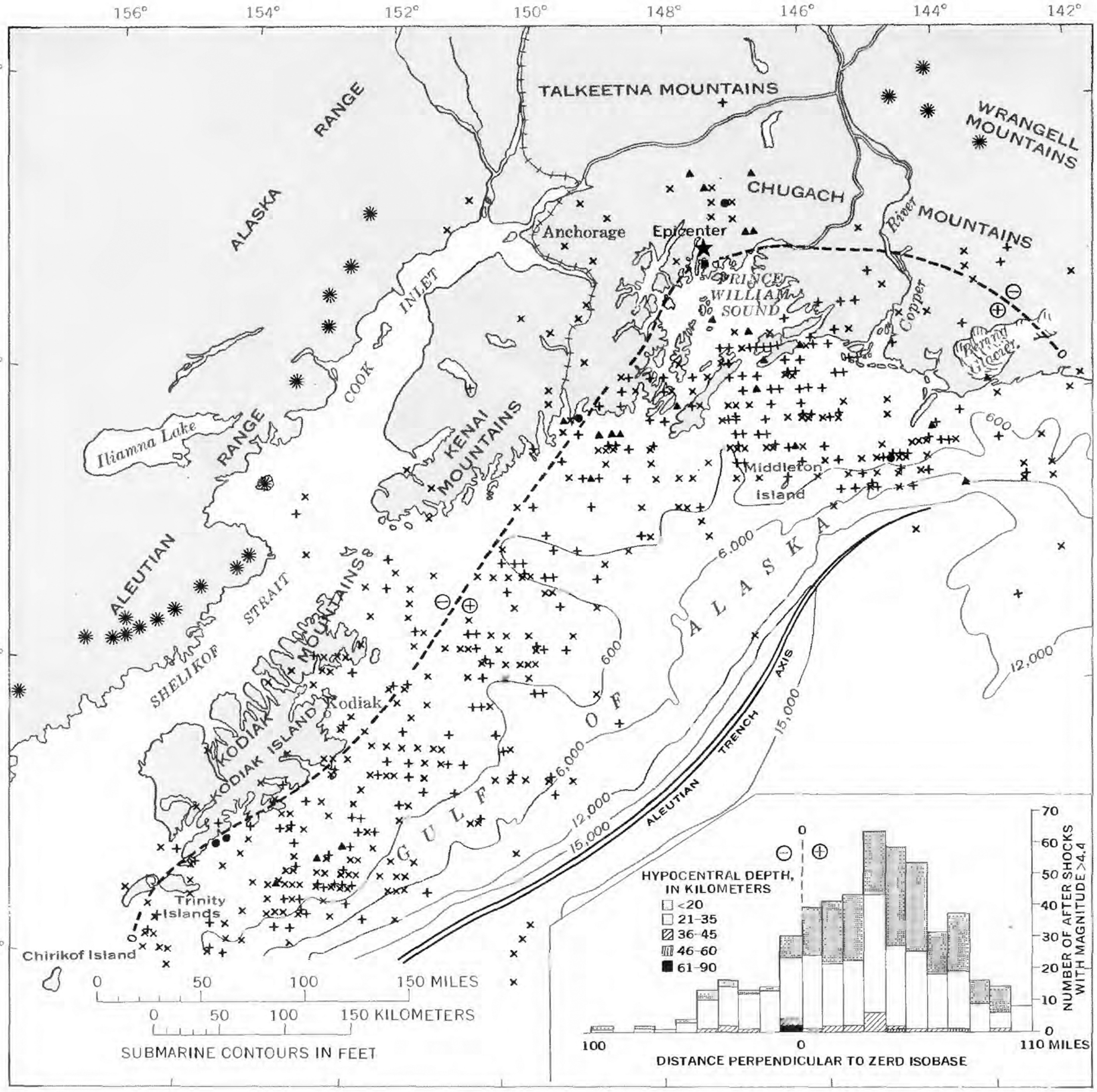
THE AFTERSHOCKS

The epicenters and depth distributions of 598 aftershocks with Richter magnitudes equal to or greater than 4.4 recorded through December 31, 1964, are shown on figure 2. Of these, the largest shock had a magnitude of 6.7, six were larger than 6.0, 127 were between 5.0 and 6.0, and the remainder were less than 5.0 (U.S. Coast and Geodetic Survey, Preliminary Determination of Epicenter cards). During this same period of time, thousands of smaller aftershocks were also recorded.

The aftershock sequence diminished rapidly in frequency and intensity. All aftershocks with magnitudes larger than 6.0 recorded teleseismically by the U.S. Coast and Geodetic Survey, occurred within the first several hours and most of those that were strongly felt occurred within 3 weeks of the main event. Daily frequency of all shocks dropped rapidly from a high of about 120 the first day after the earthquake to 30 in 5 days, 15 in 10 days, and a steadily decreasing number thereafter (Jordan, Lander, and Black, 1965, p. 1324). As noted by Jordan and associates, the number of aftershocks at the beginning of the series may be much larger than indicated, because the first aftershocks are commonly masked by the train of large amplitude waves generated by the main shock and these waves may have a duration of several hours at distant stations.

The larger aftershocks ($M \geqq 4.4$) were concentrated mainly in an arcuate belt 600 miles long by as much as 200 miles wide that roughly parallels the continental margin. From the epicenter of the main shock it extends 425 miles southwestward to the vicinity of the Trinity Islands and about 175 miles eastward nearly to long 142° W. (fig. 2). Most of the largest aftershocks in this belt ($M \geqq 5.0$) were situated over that part of the belt lying seaward from the zero isobase between the major zones of uplift and subsidence. Smaller aftershocks ($M \geqq 4.4$) were spread over a larger area that extends inland from the zero isobase beneath the coastal mountains that border the Gulf of Alaska, and the smallest aftershocks ($M < 4.4$) were scattered over an even larger area. There was a distinct concentration of activity within the aftershock belt on the Continental Shelf at the southwestern end and in the area southeast of Prince William Sound near the northeastern end. Burk (1965, p. 150) noted that the fairly sharp southwestern limit of the aftershock belt approximately coincides with a transverse structural boundary on the Alaska Peninsula, and he suggested the possibility that a transverse fault marks the southwestern margin of the crustal block involved in the 1964 earthquake. Only about 3 percent of the aftershocks with magnitudes of 4.4 or more were outside the main belt of activity along the continental margin. Their epicenters were widely scattered, mainly in the Shelikof Strait-Cook Inlet areas and on the ocean floor seaward from the Aleutian Trench axis. Significantly, none of the aftershocks were centered as far inland as the chain of active volcanoes in the Aleutian Range, southern Alaska Range, and the Wrangell Mountains. The possibility that they were directly related to vulcanism is thus ruled out.

Depth distribution of the aftershocks is less perfectly defined than their epicentral positions because of (1) inherent errors in the determination of hypocenters in areas of uncertain crustal structure and seismic velocity, and (2) the wide spacing of the seismographs on which the shocks were recorded. Algermissen (1965) found that hypocenters of aftershocks with magnitudes greater than 5.0 were at depths between 3 and 25 miles (5–40 km) and they average about 12 miles (20 km). Only 25 of the earthquakes with magnitudes of 4.4 or more were deeper than 22 miles (35 km), the deepest ones being less than 56 miles (90 km). According to Page (1967; also unpub. data), the microaftershocks—which had a spatial distribution similar to that of the aftershocks—were at depths of 22 miles (35 km) or less. He



EXPLANATION

+ <20 · × 21-35 · ▲ 36-45 · ● 46-60 · ■ 61-90

Epicenter of aftershock showing depth of focus, in kilometers

Approximate zero isobase between major zones of tectonic uplift and subsidence

Active or dormant volcano

Aftershock data after R. A. Page, Jr., Lamont Geological Observatory of Columbia University (written commun., May 1966)

2.—Distribution and depth of aftershocks ($4.4 \leqq M \geqq 6.7$) from March 27, to December 31, 1964.

suggests that subcrustal earthquakes in and near the aftershock region may have represented the normal seismicity of the region. Aftershock hypocenters do not fall into any well-defined planar zone although there is a vague tendency towards a slight deepening of their lower limit beneath the continent. The proportion of hypocenters deeper than 12 miles relative to those shallower than 12 miles also shows an increase in the same direction (fig. 2).

Large earthquakes at shallow and intermediate depths are thought by most geologists and geophysicists to result from sudden rupture, or faulting, in strained rocks. Aftershocks which follow large earthquakes presumably represent continuous adjustments of the strained volume of rock, or focal region, within which faulting occurred (Benioff, 1951). According to this model, therefore, faulting associated with the 1964 earthquake was largely confined to the part of the continental margin extending roughly 150–200 miles northward from the axis of the Aleutian Trench, it was limited in depth to the crust, or perhaps the uppermost part of the mantle, and its lower limit may deepen slightly beneath the arc.

FOCAL MECHANISM STUDIES

Mechanism studies of the main shock, of a number of larger aftershocks, and of one preshock that occurred about 7 weeks prior to the earthquake provide data relevant to the fault orientation and sense of displacement at the earthquake foci. Body-wave solutions define a pair of orthogonal planes at the focus, one of which presumably contains the active fault surface. Inherent in the focal mechanism studies are the basic assumptions of an elastic-rebound source and initial displacements at the earthquake foci that approximately reflect the regional stress field.

Focal mechanism studies of *P*-waves for the main shock yield one well-defined nodal plane that strikes between N. 61° and 66° E. and dips 82°–85° SE. (Stauder and Bollinger, 1966. Algermissen, 1965; written commun., March 19, 1965). The alternative low-angle plane is restricted by the data to a plane that dips towards the northwest; Algermissen's solution suggests an inclination of about 8°. If the well-defined nodal plane is regarded as the fault plane, its strike was N. 61°–66° E. and the motion was predominantly dip-slip on a steep reverse fault with the southeast side relatively upthrown. Alternatively, if the steep nodal plane is considered to be normal to the motion, the fault would be a northwest-dipping thrust. Focal mechanism studies of the main shock alone cannot distinguish which of these two planes is the fault plane. Surface-wave studies that define the direction of rupture may permit distinction between the fault and auxiliary planes in cases where the strikes of the two planes differ significantly. However, as noted by Savage and Hastie (1966, p. 4900), surface waves do not permit a unique solution for the 1964 earthquake because the direction of rupture propagation is essentially the same for either plane.

Stauder and Bollinger (1966) determined focal mechanisms, based on combined *P*-wave first-motion and *S*-wave polarization data, for a preshock and 25 aftershocks in the Kodiak Island and Prince William Sound areas. Most of these solutions have one near-vertical nodal plane that resembles the well-defined nodal plane for the main shock; the other dips 5°–15° to the northwest or north. Strike of the steep plane is between N. 50° and 72° E. to the southwest of Prince William Sound; it is variable in the Prince William Sound area and nearly east-west to the east of the sound. This systematic variation in orientation of the steep nodal plane tends to follow a change in trend of tectonic features along the coastal belt. Four aftershock solutions in the Prince William Sound area and one located seaward from the Aleutian Trench off Kodiak Island, however, are anomalous in that they do not correspond to this general pattern.

The preshock and all the aftershock fault-plane solutions are subject to the same ambiguity of interpretation as the main shock. Thus, on the basis of all the individual solutions (except for the five apparently anomalous ones) the motion at the source of the earthquake and the related pre- and aftershocks may be (1) almost entirely dip-slip on a steeply dipping reverse fault along which the seaward side is relatively upthrown or (2) dip-slip on a northward-dipping thrust fault along which the landward block overrides the seaward block in an average S. 25° E. direction to the southwest of Prince William Sound and in a S. 10°–15° W. direction to the east of Prince William Sound (Stauder and Bollinger, 1966, p. 5295). In considering the solutions in relation to one another, Stauder and Bollinger observe that in the first alternative the faulting may consist of en echelon segments that follow a sinuous path roughly paralleling the curving trend of both the aftershock belt and the zero isobase between the major zones of earthquake-related vertical tectonic deformation (fig. 2). In the second alternative the thrust plane has a dip of less than 14° beneath the continent, displacement of the upper plate is relatively seaward, and the direction of motion is roughly normal to the trend of the aftershock belt and the zones of tectonic deformation.

The steep plane in solutions of the main shock, the preshock, and most aftershocks differs in strike from the tectonic trends of the region, the orientation of the earthquake focal region, and the pattern of vertical displacements associated with the earthquake. These differences suggest to Stauder and Bollinger (1966, p. 5293–5294) that the steep plane corresponds to the auxiliary plane rather than the fault plane. They also interpret the spatial distribution of foci for which mechanism studies were made and the nature of the inferred motions as being more compatible with thrusting on the shallowly dipping plane than to movement on the steep plane.

Solutions for the main shock and most of the other shocks suggest a maximum-stress axis at the foci of these earthquakes oriented approximately normal, or at a small oblique angle, to major structural elements within the focal region. Within the limitations of the data, this orientation is in reasonably good agreement with geodetic and geologic data cited in subsequent sections (p. I 49) that suggest dominantly tangential compressive stress on land within the region affected by tectonic movements during the earthquake.

DEFORMATION

Crustal deformation, including both vertical and horizontal movements, associated with the 1964 Alaska earthquake was more extensive than any known to have been related to a single tectonic event. Vertical movements occurred over an arcuate region that roughly parallels the continental margin for almost 600 miles from the southwestern tip of the Kodiak group of islands northeastward through Prince William Sound and thence eastward to about long 142° W. (fig. 3). In a northwest to southeast direction, the deformation extends at least 200 miles from the west shore of Cook Inlet to Middleton Island at the seaward edge of the Continental Shelf. In addition, crustal warping appears to extend inland as far as the Alaska Range and it may extend seaward to the axis of the Aleutian Trench.

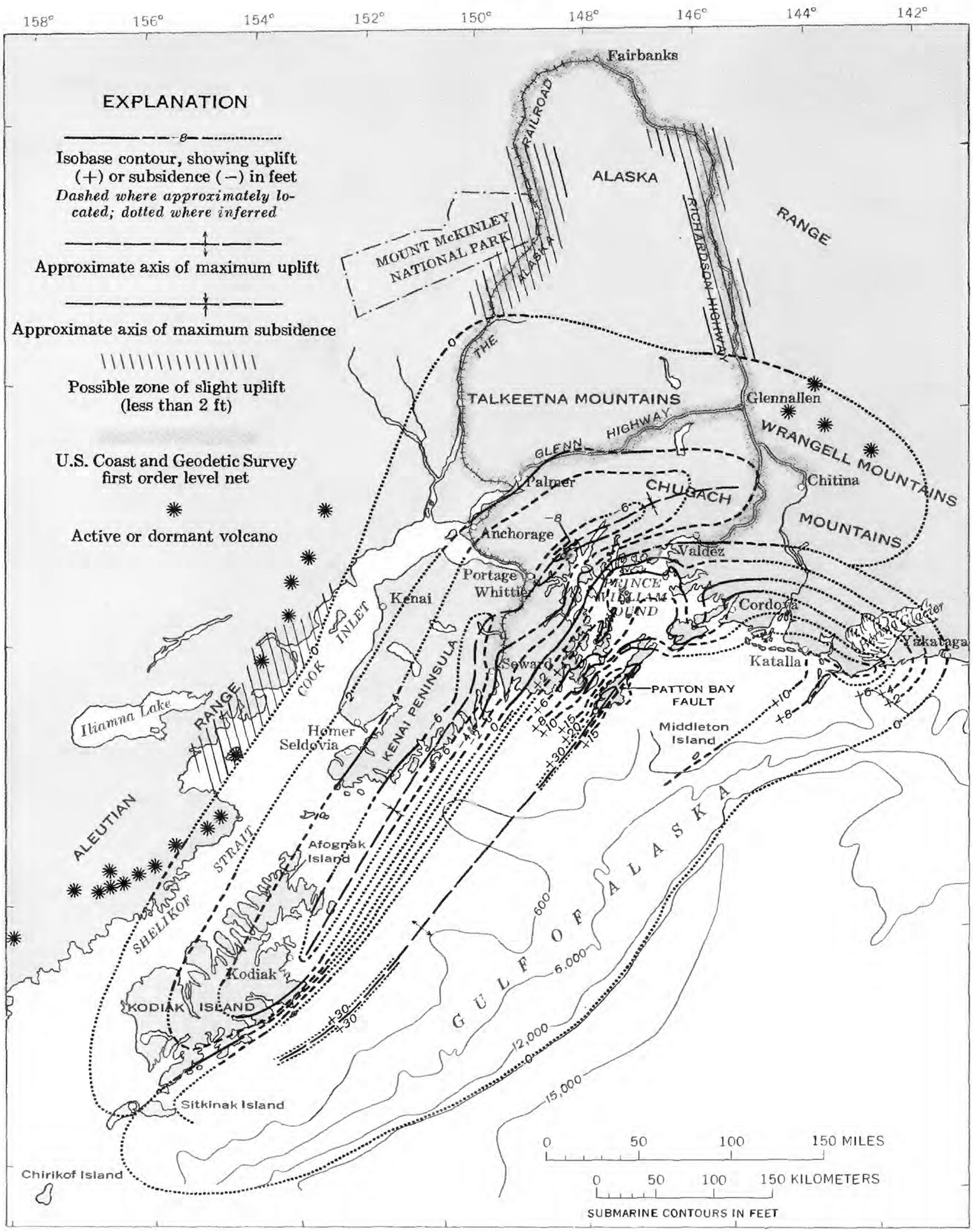
Observable tectonic deformation involved (1) regional crustal warping, including both uplift and subsidence relative to sea level, in broad zones that roughly parallel the trend of the continental margin, (2) systematic regional horizontal extension and shortening in a direction approximately transverse to that of the zones of warping, (3) displacement across longitudinal reverse faults exposed on land and on the sea floor, and (4) possible displacement on at least one wrench fault. Evidence for these various movements, their manifold effects, and their tectonic significance are discussed below.

REGIONAL VERTICAL DISPLACEMENTS

Notable tectonic changes in land level during the 1964 earthquake occurred over an area of at least 70,000 square miles, and probably more than 110,000 square miles, of south-central Alaska. The areal distribution and approximate amount of the vertical displacements are summarized on figure 3. Plates 1 and 2 show data points where quantitative measurements of vertical displacement were made, as well as the method used and year of measurement. Also shown on the figure and plates are isobase contours, or lines of equal vertical displacement based on these data, and the approximate axes of the major upwarped and downwarped zones. The deformation includes two broad major zones of warping, each about 600 miles long and as much as 130 miles wide. The seaward zone is one of uplift that includes a fringe of coast along the Gulf of Alaska, the adjacent Continental Shelf, and perhaps the continental slope; it is bordered to the northwest and west by a zone of subsidence. Slight uplift also occurred in at least three areas extending inland from the major zone of subsidence as far northward as the Alaska Range.

METHODS OF MEASUREMENT

Quantitative information on vertical displacements along the coast (pls. 1, 2) comes mainly from (1) comparison of pre- and postearthquake tide gage readings, (2) the position of the upper growth limit of certain sessile intertidal organisms relative to sea level, (3) differences in the pre- and postearthquake positions of the upper growth limits of sessile intertidal organisms or the lower growth limit of terrestrial vegetation, (4) differences in the heights of pre- and postearthquake storm beaches, (5) estimates or measurements of changes in the position of shoreline



3.—Map showing the distribution of tectonic uplift and subsidence in south-central Alaska.

markers by local residents, and (6) measured changes in the height of tidal bench marks relative to sea level. For those offshore areas where detailed preearthquake bathymetry was available, approximate vertical displacements were also obtained from comparison of pre- and postearthquake depth soundings. The amount and distribution of the vertical displacements inland from the coast (pls. 1, 2) was precisely determined along the highway-railroad routes between Seward, Anchorage, Fairbanks, and Valdez by comparison of pre- and postearthquake level lines tied to tidal bench marks.

Isobase contours plotted on plates 1 and 2, which were derived from all the sources listed above, represent absolute changes in altitude related to the earthquake. The accuracy of the contouring varies greatly from place to place depending upon the type and amount of data, but, because it is essentially an averaging process, the contouring tends to be most accurate in areas where many data points are available, and least accurate in areas where data are sparse. In general, contours shown as solid or dashed lines are estimated to be accurate at least to within ±1 contour interval; at most places they are probably accurate to within half an interval.

The various techniques used for determining vertical displacement, estimates of their relative precision, and sources of data are outlined below.

COMPARISON OF PRE- AND POSTEARTHQUAKE TIDE-GAGE READINGS

The directions and relative amounts of vertical displacement were determined from coupled pre- and postearthquake tide-gage readings made by the U.S. Coast and Geodetic Survey at two

TABLE 1.—*Land-level changes based on comparison of pre- and postearthquake tide series at tide stations in south-central Alaska*

[After Small (1966, p. 19–22) and U.S. Coast and Geodetic Survey (unpub. data, April 22, 1966). Asterisk indicates station is plotted in fig. 10; double asterisk indicates standard tide-gage station operational before and after the earthquake]

Location	Length of tide series		Land movement (feet)
	Preearthquake	Postearthquake	
Prince William Sound			
*Cordova	1950–1951 (2 years)	May–Nov. 1964	+6. 2
		May–Nov. 1965	+6. 2
*Port Gravina	July 17–Sept. 26, 1913	July 3–31, 1964	+4. 5
	June 15–Aug. 19, 1915	May 16–June 16, 1965	+4. 2
Port Fidalgo	Aug. 24–26, 1915	June 18–July 17, 1965	+2. 4
Valdez	1924–26 (3 years)	May–July 1964	−. 9
		July–Nov. 1965	[1] −3. 5
*Whittier	June 12–29, 1956	May–June 1964; Oct.–Dec. 1964	−5. 4
		July–Nov. 1965	[1] −5. 7
*Chenega Island	July–Aug. 1957	July 7–Aug. 4, 1964	+4. 8
Green Island	Aug. 23–Sept. 20, 1911	August, 1965	+6. 6
*Port Chalmers	July–Sept. 1933 (3 months)	May 19–30; July 7–Aug. 6, 1964	+10. 5
		July 19–Aug. 23, 1965	+10. 5
*Sawmill Bay, Evans Island	May, 1927 (1 month)	May 20–June 1; June 14–July 7, 1964	+7. 2
		July 29–Aug. 31, 1965	+7. 2
Hogg Bay	June 15–July 21, 1927	July 20–Aug. 16, 1965	+5. 8
Kenai Peninsula and Cook Inlet			
Day Harbor	May 20–July 13, 1928	May 7–June 30, 1965	−. 5
**Seward	1926–38 (13 years)	June 1964–Jan. 1965	−3. 6
		Jan.–Sept. 1965	−3. 6
Aialik Bay	July 6–Sept. 21, 1912	May–June 1965	−4. 5
*Two Arm Bay	July 20–Sept. 18, 1928	May 14–June 30, 1965	−5. 4
Chance Cove	May 14–July 25, 1930	May 15–June 30, 1965	−6. 6
Shelter Cove	June–Aug. 1927 (3 months)	July 1965	−5. 4
*Port Dick	Aug. 1–Sept. 26, 1930	May 21–June 30, 1965	−6. 2
**Homer	Oct. 3–Nov. 17, 1962	June–Dec. 1964; Sept.–Nov. 1965	−5. 4 [1] −5. 8
*Port Chatham		May 22–June 4, 1965	−4. 6
*Seldovia	April 30–Oct. 8, 1908	June–Oct. 1964	−3. 9
	May 8–Sept. 29, 1910	April–July 1965	−3. 8
Nikiski		June 18–July 31, 1964	−. 9

See footnote at end of table.

TABLE 1.—*Land-level changes based on comparsion of pre- and postearthquake tide series at tide stations in south-central Alaska*—Continued

Location	Length of tide series: Preearthquake	Length of tide series: Postearthquake	Land movement (feet)
Kenai Peninsula and Cook Inlet—Continued			
Anchorage	June–Sept. 1910; June–Oct., Dec. 1922; May–Nov. 1923; May–Sept. 1924; May–Oct. 1925	May–Oct. 1964	−2.6
		April–Oct. 1965	−2.3
Kodiak group of islands			
Carry Inlet	July 1931	July 8–Aug. 7, 1965	−3.2
Redfox Bay	July–Aug. 1926	July 1–Aug. 8, 1965	−3.4
*Tonki Bay	Aug. 23–31, 1932	July 10–Aug. 7, 1965	−5.2
Nachalni Island	July 30–Sept. 8, 1941	June 14–July 10, 1965	−3.9
Dolphin Point	Sept. 10–Oct. 1, 1941	June 7–July 8, 1965	−2.9
St. Paul Harbor		June–Oct. 1964	−5.5
		May–Nov. 1965	−5.0
*Ugak Bay	August 1932	June 12–July 10, 1965	−4.2
Port Hobron	June 1–Sept. 19, 1928	July 26–Aug. 27, 1965	−.7
Jap Bay	July 1931	July 28–Aug. 24, 1965	.0
*Lazy Bay	May 16–Sept. 29, 1929	June 11–30, July 1–Aug. 14, 1964	−.4
	June 1–Sept. 30, 1930	July 28–Aug. 15, Aug. 20–24, 1965	−.2
*Larsen Bay	Aug. 19–22, 1929	June 13–Aug. 31, 1964	−2.5
		July 18–Aug. 17, 1965	−2.4
Uyak Bay	July 14–30, 1908	July 22–Aug. 17, 1965	−1.9
*Port O'Brien	June 22–July 30, 1929	July Aug. 1964	−3.6
		June 8–July 19, 1965	−3.6
**Women's Bay	1950–59 (10 years)	April–July 1964	−5.6
Alaska Peninsula			
**Kukak Bay	July–Aug. 1949 (2 months)	July 17–Aug. 16, 1965	−.5

[1] Change between 1964 and 1965 possibly due to surficial compaction.

operative standard stations, and 34 temporary stations within the region affected by tectonic movements during the earthquake (table 1). Such measurements were made at localities where a series of pre-earthquake tidal observations had been made and where tidal bench marks had been established that were not destroyed during or before the earthquake.

The Seward and Womens Bay (Kodiak) stations were equipped with standard automatic tide gages; readings could therefore be compared for series taken immediately before and after the earthquake. These determinations of vertical displacement are probably accurate to within a few tenths of a foot. The only other standard station in the region, at Homer Spit in Kachemak Bay, could not be used directly to measure vertical tectonic displacement because it was inoperative at the time of the earthquake, and the tectonic subsidence was augmented by pronounced surficial settlement of the unconsolidated deposits that made up the spit. Because the amount of surficial settling was known from level lines to areas of relatively firm ground (Waller, 1966a, p. D13), net tectonic subsidence could be obtained by subtracting the surficial effect from total subsidence relative to sea level at the gage. This difference amounted to about 2.9 feet.

Accuracy of the changes determined at the temporary stations depends largely on the length of the preearthquake series of observations, some of which were as short as 4 days, and also on the time interval between the pre- and postearthquake series, which at one station is 54 years and at many is more than 30 years (table 1). Positions of the preearthquake tidal datum planes at these stations may not have been precisely determined originally, and (or) they may have changed from their original values because of relative land-level changes in the time interval between tide observations. At most of the temporary stations such errors are believed to be small because, where other sources of data such as estimates by local residents or measured differences between pre- and postearthquake shoreline features are available in the immediate vicinity, they tend to agree with the tide data within 0.5 foot or less. An exception is the 7.0-foot uplift of the Sawmill Bay station on Evans Island, Prince William Sound, derived from the tidal observations(table 1). It is about 2 feet too low when compared with (1) the 9.0-foot estimated uplift of shoreline features whose preearthquake heights were precisely known to residents of the area and (2) the 8.9 feet of uplift indicated by differences in the pre- and postearthquake upper growth limit of barnacles (pl. 2). The preearthquake tidal observations at this particular station consist of a 1-month series taken in 1927, 37 years before the postearthquake series with which it was compared. Therefore, either about 2 feet of uplift occurred in the area during the 37 years prior to the earthquake or the tidal measurements are in error. Because Sawmill Bay had been continuously inhabited during this time interval and because there was no indication of preearthquake changes in land level, it appears probable that the latter alternative is the correct one.

THE UPPER GROWTH LIMIT OF SESSILE INTERTIDAL ORGANISMS RELATIVE TO SEA LEVEL

Vertical displacements in coastal areas were determined mainly from more than 800 measurements of the upper growth limits of intertidal sessile marine organisms relative to sea level along the long, intricately embayed rocky coast. In Prince William Sound and Resurrection Bay, where we used outboard-motor-powered skiffs for studying these changes during 1964, measurements were made continuously along the shore at spacings of about 1–5 miles except in those places where cliffs, heavy surf, or floe ice in fiords prevented boat landings or where the upper growth limits were not well-defined. During both the 1964 and 1965 seasons, measurements of vertical displacement were also made at localities shown on plates 1 and 2 that were accessible by light plane or helicopter along the ocean coast, on the offshore islands, and around the shores of the Kodiak group of islands.

In measuring land-level change from the displacement of sessile marine organisms relative to sea level, the zonation of plants and animals between tide marks was used—a zonation that has long been recognized by marine ecologists. The intertidal zone along the predominantly steep and rocky coastline of south-central Alaska is inhabited by certain species of organisms—notably barnacles, mussels, and algae—whose vertical growth limits are usually well defined (pl. 3*A*, facing p. I 16; fig. 4). The zonation of intertidal organisms in the Prince William Sound region was studied in detail by a party headed by G Dallas Hanna of the California Academy of Sciences in 1965. To the writer's knowledge, there were no published preearthquake data on the intertidal ecology for any part of the Gulf of Alaska coast.

In particular, the common acorn barnacle, *Balanus balanoides* (Linnaeus), and closely similar forms such as *B. glandula* are widely distributed on rocky shores and form a conspicuous band with a sharply defined, readily recognizable upper limit (figs. 5–7) The common olive-green rockweed (*Fucus distichus*), which has an upper growth limit near that of the barnacles, served as a useful datum for measuring land-level changes along shores where barnacles were absent or poorly developed (pl. 3*B*, facing p. I14). The upper limit of this zone, referred to as the "barnacle line," corresponds roughly to the top of the Balanoid or Midlittoral Zone of Stephenson and Stephenson (1949); to Zone 2, the High Tide Region, or the Upper Horizon of Ricketts and Calvin (1962); and the Upper Intertidal Zone of Rigg and Miller (1949). The barnacle line usually, but not always, approximates the lower growth limit of the dark-gray-to-black encrusting lichen (*Verrucaria*) which commonly forms a black band in the splash zone immediately above the barnacles and *Fucus* (pl. 3*A*, facing p. I16; fig. 4).

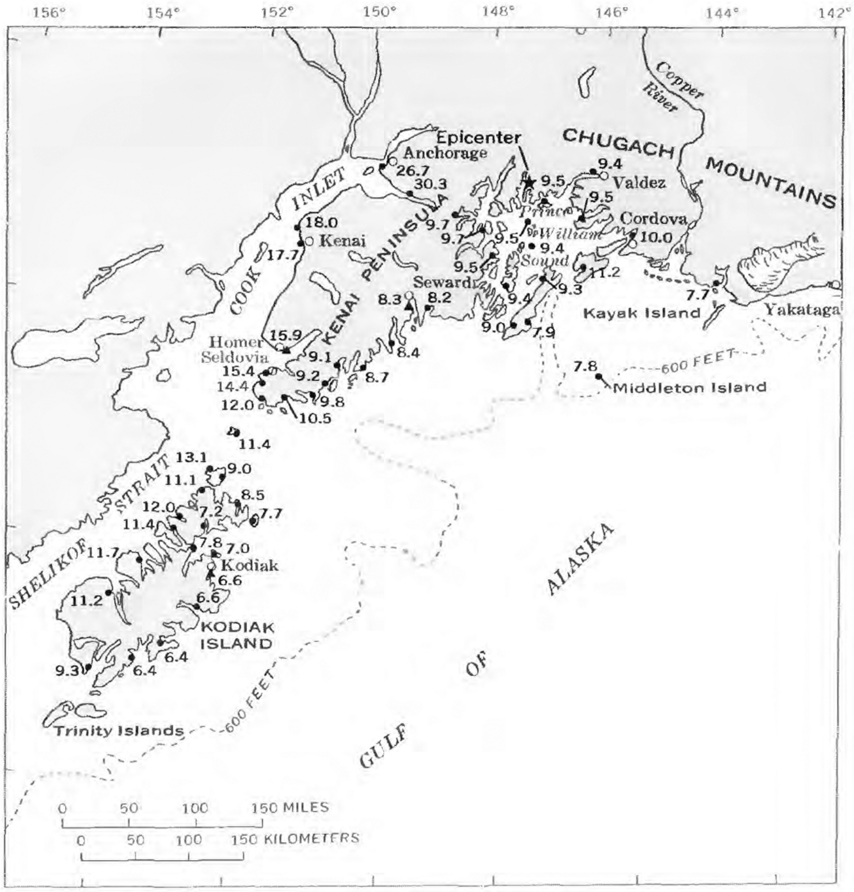
The upper limit of all intertidal organisms depends mainly on the ability of immature individuals to survive prolonged exposure to air and on the tidal characteristics at any given locality (Kaye, 1964, p. 591–592). In referring to the survival ability of barnacles, Kaye has termed this maximum exposure interval the "lethal limit." He found experimentally that it was close to 150 hours for yearling barnacles and ranged to an absolute maximum of 192 hours for mature barnacles. *Fucus*, which has a nearly identical upper growth range, must have approximately the same "lethal limit." To a lesser extent the upper growth limit of barnacles and *Fucus* depends upon a number of other factors which, in parts of Prince William Sound, locally cause the barnacle line to deviate as much as 0.6 foot from its average height at any given locality. Wave action during the lowest annual neap tides and protection from desiccation by 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 organisms' upper growth limits.

4.—Characteristic parallel bands formed by zoned intertidal marine organisms long unlifted west shore of Knight Island, Prince William Sound. Encrusting lichens form the upper dark band and brown laminarians form the lower one; light-colored barnacle zone is between the two. Photograph taken at about 2-foot tide stage, May 20, 1964.

5.—Measuring height of the barnacle line at Port Bainbridge in an area of the Kenai Peninsula uplifted 5.7 feet. Protograph taken at 4.4-foot tide stage, June 21, 1964.

6.—Barnacle line clearly marked by upper limit of light-gray barnacles and desiccated rockweed in dark patches and by the lower growth limit of dark-gray encrusting lichens of the splash zone. Uplift of 3.2 feet in Whale Bay, western Prince William Sound. Photograph taken at 3.0-foot tide stage, June 22, 1964.

7.—Barnacle line on hull of S.S. *Coldbrook*, Middleton Island. The upper growth limit of the dark band of barnacles on this vertical surface is clearly defined and at a uniform level, even though it is in a locality that was exposed to open-coast surf conditions prior to the earthquake. Indicated uplift is 11.7 feet. Photograph taken at 6.5-foot tide stage, July 26, 1965.



8.—Variations of mean tide range in areas of south-central Alaska where tide levels were used as a datum in measuring vertical displacement of the coast. Triangles show locations of permanent tide gages prior to the earthquake. Data from U.S. Coast and Geodetic Survey (1964).

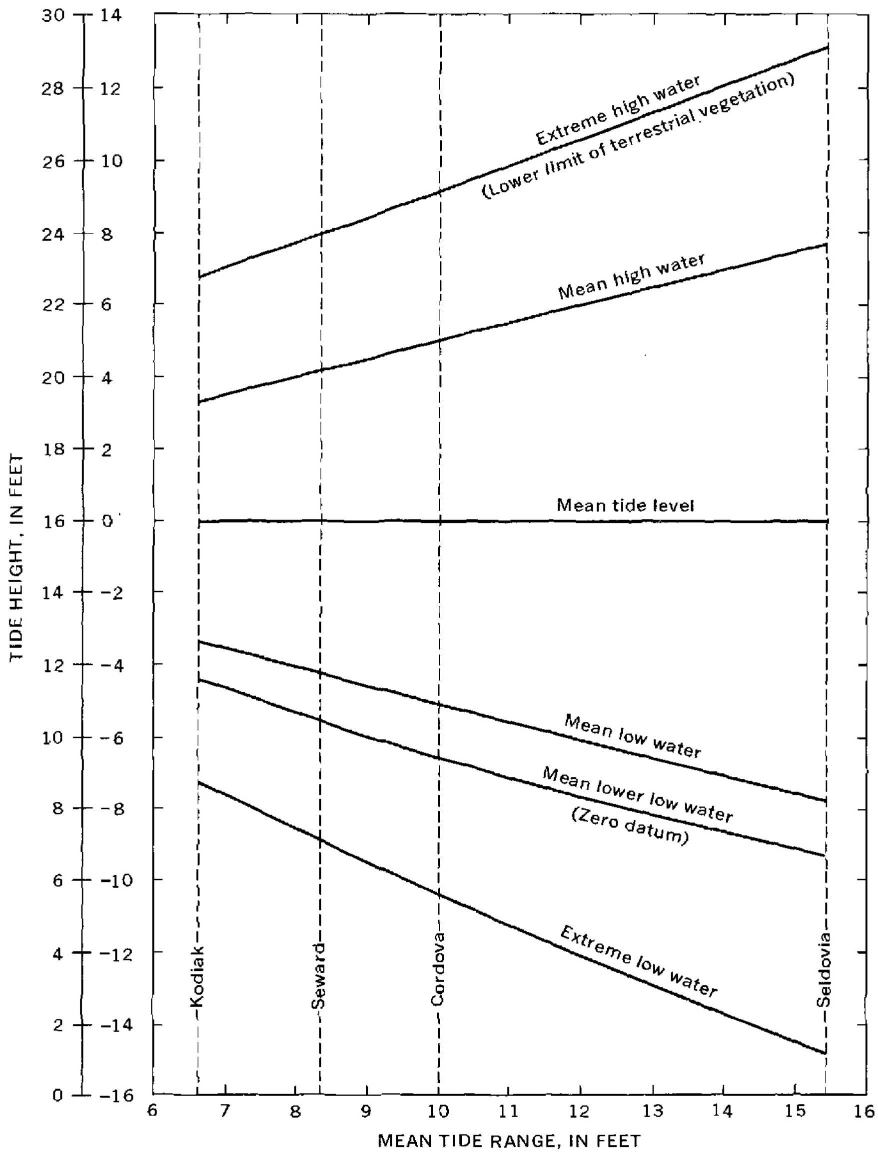
Specific data on the normal upper growth limit of barnacles and *Fucus* relative to tide levels were unavailable for the part of coastal south-central Alaska that was affected by vertical tectonic displacements associated with the earthquake at the time this study was made. However, elsewhere on the Pacific Coast of North America and in other areas of the world the height of the barnacle line roughly approximates annual mean high water along shores with a small or moderate tidal range. In the region where measurements of vertical displacements were made during this study, tidal range, and hence the height of the barnacle line, differs from place to place along the coast. For example, the mean tide range (that is the difference in height between mean high and mean low water) varies from a minimum of 6.4 feet along the ocean coast of the Kodiak group of islands to a maximum of 15.9 feet at Homer near the entrance to Cook Inlet (fig. 8). Even higher tides—as high as 30.3 feet—prevail within Cook Inlet. However, the exceptionally large tidal range, in combination with a general lack of stable rocky shores and an increasingly impoverished marine fauna toward the head of the inlet precluded use of marine organisms for measuring vertical displacement.

On figure 9 are shown the variations in the 1964 positions of six of the important sea-level datum planes corresponding to mean tide ranges of 6.4 feet to 15.9 feet in south-central Alaska.[1] Kaye (1964) has reviewed the definition of these tidal planes, the way in which they are derived from tide-gage records, and the control they exert on the zonation of intertidal organisms.

Field procedure for determining land-level changes by the barnacle-line method was to measure the height of the upper limit of barnacle or *Fucus* growth above or below water level at any stage of tide. On smooth steep rocky slopes sheltered from heavy surf, this line is sharply defined and can be readily determined to within 0.2 foot or less. On sloping shores or shores exposed to heavy surf it tends to be less regular, although even under such conditions it generally can be determined with confidence to within 0.5 foot. At most places where the barnacle line was above water, its height 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.

Stage of tide at the time of measurement was then determined from the U.S. Coast and Geodetic Survey table of predicted tides for

[1] Extreme high and low water are the highest and lowest predicted annual tides; the extreme high-water line is also the approximate lower limit for growth of terrestrial vegetation along shores sheltered from wave splash. Because tides in south-central Alaska are of the semidiurnal type, the levels of the two daily high and low tides differ, varying between a maximum (spring tides) and a minimum (neap tides), depending on the interactions of the tide-producing forces. Thus, mean high water (MHW) and mean low water (MLW) are the average heights of all high and low tides within the period of measurement. Mean lower low water (MLLW), the mean of the spring tides, is the tidal datum plane commonly used as the zero datum by the U.S. Coast and Geodetic Survey. Mean tide level (MTL) is the exact midpoint of the range MHW–MLW. It differs by a few tenths of a foot from mean sea level, the average of all hourly tide readings over a given period, because the tidal curve is not a simple sine curve, but rather is compounded of a number of simple sine curves, some of which have fixed phase relationships with respect to each other (Marmer, 1951, p. 70).



9.—Tidal parameters in areas of south-central Alaska where tide levels were used as a datum in measuring vertical displacement of the coast. Data from U.S. Coast and Geodetic Survey (1964).

the closest reference station, and, finally, the position of the barnacle line relative to the mean lower low water (MLLW) datum was calculated. For measurements made close to the 16 U.S. Coast and Geodetic Survey tide gages that were installed in the area immediately following the earthquake, we later made corrections to the actual, rather than the predicted, tides. During the period of fieldwork, it was found that few tides deviated by as much as 1.5 feet from predictions; most were within a few tenths of a foot of predicted values.

Differences in the height of the barnacle line in local areas of similar tide range provided a powerful tool for determining relative changes in land level along the coast even where the absolute change was not known. Absolute uplift or subsidence relative to sea level at any given locality was taken as the difference in height between the measured elevation of the barnacle line and the "normal" preearthquake upper growth limit for the barnacles and *Fucus* as indicated in figure 10.

The "normal" preearthquake upper growth limit of barnacles and *Fucus* relative to MLLW was determined empirically at the 17 localities listed on figure 10 where the amount of vertical displacement was known from Coast and Geodetic Survey tide gage readings. Its position at these localities was taken as the measured height relative to MLLW corrected for the amount of tectonic uplift or subsidence indicated by the tidegage readings. On figure 10 these heights are plotted against the mean tide range at the station. The least-square curve through these points represents the average preearthquake height of the barnacle line for mean tides ranging from 6.4 to 15.9 feet in the study area. The curve suggests that the preearthquake barnacle line was close to mean high water for the lowest mean tides and that it lowered progressively relative to MHW with an increase in mean tide range. For the lower tides of 6.4–10.0 feet which prevail along most of the Gulf of Alaska and in Prince William Sound, it ranged from MHW to 0.6 foot below MHW level; for the higher mean tides of as much as 15.4 feet at Seldovia in lower Cook Inlet and Shelikof Strait, the barnacle line dropped to 1.7 feet below MHW.

The derived curve for the approximate height of the barnacle line is only a crude approximation to the actual position of the barnacle line. In addition to uncertainties in measurement of the position of the postearthquake barnacle line relative to sea level, it incorporates inherent errors in the determination of land-level change at the temporary tide-gage stations as discussed previously (p. I 11).

EFFECTS OF THE EARTHQUAKE ON INTERTIDAL ORGANISMS

A (right).—Sharply zoned intertidal marine organisms along west shore of Port Bainbridge. The barnacle line, uplifted 6.1 feet, is at the contact between the upper black band of encrusting lichens and the light-gray band of barnacles. Greenish material in the lower part of the rock face is marine algae; its upper limit marks the approximate position of the postearthquake barnacle line. Photograph taken at 3.1-foot tide stage, June 18, 1964.

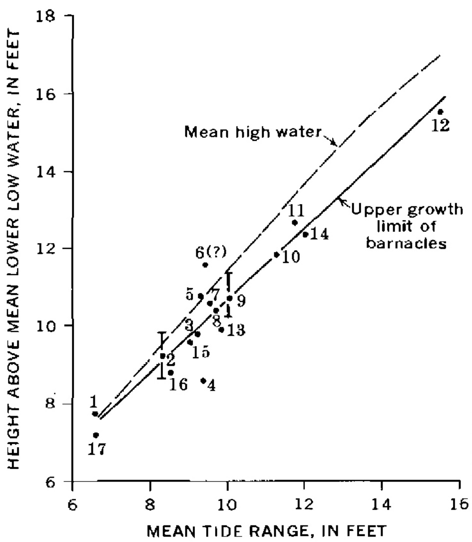
B (left).—Barnacle line defined by upper growth limit of olive-brown rockweed and lower limit of dark-gray encrusting lichens. Shoreline shown (Malina Bay, Afognak Island) subsided about 3 feet during the earthquake. Photograph taken at 6.5-foot tide stage, July 20, 1964.

C (right).—Postearthquake yearling barnacles (light gray) among preearthquake barnacles (yellow). Within 4 months after the earthquake the new crop of barnacles had base diameters of as much as 0.3 inch. Large divisions in upper scale are inches. Photograph taken August 2, 1964.

EFFECTS OF THE EARTHQUAKE ON SHORELINE FEATURES

A.—Living (olive-green) and desiccated (dark-brown) *Fucus* along the shore of Glacier Island, Prince William Sound. The top of the band of desiccated algae was near the preearthquake barnacle line and the top of the band of living algae was near the postearthquake barnacle line. The 3.0-foot difference between their elevations was a measure of the tectonic uplift in this area. Photograph taken at 8.8-foot tide stage, June 13, 1964.

B.—Extensive area of brown terrestrial vegetation at Kiliuda Bay, Kodiak Island, killed by salt-water immersion after about 4 feet of tectonic subsidence and an unknown amount of surficial subsidence. Dikelike gray ridge of beach gravel was built up in adjustment to the new higher base level. The area behind this beach ridge may eventually become a shallow lagoon. Photograph taken July 17, 1964.



Data Points
1. Womens Bay, Kodiak Island
2. Seward, Kenai Peninsula (average of 2 measurements, range indicated by vertical bar)
3. Chenega Island, Prince William Sound
4. Lazy Bay, Kodiak Island
5. Port Chalmers, Montague Island, Prince William Sound
6. Sawmill Bay, Evans Island, Prince William Sound
7. Port Gravina, Prince William Sound
8. Whittier, Prince William Sound
9. Cordova, Prince William Sound (average of 28 measurements; range indicated by vertical bar)
10. Larsen Bay, Kodiak Island
11. Port O'Brien, Kodiak Island
12. Seldovia, Kenai Peninsula
13. Port Dick, Kenai Peninsula
14. Port Chatham, Kenai Peninsula
15. Two Arm Bay, Kenai Peninsula
16. Tonki Bay, Afognak Island
17. Ugak Bay, Kodiak Island

10.—Height of upper growth limit of barnacles above mean lower low water for mean tide ranges of 6.4–15.9 feet in south-central Alaska. The least-square curve for the upper growth limit of barnacles is based on measurements made at 17 stations (dots) where the vertical displacement was determined independently by the U.S. Coast and Geodetic Survey (Small, 1966) from tide-gage observations. Standard deviation is 0.627 foot.

However, the validity of using the empirically determined barnacle line was generally confirmed by our observations in late 1964 and in 1965 that the upper growth limits of new postearthquake barnacles and *Fucus* were generally within about ±1 foot of this line.

The precision of the land-level changes determined by the barnacle-line method varies within wide limits because of the numerous variables involved in making the measurements and the assumptions inherent in the presumed preearthquake position of the barnacle line. The measurements are generally within 1 foot of changes estimated by local residents or found by means of other techniques. Under the least favorable combination of circumstances, such as along segments of the coast exposed to heavy surf or swells or in areas of high and erratic tides in Cook Inlet and Shelikof Strait, measurements may locally be in error by as much as 2½ feet. In such areas, however, more reliance was placed on changes indicated by those techniques that do not require use of tide level as a datum.

DIFFERENCES IN THE EXTREME PRE- AND POSTEARTHQUAKE GROWTH LIMITS OF SESSILE INTERTIDAL ORGANISMS AND TERRESTRIAL VEGETATION

Throughout the region affected by tectonic land-level movements, postearthquake changes in the upper growth limit of barnacles and *Fucus* or in the lower growth limit of terrestrial vegetation provided direct indications of the direction and approximate amount of movement. Thus, along uplifted shores a band of dead barnacles, *Fucus*, and other sessile organisms developed within 2 months after the earthquake. The height of this band reflected the amount of uplift (pl. 4*A*). By July of 1964 a new postearthquake line of young barnacles, and in some places *Fucus*, was well established on most shores. The height of this line above or below the preearthquake line furnished a direct measure of the amount of uplift or subsidence (pl. 3*C*; fig. 11). Similarly, at many places, the amount of subsidence could be clearly determined within 2 months after the earthquake from certain ephemeral features such as the elevation to which the highest spring tides inundated terrestrial vegetation (fig. 12). By the 1965 field season, land plants had become sufficiently well established over much of the uplifted shore that the approximate amount of uplift could be determined from differences in the pre- and postearthquake lower growth limits (fig. 13). This method was particularly useful in areas of uplift such as Middleton Island where, for some unknown reason, barnacles and *Fucus* had not become established in the intertidal zone even a year after the earthquake.

Land-level changes determined from differences in pre- and postearthquake positions of the barnacle line or of the lower limits of terrestrial vegetation provide reasonably precise values for the tectonic movements where the postearthquake growth limits have had time to reach a position in equilibrium with the local tides and where the earthquake-related displacements have not caused significant changes in the tidal characteristics. Such measurements are thought to represent the actual vertical change at least to within 1.0 foot, and probably to within 0.5 foot in most places.

DIFFERENCES IN THE HEIGHTS OF PRE- AND POSTEARTHQUAKE STORM BEACHES

Along uplifted sandy shores on Montague Island in Prince William Sound and along the linear stretch of coast east of Kayak Island, the amount of uplift could be approximated in 1965 from the relative positions of pre- and postearthquake storm beaches (fig. 14). The accuracy of such measurements is difficult to evaluate, although where they could be

11.—Conspicuous white band of postearthquake barnacles along the shore of Kizhuyak Bay, Afognak Island. The difference in elevation between the upper growth limit of the yearling barnacles in the photograph and the preearthquake barnacles, which were at water level, indicates at least 3 feet of tectonic subsidence. Photograph taken July 20, 1964.

12.—Drowned brush and trees along shore of Harriman Fiord, Prince William Sound. The color change (arrow) between the dead brown foliage (light gray) below and the green foliage (darker gray) reflects the position of the postearthquake extreme high-tide line. The difference in elevation between the lower growth limit of terrestrial vegetation and the new extreme high-tide line provided a measure of tectonic subsidence, which was 7.2 feet at this locality. Photograph taken at 0.9-foot tide stage, June 10, 1964.

13.—Wild flowers and grass growing among dead barnacles (white) on shore of Middleton Island uplifted about 11 feet. The differences in the lower growth limits of pre- and postearthquake terrestrial vegetation provided a direct indication of the approximate amount of uplift. Photograph taken July 26, 1965.

14.—Coast at Cape Suckling uplifted about 13 feet during the earthquake. The difference in elevation between the postearthquake storm beach (marked by band of light-colored driftwood) and the preearthquake storm beach, which was above the base of the sea cliff, provided a crude measure of the uplift. The smooth area between the upper limits of driftwood and the sea cliff is now a marine terrace, and the former island in the foreground is a stack on its surface. The flat surface on the stack is probably an older marine terrace. Photograph taken at about zero tide stage, July 24, 1965.

compared with changes at nearby rocky shores, they appear to give results consistent, within about 3 feet, with those obtained from barnacle lines. The measurements between Kayak Island and Yakataga (pl. 1) are particularly uncertain because the shore there consists of active sand dunes that had partially concealed the old storm-beach line by the time measurements were made in 1965. This method was used only where no other means was available for measuring vertical displacement.

ESTIMATES OR MEASUREMENTS BY LOCAL RESIDENTS

Where possible, data on local land-level changes were obtained from local residents in interviews and on form questionnaires. The amount of these changes and the confidence limits expressed by observers are shown on plates 1 and 2. Most of the estimates were made by fishermen, mariners, loggers, and other coastal residents who had long experience in observing the levels of local tides relative to familiar shoreline features. Consequently, most of their estimates or measurements of the vertical displacements are probably correct to within a foot or less.

HEIGHT OF TIDAL BENCH MARKS RELATIVE TO SEA LEVEL

At a few localities in Prince William Sound, changes in land level were determined by leveling from the water surface to U.S. Coast and Geodetic Survey tidal bench marks of known preearthquake elevation. The accuracy of these determinations, which depends mainly upon the precision of leveling and the degree to which the actual tides at the time of measurement correspond with predicted tides, is believed to be within 0.5 foot.

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COMPARISON OF PRE- AND POSTEARTHQUAKE DEPTH SOUNDINGS

Submarine control for the offshore uplift indicated southwest of Montague Island is provided by comparisons of detailed bottom soundings taken by the U.S. Coast and Geodetic Survey in 1927 and after the earthquake in 1964 (Malloy, 1964, p. 1048–1049; 1965, p. 22–26). Because of the technical problems involved in carrying out such surveys, however, the inferred submarine displacements could locally be in error by 10 feet or more.

COMPARISON OF PRE- AND POSTEARTHQUAKE LEVELINGS

The amount and distribution of the vertical tectonic movements inland from the coast were defined by the U.S. Coast and Geodetic Survey's prompt releveling of 722 miles of previously surveyed first-order level lines connecting the cities of Seward, Anchorage, Valdez and Fairbanks (pl. 1; fig. 3). Details of the methods used in these surveys were presented by Small (1966); changes in elevation between successive levelings at bench marks along these routes have been tabulated by Wood (1966, p. 124–130). The changes indicated by this method are probably accurate to within a few tenths of a foot.

Small (1966, p. 2) has cautioned that observed divergences in the results of the original leveling dating from 1923 and the postearthquake leveling in 1964 and 1965 may not be due entirely to movement associated with the earthquake because there is evidence for minor gradual regional movement along some lines where repeated levelings were made prior to 1964. At a few bench marks along these routes, which were located on unconsolidated deposits, the indicated subsidence was anomalously large. Such anomalous measurements obviously represented changes caused by accidental displacement of the bench marks or by local phenomena such as frost heaving, surficial settling, or thawing of permafrost rather than to tectonic movements; they are not shown on plate 1.

DISTRIBUTION OF LAND-LEVEL CHANGES

Figure 3 summarizes the known and inferred areal distribution of land-level changes. The deformation extends for almost 600 miles along the Gulf of Alaska coast from the southwest tip of the Kodiak group of islands through the Prince William Sound region and eastward to the vicinity of Yakataka where it seems to die out. The deformed region consists essentially of (1) a broad zone of subsidence centered along the axis of the Kodiak-Kenai-Chugach Mountains, (2) a major zone of uplift that borders it on the seaward side and extends from the coast onto the sea floor, and (3) a zone of slight uplift that borders it on the landward side and extends northward into the Alaska and Aleutian Ranges. The distribution of subsidence and uplift in these three zones is described below.

THE ZONE OF SUBSIDENCE

The zone of tectonic subsidence includes almost all of the Kodiak group of islands, most of the Kenai Peninsula, the northern and western parts of Prince William Sound, and probably the western segment of the Chugach Mountains (pls. 1, 2; fig. 3). Areas of subsidence in most rocky embayed coastal areas are clearly defined by the various criteria outlined in the preceding section or by qualitative indicators of shoreline submergence. In sheltered embayments the changes may be noticeable where the subsidence is as little as 1 foot. Such effects were most pronounced in areas of lowest mean annual tide range (fig. 8), inasmuch as both the frequency and duration of shoreline immersion for a given amount of subsidence vary inversely with the tidal range.

Along coasts with large tidal ranges and nonrocky shores, such as the part of Cook Inlet north of Homer, the subsidence is known only from tide gage readings near Kenai and at Anchorage, from a few observations by local residents, and from leveling along the coast near the head of the inlet. In this area, and along Shelikof Strait, the northwestern limit of the zone of subsidence is poorly defined. It seems to be close to the west side of the inlet from Redoubt Bay southwestward to Kamishak Bay and probably extends inland between the south side of that bay and the general area of Katmai Bay.

Control on the distribution and absolute amount of subsidence inland from the coast is provided by the Coast and Geodetic Survey's releveling of the first-order net shown on figure 3. Subsidence was indicated on all these lines south of the approximate southern margin of the Alaska Range except in the immediate vicinity of Valdez where a few bench marks were uplifted less than 0.2 foot (pl. 2). The leveling clearly demonstrates that the subsidence extends as a broad warp without abrupt changes of level across the Kenai Mountains northward from Seward and across the Chugach Mountains north of Valdez (pl. 1). Within the Chugach Mountains, subsidence of about half a foot extends eastward at least to Chitina. The northern limits of the zone are approximately defined by the leveling along The Alaska Railroad and Richardson Highway. Because of the small measured land-level changes on those lines, errors of as little as 0.5 foot could cause shifts of as much as 30 miles in the position of the northern boundary of the zone.

MAJOR ZONE OF UPLIFT

The main zone of uplift on land, as determined from shoreline changes, includes (1) a narrow fringe of points, capes, and small islands along the seaward side of the Kodiak group of islands, (2) all but the extreme northwestern and northern parts of the Prince William Sound region, and (3) the coastal belt extending about 120 miles east of the sound. Direct indications of uplift of parts of the contiguous Continental Shelf are afforded by emergence of all the offshore islands and reefs, including Middleton Island near the edge of the Continental Shelf (pl. 1, 2; fig. 1).

The extreme southwestern limit of the zone is believed to lie between Sitkinak Island, which was uplifted about 1½ feet according to a local resident (Mr. Hall Nelson), and Chirikof Island, where there apparently was no change in level (Neal Hoisington, written commun., 1965). Its eastern limit is probably at, or just west of, Yakataga.

The trend of the isobase contours in the northeastern part of the zone of uplift (fig. 3) and the distribution of aftershocks (fig. 2) seem to justify the inference that uplift also occurred over much of the submarine part of the continental margin in a broad zone extending southwestward at least to the latitude of southern Kodiak Island. Seaward projection of the trend of the isobase contours from the area between Yakataga and Middleton Island also suggests that uplift occurred over much of the continental slope and could have extended to the toe of the continental slope, as is shown in figure 3 and on plate 1.

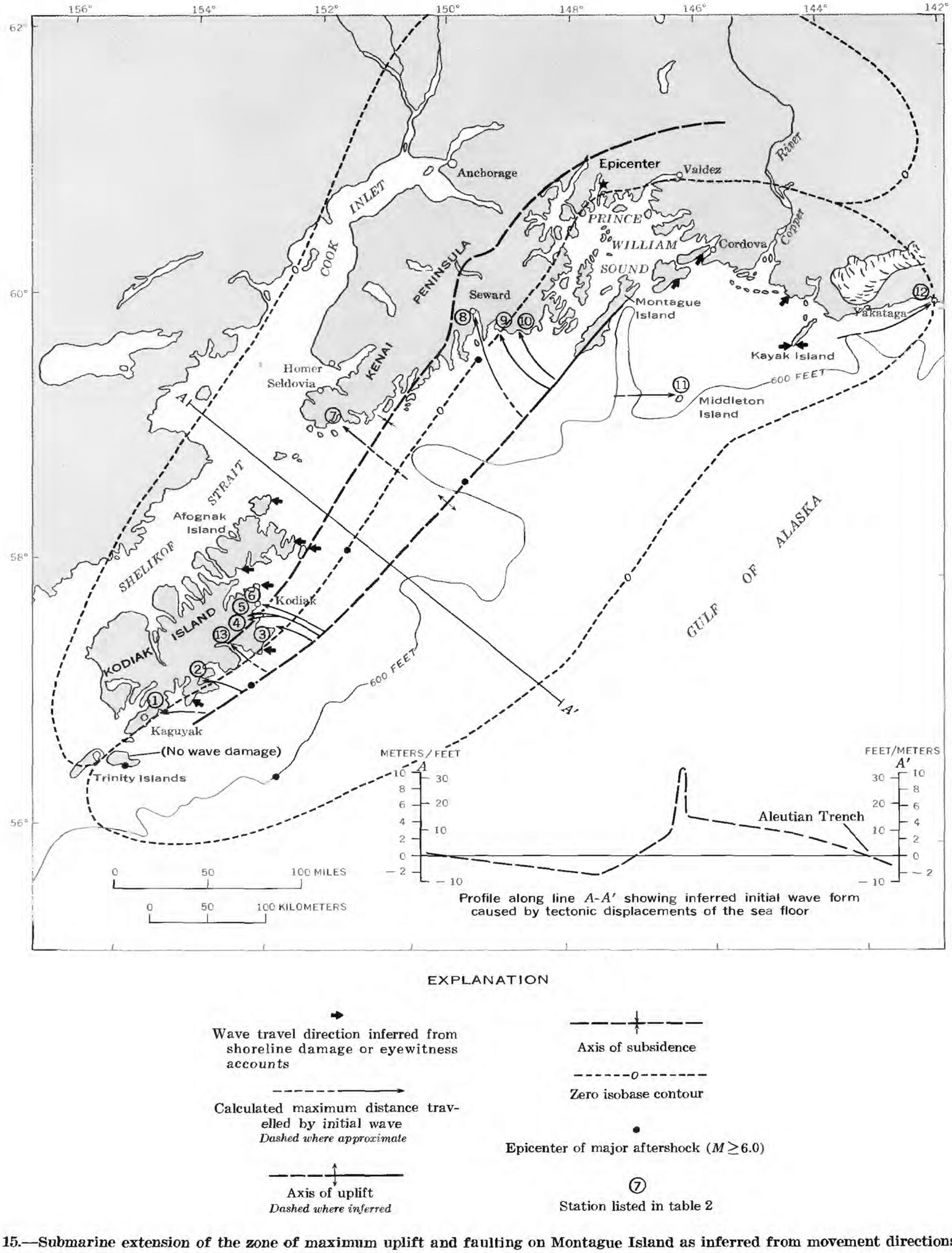
Independent evidence for uplift over a large segment of the Continental Shelf and slope comes from the seismic sea waves (tsunami) generated by the submarine movements. Because seismic sea waves are gravity waves set up in the ocean mainly by vertical disturbances of the sea bottom, the sense of displacement in the generative area can be determined under favorable conditions from the initial water motion at suitably situated tide stations. Tide-gage records outside the immediate area affected by the earthquake show an initial rise that indicates a positive wave resulting from upward motion of the sea bottom (Van Dorn, 1964, p. 166). The initial direction of water movement along the coast of the Gulf of Alaska within the area affected by the earthquake is less clear, however, because there were no operative tide gages, and in many localities the water movements reported by eyewitnesses were complicated by (1) changes of land level along the coast, (2) local waves generated mainly by submarine landslides, and (3) seiches, or other water disturbances related to horizontal tectonic displacement.

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. Distances traveled by the waves can be determined if both the wave velocity along the propagation path and the travel time are known. Because of their long wavelengths, seismic sea waves move as shallow water waves even in the deepest ocean, and their approximate velocity is given by La Grange's equation (in Lamb, 1932, p. 257):

$$V=\sqrt{gh}$$

were g is the gravitational constant, and h is the water depth along the travel path (as determined from nautical charts). Travel time is taken as the elapsed time between the main shock and the arrival of the first wave crest at shore stations. The distribution of tide gage stations outside the seismic sea-wave generative area precludes precise delineation of the source by this method. However, its general position as derived by Van Dorn (1964, fig. 8), Pararas-Carayannis (1967), and M.G. Spaeth (oral commun., Sept. 1964) is consistent with uplift in the broad zone that lies roughly between the Aleutian Trench axis and the coast and extends from the general area offshore from Yakataga southwestward to about the latitude of Kodiak.

The position of the axis of the wave source shown on figure 15 was inferred from the arrival times of the initial wave crest along the adjacent coast, the general distribution of wave damage, and the reported movement directions of the initial wave. Approximate travel times to shore stations, sense of initial water motion, and the data sources are given in table 2. These data suggest that the wave crest was generated along one or more line sources within an elongate belt that extends about 250 miles from the vicinity of Montague Island in Prince William Sound to the area offshore from Sitkalidak Island in the Kodiak group. This inference is supported by the fact that at the northeast end of this axis on Montague Island warping and faulting have resulted in uplift of 38 feet in a



15.—Submarine extension of the zone of maximum uplift and faulting on Montague Island as inferred from movement directions and calculated travel distances of seismic sea waves generated by the tectonic displacements.

TABLE 2.—*Travel times of seismic sea-wave crest to near-source observation stations*

[Locations shown on figure 15. Travel time from start of the earthquake (5:36 p.m., A.s.t.)]

Station No. (fig. 15)	Station Name	Travel time (minutes)	Reported sense of first motion	Data source
1	Kaguyak	38±5	?	Larry Matfay (radioed message received at Old Harbor).
2	Old Harbor	48	Up	Larry Matfay.
3	Cape Chiniak	38	Up	Fleet Weather Central, Kodiak Naval Station.
4	Kalsin Bay	70	?	U.S. Geological Survey stream gage.
5	Kodiak Naval Station.	63	Up	Fleet Weather Central, Kokiak Naval Station.
6	Kodiak	45±3	Down	Jerry Tilley.
7	Rocky Bay	[1] 30	Down	Guy Branson.
8	Seward	30±5	Up	Scottie McRae.
9	Whidbey Bay	19¼±½	Up	Bill Sweeney.
10	Puget Bay	20±2	Up	Sam Hatfield.
11	Middleton Island	[1] 20	Down	Dwight Meeks
12	Cape Yakataga.	60±1	Down	Charlie Bilderback.
13	Saltery Cove	[1] 30	?	Ron Hurst.

[1] Approximate.

belt about 6 miles wide (Plafker, 1968), and comparable displacements are known to have occurred on the adjacent sea floor (Malloy, 1964). Similarities in maximum wave-runup heights along physiographically comparable segments of coast, both on the Kenai Peninsula opposite Montague Island and on the ocean coast of Kodiak Island, suggest that the vertical sea-floor displacements that generated the waves in these two areas could be of the same order of magnitude. If so, the initial wave form resulting from sea-floor displacement had the approximate shape shown by profile *A–A′*, figure 15. Other factors, however, such as rate of uplift, initial slope at the wave source, and energy loss along the propagation path preclude direct correlation of runup heights with displacement at the source.

PROBABLE ZONE OF SLIGHT UPLIFT

Minor uplift, probably associated with the earthquake, has been detected in three areas adjacent to, and inland from, the zone of subsidence (fig. 3). The distribution of uplift in these three areas strongly suggests the possibility that they may be part of a continuous zone, roughly 100 miles wide, that parallels the major zones of subsidence and uplift.

Slight uplift in the Alaska Range is indicated by U.S. Coast and Geodetic Survey releveling along both the Richardson Highway and The Alaska Railroad (Small, 1966, fig. 9). Comparison of the 1964 leveling with a line run in 1952 shows general uplift along the Richardson Highway in a zone from about 25 miles north of Glennallen to within 50 miles of Fairbanks (fig. 3). Maximum uplift recorded on this line was 0.89 foot near the center of the zone, with irregular but generally progressive decreases toward the north and south. Land-level changes indicated by comparison of 1922 and 1964 levelings along The Alaska Railroad are less consistent but they are predominantly positive and are as much as 0.36 foot in a zone where the line crosses the Alaska Range. The relatively large amount of change (0.89 ft) in only 12 years between the successive surveys on the Richardson Highway strongly suggests that at least the major part of the measured changes were probably associated with the earthquake. Furthermore, the fairly systematic rise and fall in the amounts of uplift across the Alaska Range suggest that uplift is not due to surveying errors or errors inherent in tying the level lines to tidal datum planes. Thus, it is tentatively concluded that the uplift along these two lines represents earthquake-related uplift over a broad zone centered in the general area of the Alaska Range.

Residents along the Iliamna, Chinitna, and Tuxedni Bays on the northwest shore of Cook Inlet (pl. 1; fig. 3) report a decrease in the height of tides after the earthquake that suggests shoreline uplift of 1–2 feet. There is little doubt about the validity of these estimates, particularly in Iliamna and Tuxedni Bays, where reference marks existed whose pre-earthquake relationship to tide levels were precisely known. Between 1963 and 1965 a slight uplift, presumably related to the earthquake, occurred also on the west side of Augustine Island (R. L. Detterman, oral commun., 1965) where beach berms have been lowered ½–1½ feet. The observed changes along the northwest side of Cook Inlet strongly suggest slight tectonic uplift of the shoreline during the earthquake. They cannot be explained by changes in the tidal characteristics due to regional subsidence of the entrance to Cook Inlet, inasmuch as deepening of the entrance would be expected to increase, rather than decrease, the height of the tides by facilitating diurnal movement of the tidal prism.

GEOMETRY OF THE DEFORMATION

The pattern of absolute vertical deformation associated with the earthquake is indicated by the isobase contours and profiles of plates 1 and 2 and is shown at a smaller scale in figure 3. The isobase contours may be pictured as the amounts of vertical displacement of an imaginary surface that was horizontal before the earthquake. The resulting map, therefore, is a special form of structure-contour map showing the configuration of the deformed surface. The maps and profiles indicate that the deformation occurred in three broad elongate warps, each of which is from 100 to 130 miles wide and has axes that roughly parallel the trend of the continental margin. Subsidence occurred in the middle warp and uplift in the adjacent warps on the seaward and landward sides. The zero isobases between the zone of subsidence and the adjacent zones of uplift are axes of tilt across which the sense of vertical displacement relative to the preearthquake position changes gradually. No abrupt changes of level have been found between the adjacent zones that would indicate vertical fault displacement between them.

ZONE OF SUBSIDENCE

The zone of subsidence is a synclinal downwarp whose axis is situated roughly along the crest of the coastal mountain ranges. The axis of subsidence plunges gently northeastward from the Kodiak Mountains and southwestward from the Chugach Mountains to a low of 7½ feet on the south coast of the Kenai Peninsula. In cross section the downwarp is strongly asymmetrical with an average tilt in the middle part of the deformed region of about 1 foot per 14 miles from the landward side towards the axis and a much steeper average tilt of 1 foot per 2–3½ miles on the seaward side. The prevailing simple synclinal form of the downwarp is broken only by the slight warping of the tilted surface near the axis of subsidence immediately north of Seward in an area where both triangulation and geologic data suggest the possibility of minor earthquake-related movement on a conspicuous north-south-trending lineament.

Apparent reduction in crustal volume within the zone of subsidence, as calculated by summing the average volumes included between successive isobase contours on plate 1, is about 29 cubic miles. Total area of the zone affected by subsidence is about 48,000 square miles, and the average amount of subsidence within it is roughly 2½ feet.

MAJOR ZONE OF UPLIFT

The major zone of uplift along the continental margin is a broad upwarp with a maximum amplitude of 15 feet, upon which is superimposed a narrow belt, less than 10 miles wide, in which there has been strong uplift associated with displacement on reverse faults. The axis of the uplifted zone trends southwestward from Montague Island, presumably to the area offshore from Sitkalidak Island in the Kodiak group. Maximum uplift along this axis on Montague Island is 38 feet and it may be as much as 50 feet on the sea floor.

The position of the axis of uplift to the northeast of Montague Island is uncertain; it may continue offshore from Hinchinbrook Island and the Copper River Delta to intersect the coast at Cape Suckling where 13 feet of uplift was measured. Because of the scarcity of data points in the Cape Suckling area, the shape of the deformed surface there cannot be closely defined and the possibility cannot be ruled out that the large amount of uplift there may reflect local warping or faulting.

The part of the upwarp available for observation in Prince William Sound has an irregular shape that suggests combined tilting and warping. As indicated by profiles $A–A'$ and $B–B'$ on plate 1, the landward slope outside the narrow belt of extreme uplift averages 1 foot per 2.1 miles northwest of Montague Island and only 1 foot per 7.4 miles north of Hinchinbrook Island. Local tilts as high as 1 foot in 185 feet occur within the belt of extreme uplift and surface faulting in southern Montague Island. The isobases in the central and southeastern part of Prince William Sound reflect a broad undulating platform 4–8 feet above its preearthquake position. In at least two areas of western Prince William Sound, local flattening or even reversals of slope are indicated.

Data on the configuration of the upwarped surface in the Kodiak Island area, although less conclusive, suggest northwestward tilting that is as steep as 1 foot per mile at Narrow Cape.

Little is known about the shape of that part of the upwarped zone that is seaward from the axis of uplift because only a few points are available for observation. The slope between the most southeasterly capes of Montague Island and Middleton Island 50 miles to the southeast averages 1 foot per 11 miles, but the shape of the surface in the water-covered area between these points is conjectural. Nor is it known whether the uplift seaward from Middleton Island dies out gradually toward the toe of the continental slope, as inferred on the profiles on plate 1, or whether it terminates abruptly in one or more faults or flexures on the slope.

The apparent increase in crustal volume within the major zone of uplift is much less certain than that involved in subsidence, because the distribution of uplift in extensive submarine areas must be inferred from the trend of isobase contours in the northeastern part of the zone and a few offshore control points. If the deformation has the general form shown by the profiles on plate 1, the volume increase would be approximately 89 cubic miles, or roughly three times the decrease in the zone of subsidence. Total area of the uplifted zone is inferred to be roughly 60,000 square miles. Average amount of uplift is about 6 feet, except in the narrow axial belt of uplift and faulting extending southwestward from Montague Island, where it is probably 30 feet or more.

PROBABLE ZONE OF SLIGHT UPLIFT

Where the broad slight upwarp landward from the zone of subsidence is crossed by level lines, its axis seems to be centered along the crest of the Alaska Range and its maximum indicated uplift is 0.89 foot on the Richardson Highway line and 0.35 foot on The Alaska Railroad line. The upwarp crossed by these two lines of leveling may be part of a continuous zone that extends into the Aleutian Range of the Alaska Peninsula where uplift of as much as 1.5 feet has been reported at several places in the Kamishak Bay-Tuxedni Bay area. A rough estimate of the apparent increase in crustal volume in the zone of slight uplift, based on an average uplift of 0.3 foot and an area of 24,000 square miles is about 1.0 cubic miles.

EARTHQUAKE FAULTS

Faults on land associated with the 1964 earthquake were found only at two localities on southwestern Montague Island in Prince William Sound and on the subsea continuation of one of these faults southwest of the island. Comparable faults entirely on the sea floor may have gone undetected. As far as could be determined no definite movement occurred along any other faults on land, although faulting at depth is suspected in some areas of unconsolidated surficial deposits characterized by linear zones of landslides or surficial cracks.

MONTAGUE ISLAND FAULTS

The location of, and displacement across, the earthquake faults on and near Montague Island are shown on plate 1. Their surface characteristics and tectonic significance are briefly summarized in the following paragraphs. In a separate volume of the Geological Survey's series of papers on the Alaska earthquake, they are described in more detail (Plafker, 1967b).

The longer of the two faults, the Patton Bay fault, is represented by a complex system, 22 miles long, of en echelon reverse faults and associated flexures with an average N. 37° E. strike. Surface dip of the fault is northwest at about 85° near its southern end and 50°–75° elsewhere along the scarp. Displacement on the fault is almost entirely dip slip—the northwest side upthrown relative to the southeast side. The maximum measured vertical component of slip is 20–23 feet, and maximum indicated dip slip is about 26 feet. A left-lateral displacement component of less than 2 feet near the southern end of the fault is probably a local phenomenon related to a change in strike of the fault that causes it to trend at an oblique angle to the N. 53° W. principal horizontal stress direction.

The Patton Bay fault system was traced by the U.S. Coast and Geodetic Survey (Malloy, 1964) for at least 17 miles on the sea floor southwest of Montague Island. Indirect evidence, from the distribution of large aftershocks associated with the earthquake and from the distribution of submarine scarps, suggests that the faulting on and near Montague Island occurred at the northeastern end of a reactivated submarine fault system. This system approximately coincides with the axis of uplift inferred from seismic sea waves between the southeast coast of Kodiak Island and Montague Island (fig. 15). The fault apparently dies out on its northwestern end, although the possibility cannot be ruled out that it is offset en echelon towards the southeast (in a righthanded sense) and continues northeastward offshore from Montague Island at least as far as Hinchinbrook Island.

The shorter of the two faults, the Hanning Bay fault, is a virtually continuous reverse fault with an average strike of N. 47° E. and a total length of about 4 miles. Dip of the fault is 52°–75° NW. at the surface. Displacement is dip slip except for a left-lateral strike-slip except for a left-lateral strike-of a foot near the southern limit of the exposure. The maximum measured vertical component of slip is 16⅓ feet near the middle of the fault, the indicated dip slip at that locality being about 20 feet.

The two reverse faults on Montague Island and the postulated submarine extension of the Patton Bay fault constitute a zone within which crustal attenuation and maximum known uplift occurred during the earthquake. Nevertheless, there are no significant lithologic differences in the rock sequences across them to suggest that these faults form major tectonic boundaries. Furthermore, their spatial distribution relative to the regional zone of tectonic uplift associated with the earthquake, to the earthquake focal region, and to the epicenter of the main shock suggests that they are probably subsidiary features, rather than the primary faults along which the earthquake originated.

OTHER POSSIBLE EARTHQUAKE FAULTS ON LAND

As far as could be determined, there are no other surface faults on land along which movement occurred during the earthquake. A careful search for renewed movement on known preexisting faults did not reveal any detectable surface displacements. Nor were any anomalous abrupt changes found in amounts of vertical movement along the coast or along level lines inland from the coast that would suggest significant displacement on concealed faults. All reports of suspected faulting that were checked in the field proved to be landslides or surficial cracks in unconsolidated deposits. It is reasonably certain that if additional faulting did indeed occur, its surface expression is far more subtle than that on Montague Island.

Some of the linear belts of concentrated surficial cracking and landsliding may reflect displacements on concealed faults. Foster and Karlstrom (1967, p. F24) suggested that movement on a concealed fault may have produced a northeast-trending linear belt of conspicuous surface fissures on the Kenai Lowland in the western part of the Kenai Peninsula. However, no evidence has been found for vertical displacements where the belt crosses the U.S. Coast and Geodetic Survey level line south of Anchorage, and there is a notable absence of aftershock activity along the postulated fault.

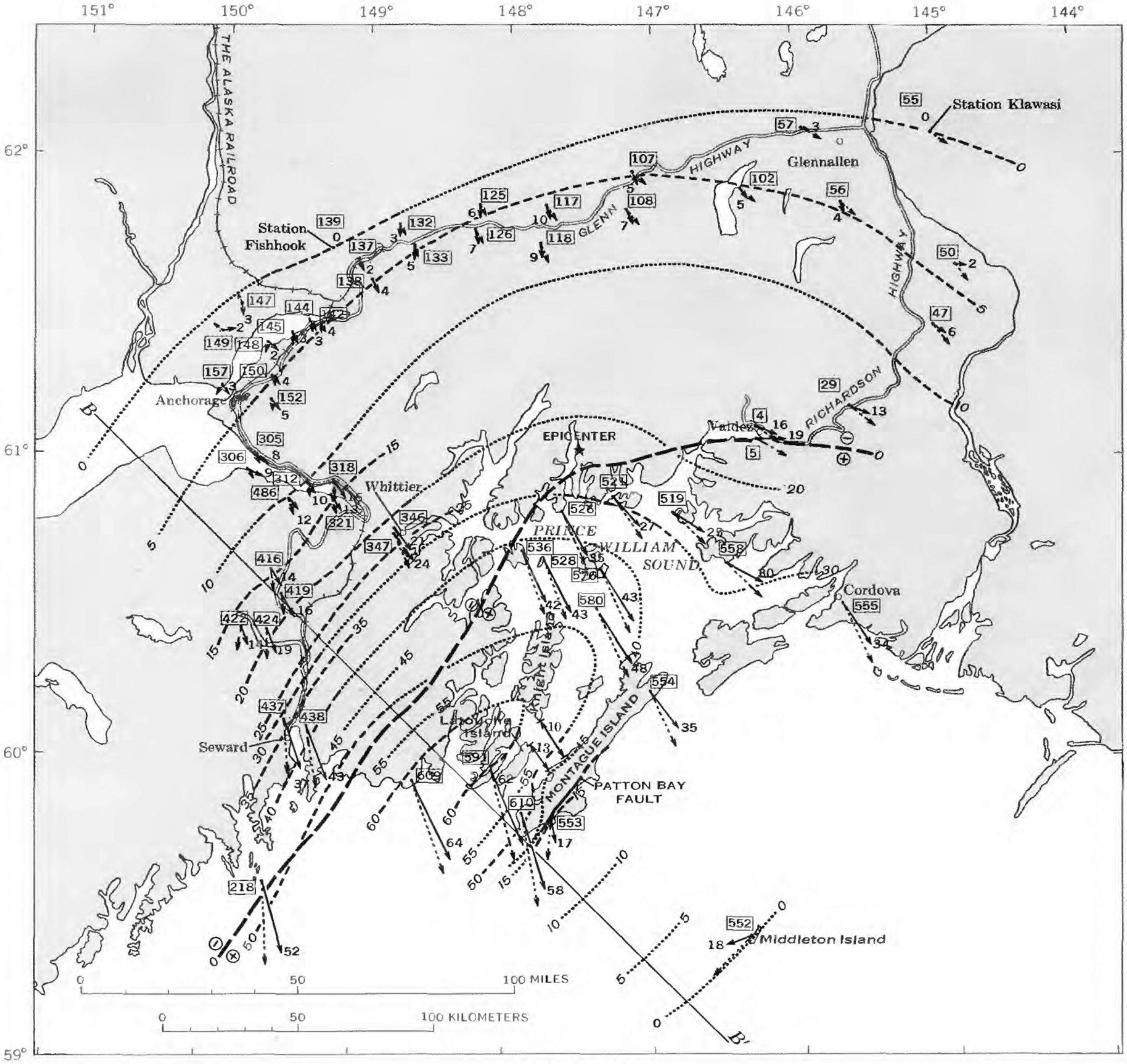
A second possible line of fault movement lies along the broad north-south-trending topographic depression, referred to here as the "Kenai lineament," that extends northward from Resurrection Bay through the valley containing the eastern arm of Kenai Lake (pl. 1). Faulting is suggested (1) by local concentrations of fissures seemingly unrelated to seismic shaking along The Alaska Railroad (D. S. McCulloch, oral commun., October 1967), (2) by reported angular changes between points on either side of the lineament as indicated by comparison of pre- and postearthquake triangulation surveys, and (3) by a distinct change in trend of isobase contours across the lineament (pl. 1). The geodetic data have been interpreted as suggesting left-lateral displacement of as much as 5 feet between stations about 4 miles apart on either side of the lineament (Wood, 1966, p. 122). These data, if correct, could indicate either slight movement on a north-south-trending concealed fault or crustal warping localized along the lineament.

HORIZONTAL DISPLACEMENTS

Although the vertical displacements that occurred during the earthquake are unusually large, they appear to be secondary to the horizontal displacements indicated by retriangulation over much of the deformed region. During 1964–65, the U.S. Coast and Geodetic Survey carried out revisional triangulations in the area shown in figure 16. The resurvey includes an area of about 25,000 square miles bounded on the west by the Seward-Anchorage highway, on the north by the Glenn Highway, and on the east by the Richardson Highway and the east coast of Prince William Sound. To the south, the resurvey extends to stations on the Gulf of Alaska coast and on Middleton Island 50 miles offshore from the coast. A tellurometer traverse was also run around the south coast of the Kenai Peninsula from Seward to Homer and from Homer to Moose Pass (at Kenai Lake) via Kenai. Because the precision of station locations obtained by the tellurometer traverse is probably too low to yield meaningful data on earthquake-related horizontal displacements, the stations are not shown in figure 16 and the indicated shifts of these stations are not considered here.

METHODS OF MEASUREMENT

Parkin (1966, p. 2–5) has described the procedures used in adjusting the pre- and postearthquake surveys. The preearthquake net consisted of: (1) a primary arc extending along the highway route from Anchorage northeastward to Valdez via Glennallen, surveyed in 1941 and 1944, (2) a second-order arc across the north shore of Prince William Sound from Valdez to Perry Island, surveyed in 1947–48, (3) a third-order arc surveyed from Perry Island to Anchorage between 1910 and 1914, (4) third-order triangulation between 1900 and 1961 for chart control across Prince William Sound and extending south to Middleton Island and westward along the southern Kenai Peninsula to Seward, and (5) a double arc from Seward north to connections at Turnagain Arm, surveyed by the U.S. Army Corps of Engineers in 1941–42. All these observations were combined into a single composite network and a



EXPLANATION

536

Triangulation station
Number refers to table 3

88

Direction of displacement relative to Stations Fishhook and Klawasi (adjustment 2, table 3)
Number is approximate change in feet

Direction of displacement relative to Station Fishhook (adjustment 1, table 3)

10

Relative northwest-southeast component of shortening indicated by resurvey of isolated segment of triangulation net
Number is measured change, in feet

30

Isothismic contours
Showing approximate southeastward component of horizontal displacement, in feet, relative to Stations Fishhook and Klawasi. Dotted where inferred

B——B′

Line of profile shown in figure 18

0 ⊖ ⊕ 0

Zero isobase between major zones of tectonic uplift and subsidence

0 60 FEET

Vector Scale
(Displacement between 0-5 feet shown as 5 feet on map)

16.—Map showing horizontal tectonic displacements in the Prince William Sound region and nearby areas. Horizontal displacements based on triangulation surveys by U.S. Coast and Geodetic Survey (Parkin, 1966, table 1).

free adjustment (an adjustment with no external constraints) was made in which one position—Station Fishhook—was held fixed. Internal scale and orientation for the net were furnished from 5 Laplace azimuths, 15 short taped base lines, and 1 tellurometer length, which were included in the adjustment as observation equations. The postearthquake triangulation survey, which was all first-order work, was adjusted in the same way as the earlier work.

Probable errors in the geographic positions of stations in southern Prince William Sound relative to the fixed station, as conservatively estimated by Parkin from the residuals, are 15–20 feet for the preearthquake survey and 6–8 feet for the postearthquake survey. These probable errors decrease progressively for stations closer to the fixed station.

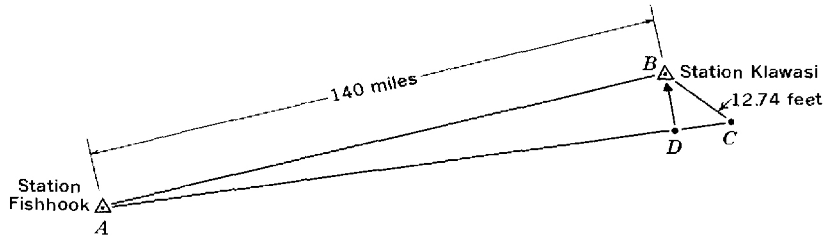
The horizontal shift of recovered stations relative to Station Fishhook between the pre- and postearthquake surveys, as computed by Parkin, are listed in table 3 as adjustment 1 and are shown graphically as displacement vectors (dashed) in figure 16. Because the postearthquake net was not carried northward to an area of stability, changes shown are relative rather than absolute. However, small angular shifts in the northern part of the net, as compared with those farther south, suggest that the northern part of the resurveyed net probably approaches an area that was not strongly affected by horizontal distortion during the earthquake. Anomalous aspects of the adjustment are (1) a gradual increase in displacement along the Glenn Highway arc east of Station Fishhook to almost 13 feet at Station Klawasi, and (2) an apparent 32-foot shift of Middleton Island southwestward in a direction almost normal to that of stations along the coast.

The first of these anomalies is eliminated, and the second considerably reduced, by an alternative preferred adjustment of the data in which two stations, Fishhook and Klawasi spaced 140 miles apart, are held fixed to provide additional orientation and scale of the postearthquake net (fig. 17). The assumption that the base line remained relatively stable in length and azimuth is justified on the basis of its position in the seismically inactive part of the net where there was only slight vertical displacement and by its orientation roughly parallel to the trend of isobase contours and normal to the trend of the horizontal shifts. The revised adjustment involves a

TABLE 3.—*Pre- and postearthquake differences in triangulation station plane coordinates*

[Data after U.S. Coast and Geodetic Survey (Parkin, 1966, and B. K. Meade, written commun., June 24, 1966). Adjustment 1: station 139 (Fishhook) held fixed; orientation and scale from azimuths and baselines of preearthquake net (Parkin, 1966, table 1). Adjustment 2: stations 139 (Fishhook) and 55 (Klawasi) held fixed for additional orientation and scale of preearthquake net. Station locations are shown in figure 16 except for stations 180, 184, 186, 188, and 218. Azimuth: north=0°.]

Station	Adjustment 1				Adjustment 2			
	Position shifts (feet)		Resultant vector		Position shifts (feet)		Resultant vector	
	ΔX (east (+)-west (−))	ΔY (north (−)-south (+))	Length (feet)	Azimuth	ΔX (east (+)-west (−))	ΔY (north (−)-south (+))	Length (feet)	Azimuth
4	+18.74	−13.73	23.23	125°	+15.51	−4.65	16.18	117°
5	+21.66	−11.50	24.52	120°	+18.60	−2.32	18.72	97°
29	+17.32	−15.39	23.17	130°	+12.37	−5.23	13.43	113°
47	+11.17	−14.32	18.16	140°	+4.17	−4.73	6.30	139°
50	+8.45	−11.84	14.55	145°	+1.96	−0.54	2.03	105°
55 (Klawasi)	+10.06	−7.82	12.74	130°	0	0	0	----
56	+10.08	−10.47	14.53	135°	+2.23	−3.29	3.97	146°
57	+10.74	−6.82	12.72	120°	+2.25	−1.23	2.56	119°
102	−10.05	−9.03	13.51	130°	+3.34	−3.60	4.91	137°
107	+8.47	−7.73	11.47	130°	+2.99	−4.16	5.12	144°
108	+8.01	−10.23	12.99	140°	+3.14	−6.13	7.05	154°
117	+5.26	−8.88	10.32	150°	+1.57	−6.20	6.40	166°
118	+4.64	−12.48	13.31	160°	+1.62	−9.30	9.44	170°
125	+3.83	−7.09	8.06	150°	+1.10	−5.41	5.52	169°
126	+3.88	−9.23	10.01	155°	+1.54	−7.29	7.45	168°
132	+1.46	−3.91	4.17	160°	+0.17	−3.12	3.12	177°
133	+1.45	−6.61	6.77	170°	+0.23	−5.40	5.40	178°
137	+0.83	−2.37	2.51	160°	+0.81	−1.81	1.97	157°
138	+1.55	−4.74	4.99	160°	+1.42	−3.75	4.01	159°
139 (Fishhook)	0	0	0	----	0	0	0	----
142	+0.58	−4.71	4.75	175°	+1.93	−3.76	4.22	153°
144	+0.40	−3.68	3.70	175°	+1.82	−2 90	3.42	148°
145	−0.49	−3.41	3.45	190°	+1.35	−2.70	3.02	153°
147	−1.84	+3.15	3.65	330°	+0.37	−2.61	2.64	172°
148	−0.51	−2.04	2.10	195°	+1.88	−1.53	2.42	129°
149	−2.46	+2.06	3.21	310°	+0.42	+1.68	1.73	76°
150	−0.07	−4.38	4.38	180°	+2.70	−3.27	4.24	140°
152	+0.58	−5.56	5.59	175°	+3.70	−4.18	5.25	143°
157	−1.54	−2.43	2.88	210°	+2.19	−1.95	2.93	132°
218	+5.08	−58.20	58.42	175°	+15.34	−49.67	51.98	163°
305	+2.77	−4.83	5.57	150°	+6.99	−2.80	7.55	112°
306	+3.29	−5.50	6.41	150°	+7.79	−3.40	8.50	114°
312	+2.96	−10.69	11.09	165°	+6.80	−7.51	10.11	138°
318	+3.26	−15.58	15.92	170°	+6.66	−12.07	15.81	140°
321	+3.94	−14.22	14.76	165°	+7.58	−10.51	12.96	144°
346	+9.68	−21.89	23.93	155°	+12.96	−16.78	21.20	142°
347	+10.72	−24.23	26.50	155°	+14.12	−19.00	23.67	143°
416	+2.63	−14.85	15.08	170°	+8.26	−10.89	13.67	143°
419	+2.64	−18.23	18.42	170°	+8.46	−13.89	16.26	149°
422	−1.08	−17.25	17.28	185°	+5.88	−12.91	14.19	156°
424	+1.85	−21.59	21.67	175°	+8.35	−16.88	18.83	154°
437	+5.34	−41.20	41.54	175°	+12.88	−34.86	37.16	160°
438	+8.84	−46.34	47.18	170°	+16.26	−39.65	42.85	158°
486	+4.09	−11.98	12.66	160°	+8.41	−8.89	12.24	137°
519	+22.13	−22.78	31.76	135°	+21.22	−13.76	25.29	123°
521	+19.46	−28.40	34.43	145°	+18.18	−20.46	27.37	138°
526	+18.01	−37.17	41.30	155°	+18.70	−29.84	35.22	148°
528	+18.19	−45.85	49.33	160°	+19.76	−38.18	42.99	153°
536	+16.68	−44.77	47.78	160°	+18.67	−37.72	42.09	154°
552	−21.56	−22.61	31.24	225°	−17.02	−6.32	18.16	250°
553	−3.14	−28.77	28.94	185°	+2.34	−17.05	17.31	172°
554	+18.63	−39.50	43.67	155°	+20.81	−28.26	35.09	144°
555	+20.85	−41.30	46.26	155°	+19.35	−28.32	34.29	146°
558	+25.94	−23.87	35.25	135°	+25.18	−13.30	29.45	118°
576	+22.24	−45.30	50.46	155°	+23.51	−36.48	43.40	147°
580	+23.59	−48.63	54.05	155°	+25.25	−39.43	47.82	147°
591	+18.60	−67.40	69.92	165°	+23.88	−57.59	62.34	157°
609	+22.06	−66.56	70.19	160°	+28.91	−57.49	64.34	153°
610	+11.30	−66.35	67.30	170°	+17.12	−55.23	57.82	163°



17.—Schematic diagram illustrating the method of deriving triangulation adjustment 2 from adjustment 1. According to adjustment 1, the postearthquake position of line *AB* is given by *AC*; point *B* shifted 12.74 feet S. 50° E. to point *C*. For adjustment 2, which assumes no change in distance or azimuth between *A* and *B*, the postearthquake net was rotated counterclockwise 0.0000124277 radians (angle *BAC*) and reduced in scale by the amount *DC*, or a factor of 0.9999881256 (the ratio of *AB*/*AC*).

counterclockwise rotation of the earthquake net of 0.0000124277 radians and a decrease in scale by a factor of 0.9999881256—changes probably well within the limits of error of these surveys. The resulting horizontal shifts, which appear to be more consistent with the vertical displacements, are given in table 3 as adjustment 2 and are plotted vectorially in figure 16 as solid lines. Unless otherwise specified, horizontal displacements referred to in the following sections are those of adjustment 2.

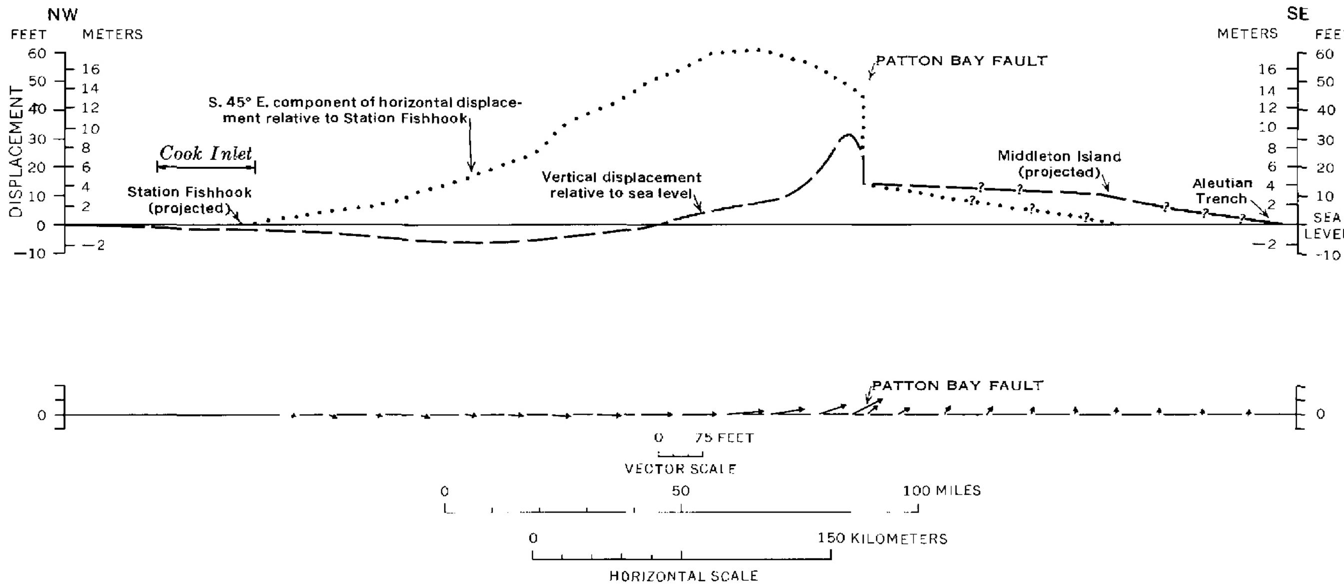
AMOUNT AND DISTRIBUTION OF THE DISPLACEMENTS

Absolute magnitudes and precise directions of the horizontal changes cannot be determined because the preearthquake triangulation net consisted mainly of third-order surveys and because the postquake survey, all of which was precise first-order work, was not carried northward to an area unaffected by the earthquake. Nevertheless, most of the changes are so large and systematic that there can be little doubt that they are in the general direction and are of the order of magnitude indicated by comparison of the two surveys. The true orientation and amount of displacement of the stations on the southeast shore of Montague Island (553) and on Middleton Island (552) are especially uncertain. This uncertainty exists because (1) both stations are in a part of the net where large differential earthquake-related vertical movements may have caused significant horizontal shifts in their positions, (2) the preearthquake triangulation involving these stations was only third order and the stations were tied to the net in Prince William Sound through several figures that are geometrically weak, and (3) the stations are situated near the extremity of the net where errors in displacement relative to the fixed stations are likely to be at a maximum. As a consequence, errors inherent in the adjustments could equal or exceed the observed displacements of these two stations in either of the two alternative adjustments.

The pattern of horizontal displacements relative to stations Fishhook and Klawasi during the time between the surveys is brought out by the displacement vectors (solid) in figure 16. Except for Middleton Island (552), they show relative seaward movements that are predominantly toward the south-southeast in the western part of the area, almost due southeast in the central part, and east-southeast in the eastern part. Over the central part of the net, the magnitude of the displacements relative to the base line increases progressively from the base line to a maximum of 64 feet at station 609 on the mainland immediately west of Prince William Sound, after which it decreases towards the southeast. In the western and eastern parts of the net, displacements show a progressive increase in magnitude to the most seaward stations amounting to as much as 52 feet south-southwest from Seward (218) and 34 feet near Cordova (555). In addition, resurveys of small isolated triangulation nets spanning the straits from Latouche and Knight Islands to Montague Island indicate relative shortening of 10–13 feet in a northwest-southeast direction (fig. 16).

The overall pattern of movement relative to the fixed stations is emphasized by the isothismic contours (lines of equal horizontal movement) in figure 16 which show the approximate component of horizontal displacement in a S. 45° E. direction, or nearly parallel to the average trend of the vectors in the same area. Contours are based on the displacement vectors (adjustment 2, table 3) and on relative horizontal movements within an isolated segment of the triangulation net between Montague, Latouche, and Knight Islands (Parkin, 1966, p. 9; C. A. Whitten, written commun., 1965).

Isothismic contours in figure 20 indicate that the entire area from the northern arc of the net to southwestern Prince William Sound showed a relative extension in a seaward direction, whereas the part of the net southeast of the Knight Island-Latouche Island area showed a relative shortening northwest-southeast direction. In other words, the position of Middleton Island, and perhaps the area southeast of the island,



18.—Profile showing measured and inferred tectonic displacements (above) and vectorial sum of the horizontal and vertical movements (below) along section *B–B′* of plate 1 and figure 16.

remained essentially fixed relative to the base line, whereas the intervening area was displaced in a relative seaward direction, the amount of displacement attaining a maximum in the Latouche-Knight Island area.

Triangulation data indicating shortening across Montague Island agree well with the observed imbrication on reverse faults in this area (Plafker, 1967b, p. G40–G41). However, the exact amount of shortening is uncertain because the geographic positions of the stations on the seaward side of Montague Island (553) and on Middleton Island (552) are subject to large errors that may equal or exceed the indicated amounts of displacement at these stations. If the S. 45° W. component of displacement dies out at Middleton Island, as inferred in figure 16, the average contraction between that point and the 60-foot isothismic contour is 60 feet. That this amount of contraction may not be unreasonable is indicated by (1) the 10- to 13-foot shortening in a northwest-southeast direction indicated by reobservation of the small isolated triangulation net spanning Montague Strait between Montague Island and Latouche and Knight Islands, (2) the horizontal shortening of at least 9.3 feet, and possibly as much as 19 feet, across the Patton Bay fault that is indicated by surface mapping, and (3) the pronounced crustal warping that occurred on and near Montague Island.

RELATIONSHIP TO REGIONAL VERTICAL DISPLACEMENTS AND SURFACE FAULTS

A genetic relationship between the horizontal and vertical regional displacements is strongly suggested by the orientation of the horizontal displacement vectors in a direction roughly normal to the trend of the isobases and by approximate coincidence of the maximum vertical displacements with areas of maximum transverse extension or contraction. This relationship is brought out by the profiles in figure 18 which show magnitudes of the horizontal displacements in a relative S. 45° E. direction and the vertical displacements relative to sea level along line *B–B′* of figure 16. Also shown in figure 18 are the vectorial sums of the horizontal and vertical displacements along the line of profile, that is, the direction and relative amount of movement of points on the ground surface along this line.

The horizontal displacement data indicate that the zone of subsidence extended tranversely by an average of 1.1×10^{-4}, or 1.1 parts in 10,000, and reached a maximum of about 3×10^{-4} slightly seaward from the axis of the subsided zone. By contrast, at least part of the zone of uplift seems to be one of net transverse shortening resulting from crustal warping and reverse faulting. Average contraction across the uplifted zone as far seaward as Middleton Island is about 10^{-4} and it averages as much as 8×10^{-4} across the narrow belt of maximum uplift on Montague Island. Presumably, a comparable relationship exists between horizontal displacements and the earthquake-related vertical movements that ocurred outside the retriangulated area. Extension of the retriangulation net over this area could provide a definitive test of this assumption.

In a general way, the displacement vectors on either side of the Patton Bay fault (stations 610 and 553) are consistent with the field observations that the fault has undergone reverse movement with resultant crustal shortening by imbrication in the dip direction. In detail, however, there is an unresolved discrepancy between the observed dip-slip movement on the Patton Bay fault and the apparent left-lateral strike-slip shift of triangulation stations on either side of it (Plafker, 1967b, fig. 35). The discrepancy was reduced by the readjustment (adjustment 2, table 3) used here, but not altogether eliminated. Absence of an observable component of lateral slip on the fault suggests either that the displacement was taken up largely by horizontal distortion between the fault and the two triangulation stations or, more probably, that an error has been introduced into this part of the triangulation adjustment through a slight clockwise rotation of displacement vectors.

It is significant that, regardless of the details of the horizontal displacements, the triangulation data suggest rebound of a broad segment of the continental margin that had been elastically compressed and shortened by at least 64 feet prior to the earthquake. The vectors in figure 16 show the general sense and amount of the rebound within the retriangulated area. This indicated rebound implies preearthquake regional compression oriented parallel to the trend of the vectors, or roughly normal to the continental margin and trend of the eastern end of the Aleutian Arc.

TIME AND RATE OF THE DEFORMATION

Instrumental records of the time and rate of tectonic movements in the deformed area are nonexistent. Three standard tide gages at Seward, Kodiak, and Homer were located where they might have been able to record the vertical land movements relative to sea level had they been operative during the earthquake. However, the Seward gage was destroyed in a submarine landslide at the time of the earthquake, the Kodiak gage with the marigram for the month of March was lost when it was washed away by seismic sea waves half an hour after the earthquake, and the Homer gage was made inoperative by the shaking. There were no accelerographs in the affected region to record the horizontal movements. As a consequence, the time and rate of the movements can only be inferred from the reports of eyewitnesses, from photographs taken after the earthquake, and from the water and atmospheric disturbances generated by the movements.

EARTHQUAKE-RELATED MOVEMENTS

Numerous eyewitness reports of immediate withdrawals of water from uplifted coastal areas indicate that much, if not all, of the deformation occurred during the $1\frac{1}{2}$–5 minutes of violent tremors. In most places, however, immediate water disturbances resulting from submarine slides or other causes precluded estimates of relative changes in level for several hours or days after the earthquake. In an area uplifted 6.3 feet, one eyewitness (Gordon McMahan, oral commun., 1964), thought that the displacements were perceptible as a series of distinct upward accelerations during the earthquake. Another eyewitness (Guy Branson, oral commun., 1964), from an area that subsided 5 feet, described a definite dropping or sinking sensation toward the end of the strong ground motion "as when a plane hits an air pocket." No other observers reported perceptible accelerations in the direction of the tectonic displacements.

All of the uplift and surface faulting at the southwest tip of Montague Island occurred prior to March 30th. On this date the uplifted platform at Cape Cleare and a part of the Patton Bay fault were photographed during a reconnaissance flight (fig. 20). Jim Osborne, a bush pilot who knows the Prince William Sound area intimately and is an exceptionally perceptive observer, informed me that all of the shoreline displacements took place prior to the morning of March 28—the day he first flew over the area after the earthquake. According to Osborne, there were no noticeable shoreline changes after the 28th. His evaluation is corroborated by residents along the coast in all areas affected by the tectonic displacements. Movement along strongly uplifted shores occurred at least fast enough to trap many mobile marine animals such as small fish, starfish, and snails above the tide level (fig. 24; p. I 36).

That a substantial fraction of the net vertical displacement occurred very rapidly is also suggested by the pattern of seismic air and sea waves. The peaks between compression and rarefaction on the La Jolla microbarograph record (Van Dorn, 1964, fig. 5) were 7 minutes apart, a difference which suggests a peak-to-peak separation at the origin of about 83 miles; this figure is in close agreement with the observed spacing between the axes of uplift and subsidence. As noted by Van Dorn, the recorded disturbance could only have been produced by vertical motions over a very large area, and in a time interval of the order of that required for an acoustic wave to propagate across the dimensions of the generator. The elapsed time between the earthquake and the arrival of the initial wave crest along the ocean coast of the Kenai Peninsula further suggests that the initiating disturbance along the submarine extension of the axis of maximum uplift southwest of Montague Island (fig. 15) occurred during, or within a few minutes after, the earthquake.

There is no direct evidence as to when the horizontal displacements, which in some inhabited localities were as much as 60 feet, occurred. No observers reported strong systematic horizontal movements at any time during the main shock, nor could such movements be inferred with confidence from the incomplete data on the directions in which objects or structures fell. Nevertheless, as suggested on page I39, horizontal displacements probably occurred during the earthquake, and at a rate fast enough to cause waves in some bodies of surface water. Accelerations due to the permanent displacements probably were undetected by observers because they were masked by the strong ground motions resulting from the transient elastic seismic waves.

PREEARTHQUAKE MOVEMENTS

Vertical changes in the position of the shore relative to sea level have been noted within a period of hours prior to some major earthquakes in Japan (Imamura, 1930, p. 141). These changes, which have been termed "acute" tiltings or deformations by Japanese scientists, have been a subject of special interest because of their obvious potential importance in earthquake prediction.

During the field investigation of the 1964 Alaska earthquake, an effort was made to ascertain whether any premonitory changes of level were noted by residents in coastal areas or were recorded on operative tide gages. The only suggestion of preseismic changes was an observation made by an officer of Fleet Weather Central at the Kodiak Naval Station to the effect that tides in the area were at least 1½, and possibly 2½, feet lower than normal a few days before the earthquake and that the low tides were apparently unrelated to atmospheric conditions (Lt. C. R. Barney, oral commun., 1964). However, the loss of the March marigram prevented documentation of the reported low tides. The Seward and Homer marigrams for the time preceding the earthquake do not show evidence of preseismic changes, nor have such changes been reported elsewhere by coastal residents.

POSTEARTHQUAKE MOVEMENTS

Relevelings, tidal observations, and gravity readings suggest either no postearthquake vertical changes or, perhaps, slight changes in the earthquake-affected region.

The most convincing indication of continued postearthquake movement comes from releveling in May–June and in October 1964 of a line 22 miles long extending northwestward from Portage on Turnagain Arm and a third releveling from Portage to Anchorage in the summer of 1965. Between the preearthquake leveling and the initial postearthquake leveling, Portage subsided 5.6 feet, the area 22 miles to the northwest subsided about 4.9 feet, and Anchorage subsided about 2.3 feet (pl. 1). Comparison of the two 1964 relevelings shows a progressive increase in divergence from northwest to southeast, which suggests additional relative subsidence of about 0.16 foot at Portage in the period between surveys (Small, 1966, p. 13). Comparison of the May–June 1964 and the 1965 leveling suggests relative postearthquake subsidence of 0.36 foot at Anchorage and 0.52 foot at Portage during this interval (Small, 1966, p. 17). Unfortunately, neither the October 1964 line nor the 1965 line was tied to tidal bench marks, so the absolute postearthquake displacements are uncertain. Furthermore, because both Anchorage and Portage are situated in areas of extensive thick unconsolidated deposits, the possibility cannot be ruled out with the data available that some or all of the indicated subsidence may be due to continued consolidation of soft sediments.

Small (1966, p. 18) also reports a gravity increase of about 0.18 mgal on Middleton Island relative to an Anchorage base station. This increase occurred between the time of a postearthquake 1964 measurement and one made in 1965 which would indicate about 2 feet of additional uplift between surveys. However, the possibility that this large difference in the successive gravity readings may be due to meter drift in one or both surveys is suggested by the fact that residents of the island did not notice changes in relative tide levels during this same interval. Two feet of uplift at Middleton Island should have been readily detectable along the shore.

A comparison of tidal observations made in 1964 and 1965 provides data on the postearthquake land-level changes at 14 of the stations listed in table 1. However, it is difficult to separate purely tectonic movements from meteorological effects and the effects of surficial compaction at gages situated on soft sediments. Tidal observations in the zone of uplift at Cordova, Port Chalmers, and Sawmill Bay showed no detectable change suggestive of continued tectonic movements, but one station, Port Gravina, apparently subsided 0.3 foot between 1964 and 1965. In the zone of subsidence, gages at Seward and Port O'Brien had no detectable change in mean sea level; five gages at bedrock sites showed slight rises ranging from 0.1 to 0.5 foot, possibly suggestive of postearthquake tectonic uplift. Comparisons of 1964 and 1965 tidal observations at Valdez, Whittier, and Homer, in the zone of subsidence, indicated apparent continued subsidence ranging from 2.6 feet at Valdez to 0.3 foot at Whittier. The postearthquake subsidence at Valdez is definitely related to seaward extension and subsidence of the thick prism of deltaic deposits on which the tide gage is situated; much or all of the subsidence at the other two sites, both of which are on thick deposits of unconsolidated sediment, could also have resulted from surficial effects.

The available data on postearthquake changes outlined above are internally inconsistent and inconclusive with reference to postearthquake vertical movements. Disregarding the Valdez, Whittier, and Homer stations, where superficial subsidence of unconsolidated deposits is known or suspected to be large, the repeated postearthquake tidal observations indicate either recovery (by uplift) of as much as 0.5 foot or no change in the subsided zone. However, the repeated levelings on the Portage-Anchorage line and repeated gravity readings at Anchorage have been interpreted as indicating either continued subsidence or stability in that part of the zone of subsidence. On the other hand, tidal observations in the zone that was uplifted during the earthquake suggests either postearthquake subsidence of as much as 0.3 foot or stability, whereas the pair of gravity measurements at Middleton Island in this zone suggest additional uplift of about 2 feet. Repeated tidal observations, levelings, and gravity readings over a longer time period will be required before definite conclusions may be drawn concerning the postearthquake pattern of adjustments in the deformed region. It is abundantly clear, from available data, however, that there was no large rapid postearthquake recovery of vertical displacement comparable to the recoveries reported after some major earthquakes along the coasts of Japan and South America.

EFFECTS OF THE TECTONIC DISPLACEMENTS

Regional vertical tectonic displacements, both upward and downward, have caused profound modifications in shoreline morphology and attendant widespread effects on the biota. Changes in the position of the shorelines relative to sea level directly affected numerous coastal installations, shipping, and the fishing and shellfish industries. A major indirect effect of the vertical movements was the generation of a train of destructive seismic sea waves that were responsible for 35 of the 115 fatalities and for much of the property damage attributable to the earthquake. The movements also appear to have generated atmospheric and ionospheric disturbances that were detectable at several places in the conterminous United States.

The systematic regional horizontal displacements may have caused waves in certain confined and semiconfined bodies of surface water, and related porosity changes may have caused temporary water losses from surface streams and lakes as well as drops in water levels of some wells that tap confined aquifers.

Because the displacements were along faults that are under water and in uninhabited places on land they did not damage any works of man. Had they occurred in inhabited areas, however, these displacements surely would have caused extensive damage to structures built across them. It is also reasonably certain that phenomena related to the reverse faulting, such as the landsliding, extension cracking, and severe warping that occurred in a belt as much as 3,000 feet wide adjacent to the fault traces (Plafker, 1967b), would have been a definite hazard for engineering works.

Most of the effects resulting from vertical movement of the shoreline have been known from other earthquakes in coastal areas throughout the world. Especially detailed descriptions have been given by Tarr and Martin (1912) of the various physiographic and biologic effects of uplift and subsidence associated with the great earthquakes of 1899, centered near Yakutat Bay along the Gulf of Alaska coast. Effects of such movements on the works of man have also been amply documented for numerous major earthquakes along the coasts of South America, New Zealand, India, Japan, and elsewhere, most of which have been summarized by Richter (1958). Although submarine tectonic movements have long been suspected as the most probable generative mechanism for seismic sea waves, the 1964 Alaska earthquake provides what is probably the clearest evidence for a cause-and-effect relationship between these two phenomena. Atmospheric disturbances of the type associated with the 1964 earthquake have been recorded previously after large volcanic explosions and nuclear detonations, but they have never before been observed in association with tectonic earthquakes. To the writer's knowledge, there are no published reports relating surface-water disturbances or ground-water changes to horizontal tectonic displacements during previous earthquakes.

19.—Spruce trees on a spit near the mouth of Resurrection Bay killed by salt-water immersion and undermined by erosion after the land subsided about 3 feet. Photograph taken at a 9-foot tide stage, July 10, 1964.

PHYSIOGRAPHIC CHANGES

Tectonic subsidence, augmented locally by surficial subsidence of unconsolidated deposits, resulted in narrowing or, in extreme cases, complete submergence of beaches. Sea water inundated the lower reaches of some streams in subsided areas as much as 4,500 feet inland from the former mouths, and salt water encroached upon former beach-barred lakes at stream mouths or bay heads (Plafker and Kachadoorian, 1966, p. D27). Beach berms and deltas in subsided areas rapidly shifted landward and built up into equilibrium with the new, relatively higher sea levels (pl. 4*B*). Former reefs and low-lying islands along the coast were submerged, and some tombolo-tied points or capes became islands. Wave action at the higher sea levels caused rapid erosion of shorelines—especially those composed of poorly consolidated deposits that were brought within reach of the tides (fig. 19). An irreplaceable loss resulting from such accelerated erosion of these deposits was destruction of coastal archaeological sites at several places in the Kodiak Island group and on the southern Kenai Peninsula.

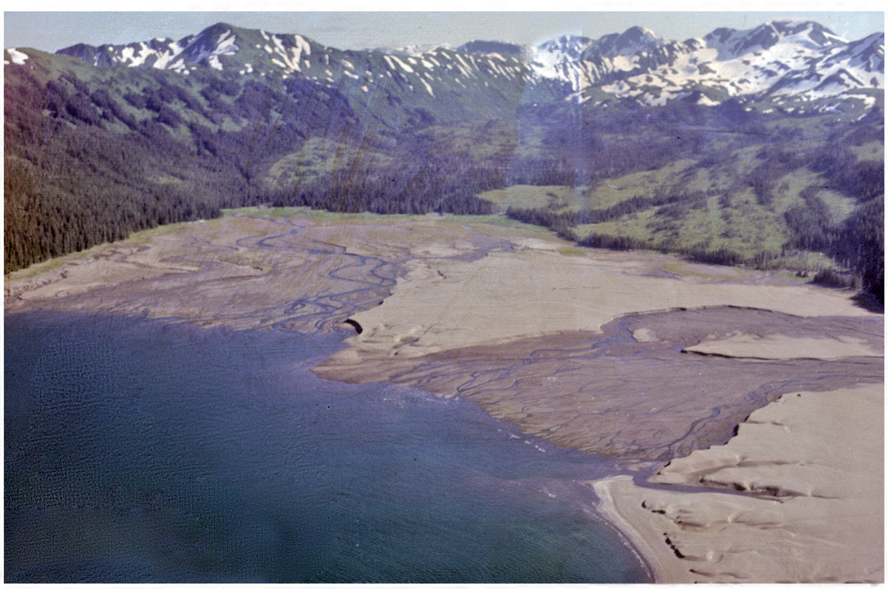
The major effect of tectonic uplift was to shift the extreme high-tide line seaward and thereby expose parts of the littoral and, at some places, the sublittoral zones (frontispiece; figs. 14, 20). In the areas of maximum uplift on southwestern Montague Island, the emergent sea floor is as much as 1,800 feet wide (Plafker, 1967b, pl. 1, 2). As a consequence, former beaches and sublittoral marine deposits were rapidly incised by streams that cut down through them to new, relatively lower base levels (fig. 21). In many places, beach-barred lakes were drained in varying degrees by incision of their outlet streams. About 8 or 9 feet of uplift at the outlet of shallow Bering Lake, which formerly was reached by high tides, caused the lake to be suddenly reduced in area by about 4 square miles to a third its preearthquake size. Beaches and deltas developed below, and seaward from, their previous positions (fig. 14). Along the uplifted shores, preearthquake beaches, sea cliffs, driftwood lines, sea caves, notches, stacks, and benches were elevated above their normal position relative to sea level. Similarly, in offshore areas, uplift created new islands and exposed reefs at stages of tide when they formerly were under water.

TILTING OF LAKE BASINS

Regional tilting or warping of the land surface seems to have caused permanent shoreline changes at Kenai and Tustumena lakes on the Kenai Peninsula. It may have had comparable effects on other lakes for which observational data are unavailable.

Tilting of Kenai Lake, which is about 25 miles long, is indicated by changes in the relative position of the bench marks that had been established near its ends prior to the earthquake. Although the accuracy of some of the recovered bench mark positions is open to question, the postearthquake survey suggests that the western end of the lake sank 3.0 feet with respect to the east end, and that the dip of the tilted surface is N. 72° W. at 1 foot per 5.4 miles (McCulloch, 1966, p. A29). These data are corroborated by the fact that the west end of the lake is close to the axis of subsidence (pl. 1) and that residents report a relative lower-

20.—Rocky surf-cut platform a quarter of a mile wide at Cape Cleare, Montague Island, exposed by 26 feet of tectonic uplift. The white band on the upper part of the platform consists mainly of barnacles and calcareous worm tubes; brown algae, or "kelp," cover much of the surface below the barnacle zone. Photograph taken at about zero tide stage, March 31, 1964. Compare with frontispiece, taken 2 months later in same general area.



21.—Bay-head deposits in MacLeod Harbor, Montague Island, deeply incised by stream erosion following about 33 feet of uplift. Arrows indicate the positions of pre- and postearthquake high-tide shorelines. Photograph taken August 6, 1965.

ing of the lake level at the eastern end after the earthquake (McCulloch, 1966).

The long axis of the 20-mile-long Tustumena Lake and its outlet stream, the Kasilof River, are oriented northwest-southeast, or almost normal to the projected trend of isobase contours in the area (pl. 1). After the earthquake, water levels at the inlet end of the lake reportedly rose above the banks; about 2 feet of southeastward tilt of the lake basin is thus suggested (J. D. Reardon, oral commun., 1965). The amount of tilt across the basin, as indicated by reported relative changes in lake levels, is in good agreement with that suggested by the spacing of isobase contours projected from the coast into the Kenai Lowland area (pl. 1).

TILTING OF RIVER DRAINAGES

Regional tilting may also have temporarily reduced the flow of certain rivers, such as the Copper, Kenai, and Kasilof Rivers, whose flow directions were opposite to the regional tilt (pl. 1). The Kasilof River was reduced to a trickle the day after the earthquake (Alaska Dept. Fish and Game, 1965, p. 23) and the Copper River reportedly ceased flowing at its mouth for several days. Immediately after the earthquake the Kenai River for almost a mile below the Kenai Lake outlet temporarily reversed its direction and flowed back towards Kenai Lake (McCulloch, 1966, p. A28), but it is not clear to what extent this reversal was due to tilting and to what extent it was related to the seiching of the lake.

Because rivers and lakes were approximately at their lowest annual levels when the tilting occurred, slight changes in gradient caused disproportionately large changes in discharge. The changes probably were largely related to upstream tilting of the larger lake basins in the drainage systems with consequent reductions or reversals of discharge until the basins once again filled to the spillover point. To some extent, however, the reduced flow in the river channels may have resulted from the lowered gradient of the beds. The regional tilting averaged 1 foot per 4.8 miles in the lower Copper River drainage and 1 foot per 10 miles or less in the Kenai Lowland. Other causes, such as channel blockage by river ice or landslides, may also have contributed to the reported temporary declines in discharge.

22.—Spruce trees in Nuka Passage on the southern Kenai Peninsula killed by repeated inundation with salt water in an area of 6.3 feet of tectonic subsidence. Algae and animals of the upper littoral have encroached upward into the former terrestrial environment. Inset shows a barnacle (white) and numerous *Littorina* or "periwinkles" (gray) on the roots of a tree. Photograph taken July 22, 1965.

BIOLOGIC CHANGES

Vertical displacements of the shoreline strongly affected both the fauna and the flora over a vast segment of coastal southcentral Alaska. Some of these effects were apparent within days after the earthquake; others, which depend upon the complex interrelations of one organism to another and to their habitat, will not be known for a long time. G Dallas Hanna, who studied the biologic effects of the earthquake in the littoral zone, has given a graphic summary of these earthquake-related changes (Hanna, 1964). The results of detailed governmental and private studies of the effects of the earthquake on intertidal organisms, land plants, and fish are to be reported in the Biology Volume of the planned series of publications of the Committee on the Alaska Earthquake of the National Academy of Sciences (W. L. Petrie, oral commun., 1968).

The most conspicuous effect of subsidence was the fringe of terrestrial vegetation killed by salt-water inundation at periods of high tides (pl. 4*B*; figs. 12, 19). Virtually all noncliffed shorelines that subsided more than 3 feet clearly showed fringes of dead vegetation within 2 months after the earthquake. In some sheltered localities at which vegetation extended down to the extreme high-tide line, dead vegetation was noticeable even where subsidence was as little as 1 foot.

Trees, bushes, beach grass, and muskeg along many former beaches were killed and partially buried in gravel or sand. Extensive areas of coastal marshland and forest that formerly had provided winter forage for grazing animals or nesting grounds for migratory birds were inundated. In such places, marine organisms encroached upward into the new littoral zone and it was not uncommon to find barnacles, limpets, and algae living on or among the remains of land plants (fig. 22).

The effects of subsidence on sessile intertidal marine organisms submerged below their normal growth positions were not readily apparent. Undoubtedly, individuals near the lower depth range of the species were adversely affected by the changed conditions and were gradually replaced by other organisms better adjusted to the deeper water environment.

Effects of uplift on the biota of the littoral zone were more striking than those resulting from subsidence, because the uplift caused complete extermination of organisms that were permanently elevated above their normal ranges. The width of the resultant band of dead organisms depended, of course, on both the amount of uplift and the slope of the uplifted shore. In areas where uplift exceeded the local tide range, as on islands in southern Prince William Sound, on parts of the mainland coast to the east of the Sound, and on several offshore islands on the Continental Shelf, destruction of the sessile organisms was almost absolute. Even many of the mobile forms—including starfish, gastropods, and small fish—did not survive. Some of the effects of uplift

23.—Closeup view of surf-cut surface at Cape Cleare, Montague Island, shown in frontispiece. The white coating on the rocks consists primarily of desiccated calcareous algae and bryozoans; the dark ropelike objects are stipes of laminarians ("kelp"). Photograph taken June 1, 1964.

on organisms of the littoral zone are illustrated by plates 3*A*, 4*A*; figures 5, 23, and 24. The dramatic change with time in the appearance of the shore and sea floor after about 26 feet of uplift at the southwest end of Montague Island may best be appreciated by comparing the aerial photograph taken on March 30th, 3 days after the earthquake (fig. 20), with one taken 2 months later on May 30th (frontispiece).

By August 1964 a few land plants had encroached onto the fringe of shore reclaimed from the sea, and in the summer of 1965 scattered clumps of grasses and wildflowers grew everywhere, on raised beaches and deltas and in favorable localities on rock benches amid the dead and dried remains of marine organisms (fig. 13). In a few years the bleak aspect of these fringes of uplifted shore should become subdued by a luxuriant cover of brush and timber comparable to that growing on older uplifted marine terraces in the area. By July 1965, land plants had covered much of the raised platform on Middleton Island and sea birds had already begun nesting in the former intertidal zone.

Throughout the uplifted areas in and near Prince William Sound, the mortality of all types of shellfish—including commercially important razor clams—has been estimated to be as high as 90 percent by G Dallas Hanna (oral commun., 1965). At many places where uplift exceeds the normal tide range, the clam population was literally wiped out. In such areas, the populations of birds, fish, and other animals that normally feed on shellfish must eventually readjust downward to the reduced food supply.

The potential effect of the land-level changes on the important salmon runs in the affected areas cannot be fully evaluated until the matured 1964 hatch returns from the sea to spawn. Spawning areas for pink and chum salmon, which are intertidal spawners, received major damage due to changes in land level and seismic sea waves (Alaska Dept. Fish and Game, 1965, p. 3; Thorsteinsson, 1965). Spawning areas of upstream migrants, including the red and silver salmon, where relatively unaffected by the earthquake.

Many low-lying coastal lakes that were important habitats for

24.—Mass of dead starfish in a depression on the uplifted platform shown on figure 23. Photograph taken May 31, 1964.

waterfowl and for fresh-water fish were damaged mainly by salt-water pollution related to subsidence or to draining in varying degrees resulting from uplift and incision of their outlet streams. A few of the uplifted lakes that formerly received salt water through their outlets during high tides become entirely fresh with consequent changes in the number and species of fish they can support.

GENERATION OF SEISMIC SEA WAVES

Most major earthquakes in coastal areas that involve vertical tectonic displacements beneath the sea are accompanied by seismic sea waves, and the 1964 earthquake generated one of the larger seismic sea waves of recent times (Grantz and others, 1964, p. 11–12; Van Dorn, 1964; Plafker and Mayo, 1965; Plafker and Kachadoorian, 1966; Pararas-Carayannis, 1967). Between the southern tip of Kodiak Island and Kayak Island, these waves took 20 lives and caused destruction all along the ocean coast. The waves were especially destructive along the ocean coast of the Kodiak group of islands and the Kenai Peninsula areas that had been lowered relative to sea level by tectonic subsidence or by the combination of tectonic subsidence and compaction of unconsolidated deposits during the earthquake. In addition, the waves, which were recorded on tide gages throughout the Pacific Ocean, caused 15 deaths and major damage in British Columbia, Oregon, and California.

The wave-source mechanism was initially investigated by Van Dorn (1964), who concluded that the waves were generated by a dipolar displacement of water resulting from regional tectonic warping. He inferred that the positive pole of this disturbance included much of the shallow Continental Shelf bordering the Gulf of Alaska within the major zone of uplift, and that the negative pole lay largely under land or beneath Cook Inlet and Shelikof Strait in the major zone of subsidence. From preliminary data on the amount and distribution of vertical displacements along the shore, Van Dorn (1964, p. 17) calculated that the total potential energy imparted to the positive part of the seismic sea wave by submarine uplift (assuming (1) vertical displacement of 6 feet that increases progressively from zero at the southwest end to 6 feet at the northwest end and (2) source dimensions of 240 miles by 100 miles), was 1.7×10^{14} ft-lbs (2.3×10^{21} ergs), or only about 0.01–0.05 percent of the approximately 10^{24} to 2×10^{24} ergs of seismic-wave energy released by the main shock. Pararas-Carayannis (1967), using source dimensions of 93 miles (150 km) by 435 miles (700 km) and the same average uplift as inferred by Van Dorn, arrived at a total water-wave energy of 5.88×10^{21} ergs.

The Geological Survey's subsequent studies of the vertical displacements on land and their probable extension beneath the sea provide additional data relevant to the probable configuration of the initial positive wave and its energy content. These data suggest that the initial wave form, due to vertical displacement of the sea floor on the Continental Shelf, probably had the general cross-sectional shape indicated by the profile in figure 15 and that the offshore areas involved in the uplift and the amounts of sea-floor displacement are considerably greater than was indicated by preliminary reconnaissance surveys. Thus, the general shape of the deformed surface on the Continental Shelf may be roughly approximated by a broad low-amplitude upwarp with minimum dimensions of 400 by about 75 miles, superimposed upon which is a narrow belt of maximum uplift about 6 miles wide that is inferred to extend some 350 miles southwestward from Montague Island. As indicated on profiles $A–A'$, $B–B'$ $C–C'$ plate 1, average uplift across the broad upwarp is roughly about 12 feet and that across the narrow zone is probably at least 30 feet. Because this highly simplified model does not consider the additional wave energy at the ends of the deformed region, where uplift gradually falls off to zero, or in that part lying seaward from the edge of the Continental Shelf where the deformation field is unknown, the calculated energy should be considered as a minimum.

If the initial wave form approximates that of the uplift, total potential energy transferred to the water, E_t was the sum of the energy in both the broad low-amplitude part of the wave (E_1) and in the narrow superimposed high-amplitude part (E_2). Total potential energy transferred to the water, E_t, derived by using Iida's equation (1963, p. 65), was

$$E_t = E_1 + E_2 = \frac{\rho g (h_1)^2 A_1}{2} + \frac{\rho g (h_2)^2 A_2}{2}$$

where ρ is the density of sea water, g is the gravitational acceleration, h is the average vertical displacement, and A is the area over which the deformation occurred. Therefore, by substitution,

$$E_1 = (1.1)(32)(15)^2(6)(350)(5{,}280)^2 = 4.6 \times 10^{14} \text{ ft-lbs } (6.2 \times 10^{21} \text{ ergs});$$

$$E_2 = (1.1)(32)(6)^2(75)(400)(5{,}280)^2 = 1.06 \times 10^{15} \text{ ft-lbs } (1.4 \times 10^{22} \text{ ergs});$$

and their sum,

$$E_t = 1.5 \times 10^{15} \text{ ft-lbs } (2 \times 10^{22} \text{ ergs}).$$

These figures suggest that the total potential energy in the positive part of the wave may be about an order of magnitude larger than that derived by Van Dorn, or 0.1–0.5 percent of the seismic wave energy release. According to the model used, roughly one-third of the energy was concentrated in the narrow high-amplitude part of the wave along the axis of maximum uplift and two-thirds was distributed over the low-amplitude part of the wave which has an area roughly 15 times larger. Thus, the relatively greater damage and higher wave runups along the outer coast of the Kodiak group of islands and the Kenai Peninsula, as compared to the ocean coast of Prince William Sound and the mainland east of the sound, appears to be a function of proximity to the narrow zone of high wave-energy concentration along the axis of maximum uplift.

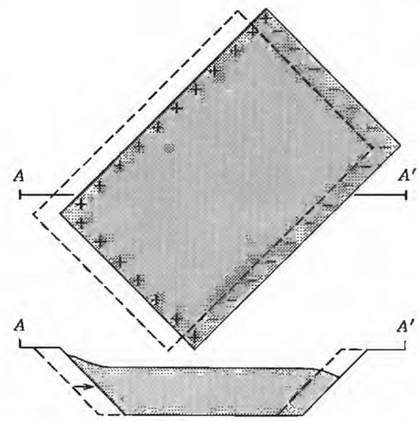
ATMOSPHERIC EFFECTS

An atmospheric pressure wave that was the atmospheric counterpart of the seismic sea waves was recorded on microbarographs at the University of California at Berkeley and at the Scripps Institute of Oceanography at La Jolla, Calif. The wave traveled at the speed of sound in air (roughly 1,050 ft per sec in the lower atmosphere), reaching Berkeley, 1,950 miles from the epicenter, 2 hours and 40 minutes after start of the earthquake (Bolt, 1964, p. 1095) and La Jolla 39 minutes later (Van Dorn, 1964, fig. 5). Travel times to these stations correspond to an initiating disturbance in the epicentral region during the earthquake. The pressure wave's signature further suggests that it was caused by the vertical tectonic displacements of the land and sea surfaces that accompanied the earthquake.

The atmospheric pressure wave also seems to have caused a traveling ionospheric disturbance that was observed in Hawaii, Alaska, and the conterminous United States on high-frequency radio sounders (Row, 1966). The disturbance at Boulder, Colo., was characterized by an abrupt onset, speeds appropriate to sound waves above 100 km in altitude, an oscillatory long-period tail, and an initial negative doppler. Computations by Row indicate that the essential features of the observations may be reproduced by sudden vertical ground displacement of the type observed in the epicentral region below a plane isothermal gravitating atmosphere.

WATER DISTURBANCES POSSIBLY RELATED TO HORIZONTAL DISPLACEMENTS

Water disturbances that accompanied the earthquake in some lakes, fiords, and rivers may have been generated by inertial effects of the water bodies as the land mass was displaced horizontally beneath them. Horizontal movement of a deep steep-sided basin or fiord, if it occurred fast enough, would be expected to impart potential energy to a contained water mass by changing its surface configuration as illustrated diagrammatically by figure 25. Thus, because of its inertia, water would tend to pile up above its original level along shores on the side of the basin opposite to the direction of displacement, and it would simultaneously be lowered along shores in the direction of displacement. For a given amount and rate of displacement, the effect of horizontal movement on the water mass would be proportionally greatest where orientation of shores is normal to the direction of horizontal movement and relatively steep basin sides permitted the maximum energy to be transferred from the basin to the contained water mass.



25.—Schematic diagram illustrating the postulated effect of a sudden horizontal displacement on water in an enclosed basin, the amount of displacement assumed to be small relative to the dimensions of the basin. Dashed lines indicate the original position of the basin, solid lines the position after displacement. Symbols along the basin margin in the plan view (above) indicate shores along which an initial rise (+) or drop (−) in water level would occur; profile *A–A'* shows a possible configuration of the water surface immediately after the displacement.

McCulloch (1966, p. A39) has reported uninodal and multinodal seiche waves in Kenai Lake with half-wave amplitudes of 5–6 feet and initial runup heights that were locally as much as 30 feet. He inferred that they were generated by a tectonic tilting of the lake basin that amounted to no more than 3 feet. A possible alternative explanation, however, is that the waves and seiche in Kenai Lake—a lake which lies in a long narrow steep-sided glacial valley—resulted mainly from the 15–25 feet of south-southeast horizontal translation of the lake basin that accompanied the earthquake in that area (fig. 16). Because of the irregular shape of the basin and uncertainties regarding the rate at which the horizontal displacements occurred, it is not possible to determine quantitatively whether the horizontal displacements alone or in combination with tectonic tilting could generate the waves recorded at Kenai Lake.

Sudden rises of water level during or immediately after the earthquake, observed at numerous coastal localities where there was no evidence for submarine sliding, strong tilting, or faulting could also have been caused by the horizontal displacements. Within Prince William Sound, where horizontal displacements in a south-to-southeast direction ranged from about 20 to 62 feet (fig. 16), local waves of unknown origin were responsible for the loss of at least 28 lives and caused extensive property damage at Chenega, Port Ashton, Port San Juan, Port Oceanic, Perry Island, and probably at Port Nellie Juan and Point Nowell (Plafker and others, 1969). Similar waves that did not cause damage also were reported at Port Wells, Unakwik Inlet, Tatitlek, Naked Island, and several other localities in Prince William Sound. Much of the damage from local waves was concentrated along east-west- to northeast-southwest-trending shores in semiconfined bays or along deep steep-sided fiords and straits. These waves, which appeared at widely separated localities in the sound within minutes after the earthquake was first felt, must have been generated locally and almost simultaneously. Most eyewitnesses observed a single large wave with runup as high as 70 feet (as at Chenega), preceded or followed by much smaller waves at intervals of a few minutes. The sudden onset, short period, and local distribution of the waves distinguish them from the train of long-period seismic sea waves generated in the Gulf of Alaska that did not reach the outer coast of the Kenai Peninsula until about 20 minutes after the start of the earthquake. That the waves may have been generated by relative seaward movement of the land mass in Prince William Sound is suggested by (1) their appearance during the earthquake, (2) their occurrence in an area where there were large horizontal displacements, and by (3) the orientation and configuration of the affected shorelines.

Unexplained waves were also observed in widely scattered coastal areas of the Kenai Peninsula and Kodiak Islands, where retriangulation data are unavailable but where significant horizontal displacements probably occurred. For example, waves as high as 9 feet were reported by eyewitnesses in the Homer area during and immediately after the earthquake. Such waves could not be attributed to sliding, slumping, or other causes (Waller, 1966a, p. D3–D4). The curious breaking and surging of the waves on the tidal flats suggested to one observer that "the land was being shoved under the bay" (Waller, 1966a, p. D4). Rapid, calm rises in water level of 9 feet at Kodiak (Plafker and Kachadoorian, 1966, p. D30) and of about 26 feet at Whittier (Plafker and Mayo, 1965, p. 15) that cannot be readily ascribed to any other cause may also have been related to horizontal displacement of the land.

In summary, horizontal displacements of the magnitude indicated by retriangulation data, if they occurred fast enough, should theoretically generate waves in water bodies of suitable size and configuration. This movement may have been the cause, or a contributing cause, of some waves observed in certain localities during or immediately after the earthquake that cannot be directly related to vertical tectonic displacements, regional tilting, seismic shaking, or submarine landslides.

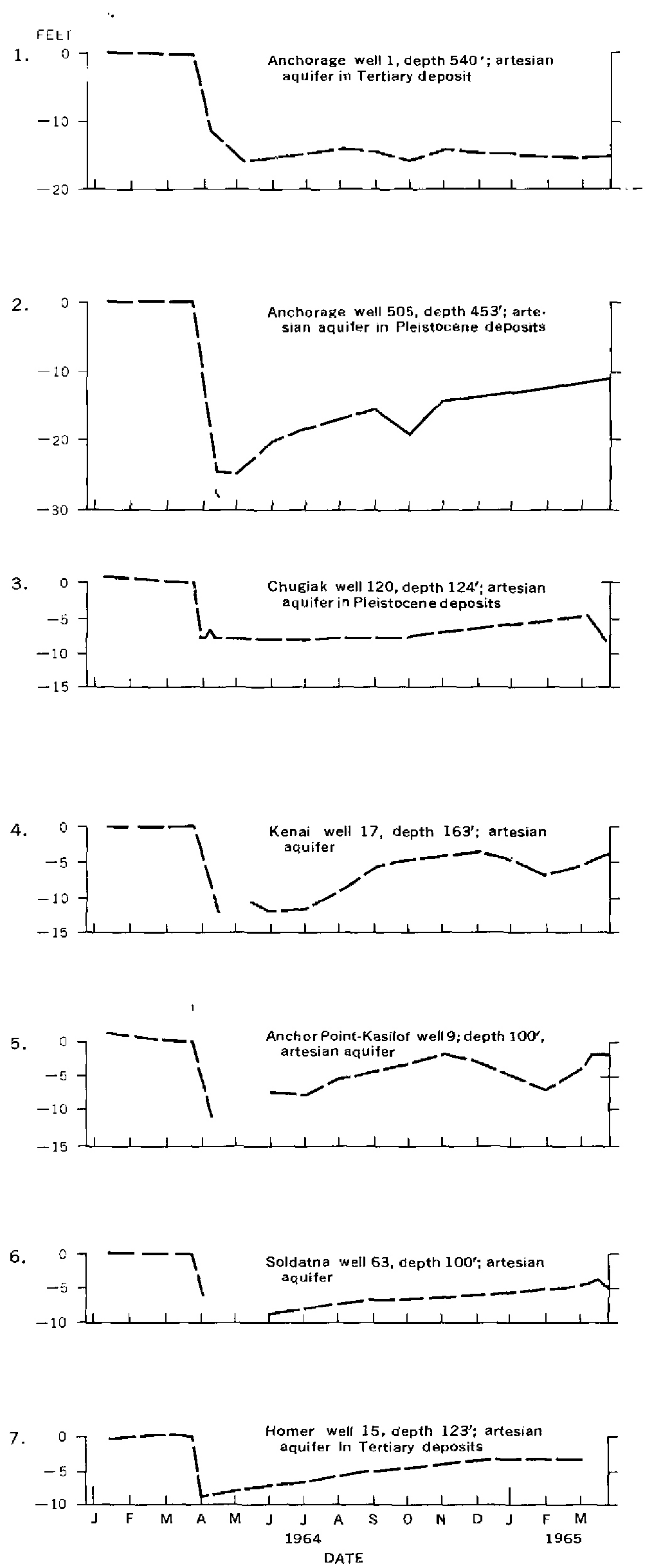
CHANGES IN ARTESIAN-WELL LEVELS

Systematic long-term drops in water levels of wells tapping confined aquifers in Pleistocene and late Tertiary strata were recorded at various widely spaced localities within the zone of tectonic subsidence (Waller, 1966a, p. D16–D18; 1966b, p. A18–A26). Records of seven representative artesian wells from Anchorage, Chugiak, and four communities on the Kenai Peninsula are shown in figure 26. The residual drops in well levels at the time of the earthquake range from about 7 to 25 feet, and none of the wells showed full recovery within a year after the earthquake.

Observed long-term changes in well levels suggest changes in the physical structure of the aquifers and a net increase in aquifer-pore space. Such changes could be caused by rearrangement of grains or fractures as a result of the horizontal extension (on the order of 2×10^{-4}) and (or) the elastic dilatation that is known or inferred to have affected the areas in which these wells are located.

A similar effect was looked for, but not found, in the oil wells of the Swanson River oil field located in the zone of tectonic subsidence near Kenai (R. I. Levorsen, written commun., 1966). Any small strain change that may have occurred probably was masked by changes in volume of the relatively compressible oil-water-gas mixture filling the pore space of the field reservoir.

The only artesian water wells in the zone of tectonic uplift are in a thick deposit of glacial drift at Cordova (Waller, 1966b, p. A20–A21). Because these wells did not have recorders installed in them, their response to the earth-



26.—Artesian well records from the zone of tectonic subsidence showing systematic drop in water levels at the time of the earthquake. After Waller (1966a, fig. 13; 1966b, figs. 14, 20, 22).

quake cannot be correlated with that of the wells in the zone that subsided. However, comparison of measurements in three wells made in July 1962 with measurements made 4 months after the earthquake showed about a 1-foot rise in water level, rather than the residual drop that characterized wells in the subsided zone.

STREAMFLOW, LAKE LEVELS, AND SHALLOW WELLS

Changes in the levels of many lakes, streams, and shallow wells in unconfined aquifers were observed at numerous localities within the zone of tectonic subsidence. In general, the reported changes involved temporary water losses (Waller, 1966b, p. A8–A11; Plafker and Kachadoorian, 1966, p. D23–D24). One of the more probable causes for such changes is an increase of intergranular or fracture porosity in the surrounding materials consequent upon horizontal extension and elastic dilatation across the subsided zone during the earthquake.

COASTAL FACILITIES AND SHIPPING

Regional land-level changes—including both subsidence and uplift—caused direct and costly damage to homes, canneries, transportation routes, airfields, docks, harbors, and other facilities throughout the affected areas (figs. 27, 28). Many facilities that had otherwise been unaffected either by the earthquake or the destructive water waves associated with it were damaged by land-level changes. Such changes had relatively few short-term beneficial effects on the works of man. Because the various forms of damage resulting from vertical tectonic movements have already been described in detail in the various reports of this series on effects to communities (U.S. Geol. Survey Professional Paper 542) and were

27.—Road along Womens Bay, Kodiak Island, in an area of about 5.5 feet of tectonic subsidence and an unknown, but probably substantial, amount of local settling of unconsolidated deposits. Since subsidence, the road has been flooded at high tide and subject to erosion by waves. Photograph taken at 4.0-foot tide stage, July 20, 1964.

28.—Canneries and fishermen's homes along Orca Inlet in Prince William Sound placed above the reach of most tides due to about 6 feet of uplift. Photograph was taken on July 27, 1964, at a 9-foot tide stage, which would have reached beneath the docks prior to the earthquake.

summarized by Hansen and Eckel (1966) and Eckel (1967), they need not be described here.

GRAVITY CHANGES

Vertical displacements were accompanied by measurable changes in gravity at several stations where comparative pre- and postearthquake gravity readings were made (D. F. Barnes, 1966; oral commun. 1966). The stations were distributed in both the zones of subsidence and the zones of uplift where changes in elevation ranged from −5.8 feet at Portage to about +11 feet at Middleton Island. Corresponding gravity changes were between +0.5 milligals to −0.67 milligals. Barnes (1966, p. 455) noted that the gravity changes, at least in the uplifted area, tend to approximate the Bouger, rather than the free-air, gradients. Although uncertainties in relocating some of the station positions preclude firm conclusions, the data suggest that there has been a redistribution of mass in at least those parts of the deformed region where the changes correspond to Bouger gradients.

COMPARISON WITH OTHER EARTHQUAKES

In terms of areal extent of deformation and amount of residual horizontal and vertical displacement, the 1964 Alaska earthquake is one of the most impressive tectonic events ever recorded. This fact is brought out by table 4, which compares the deformation associated with the 1964 event with that of selected great earthquakes for which quantitative data are available.

The area of observable crustal deformation, or probable deformation, that accompanied the 1964 earthquake is larger than any such area known to have been associated with a single earthquake in historic times. Comparable tectonic deformations have probably occurred during other great historic earthquakes, but if so they were beneath the sea, along linear coast lines, or inland, where it generally is not possible to determine the areal extent of such features with any degree of confidence. For example, the area affected by vertical displacements during the great series of Chilean earthquakes in May and June of 1960 extended

TABLE 4.—*Comparative deformations of the 1964 Alaska earthquake and some other great earthquakes*

Earthquake	Date	Approximate Magnitude (Richter scale)	Probable area of surface warping (square miles)	Maximum vertical warping relative to sea level (feet)	Maximum relative horizontal distortion (feet)	Maximum relative fault displacement (feet) (H, horizontal; V, vertical)	Data source
1964 Alaska	March 27, 1964	8.4+	108,000 (major zones only).	+38, −7½	64	2H?; 20–23V	This paper.
Niigata, Japan	June 16, 1964	7.7	≈20,000	+4.9, −1.5	No data	Submarine only, displacement unknown.	Earthquake Research Institute (1964).
1960 Chile	May 22, 1960	8.3	17,000+	+10-13, −7.2	do	No known surface faults.	Saint-Amand (1961).
Lituya Bay, Alaska	July 10, 1958	8.0	No data; fault rupture 115+ miles long.	None reported	do	21.5H; 6V	Tocher (1960).
Gobi-Altai, Mongolia	December 4, 1957. (G.)	8.6	≈32,000 (area of faulting).	No data	do	29H; 30V	Florensov and Solonenko (1963).
Nankaido, Japan	December 20. 1946.	8.5	20,000+	+2.9, −2.0	6.2	No known surface faults.	Inoue (1960, p. 84–85).
Hawkes Bay, New Zealand.	February 3, 1931	7.9	2,000 (minimum)	+9, −2 to 3	No data	6–7H; 6–8V?	Henderson (1933).
Kwanto, Japan	September 1, 1923.	8.3	10.+	+6.6, −5.3	15	6.6V	Muto (1932); Inamura (1930).
San Francisco, Calif.	April 18, 1906	8.3	No data; fault rupture 270 miles long).	Minor	≈20	21H; 3V	Richter (1958, p. 476–486).
Yakutat Bay, Alaska	September 1899	8.5–8.6	1,300 (minimum)	+47, −7(?)	No data	8+H?; 8V (on subsidiary faults).	Tarr and Martin (1912).
Assam, India	June 12, 1897	8.7	2,000 (minimum)	No data	do	0H?; 35V	Richter (1958, p. 49–50).

north-south some 420 miles along the Pacific Ocean coast, and at least 40 miles in an east-west direction at the Gulf of Ancud near the southern end of the affected area (Saint-Amand, 1961). If the inland extent of deformation to the north is comparable to that in the Gulf of Ancud, the minimum area of surface warping on land would be 17,000 square miles. The distribution of aftershocks and the source area of the destructive train of seismic sea waves associated with the earthquake suggest further that significant movements occurred over an extensive area of the sea floor adjacent to the deformed Pacific coast of Chile.

Other major earthquakes in which long sections of the west coast of South America reportedly changed elevation ocurred in 1751, 1822, 1835, 1837, and 1906 (Richter, 1958); data on the areal distribution of the changes, however, are scarce. Warping associated with one of these, the violent Chilean earthquake of 1822, gave rise to speculation that an area of some 100,000 square miles of coastal Chile had been uplifted (Lyell, 1874, p. 94), but there are few data to support this assertion. Imamura (1930) lists 26 Japanese earthquakes that resulted in vertical displacements. Regional warping has accompanied some of the more recent Japanese earthquakes, notably the 1946 Nankaido earthquake (Inoue, 1960, p. 85) and the 1964 Niigata earthquake (Hatori, 1965, p. 133–136). Elsewhere in the circum-Pacific region, additional large-scale vertical tectonic deformation was reported following the major earthquakes of 1848, 1855, and 1931 in New Zealand, and 1762 in India and Burma. Known areas of deformation associated with all of these earthquakes, however, are but small fractions of that involved in the 1964 Alaska event.

The 38 feet of uplift on Montague Island during the 1964 earthquake is known to have been exceeded only by the 47.3 feet of uplift that occurred during the earthquakes of 1899 that were centered at Yakutat Bay, 185 miles to the east (Tarr and Martin, 1912, pl. 14). Reported submarine vertical displacement offshore from Montague Island, however, may equal or perhaps exceed the 1899 movements (Malloy, 1964, p. 1048).

Subsidences roughly equal to, or slightly larger than, the 7½ feet that accompanied the 1964 earthquake have been recorded during previous seismic events, although at many places determination of the absolute amount has been complicated by surficial slumping or compaction effects in unconsolidated deposits. For example, the largest reported coastal subsidence, 17 feet, which accompanied the Great Cutch earthquake (Lyell, 1874, p. 98) was at the mouth of the Indus River in an area where significant surficial compaction of deltaic deposits was to be expected.

Following the 1923 Kwanto, Japan, earthquake, unusually large submarine displacements, involving as much as 825 feet of uplift and 1,320 feet of subsidence on the sea floor in Sagami Bay, were inferred from a comparison of pre- and postearthquake soundings (Richter, 1958, p. 570–571). However, because of (1) possible errors in the positioning of the preearthquake survey and (2) the effects of submarine sliding, these data are of doubtful value for inferring tectonic movement; consequently they are not included in table 4.

The regional horizontal displacements associated with the 1964 earthquake, which were about 64 feet, are significantly larger than any previously recorded. By contrast, regional displacements associated with the great 1923 Kwanto earthquake, as indicated by retriangulation, were less than 15 feet (Muto, 1932, fig. 6), and the maximum relative displacement resulting from combined distortion and fault offsets associated with the 1906 California earthquake was roughly 21 feet (Reid, 1911).

Measured vertical displacements across the two subsidiary earthquake faults on Montague Island—the Patton Bay fault (20–23 ft) and Hanning Bay fault ($16\frac{1}{3}$ ft)—are considerably larger than any others of definite reverse type previously described (Plafker, 1967b). A few normal or oblique-slip faults, however, have undergone larger vertical displacements. The largest of these was associated with the 1897 Indian earthquake and reportedly was as much as 35 feet (Richter, 1958, p. 51).

TECTONIC SETTING

In the following sections the 1964 earthquake is viewed from the perspective of its broad relationship to the Aleutian Arc and to other major structural elements of south-central Alaska. Emphasis is placed on the geological and geophysical evidence for late Cenozoic tectonic movements in the immediate region affected by the earthquake. Earlier tectonic features and events, and contemporaneous features and events outside the region affected by the earthquake, are not of primary interest here, although they are mentioned where necessary to provide the setting of the principal features and events that are treated.

Most of Alaska has been studied geologically in a reconnaissance manner, but detailed mapping is still relatively uncommon and is largely confined to a few mining districts and to outcrops of potentially petroliferous or coal-bearing rocks around the margins of sedimentary basins. Gross aspects of the lithology and structure in outcrop areas are reasonably well known, but present knowledge of the geologic record in most places is inadequate for a detailed interpretation of the tectonic history. Subsurface investigations in the vast terrestrial areas of sparse outcrops and in the submarine parts of the Continental Shelf and Aleutian Arc are in a state so rudimentary that geological interpretations based upon them must be considered as very tentative. The spatial distribution of earthquakes and the rapidly increasing number of available fault-plane solutions provide valuable indirect evidence for the orientation and nature of displacements that have given rise to the earthquakes. Recent improvements in locating hypocenters of Alaskan earthquakes should contribute substantially to resolving the precise relationship of the earthquakes to known structural features.

THE ALEUTIAN ARC

The 1964 earthquake occurred in the tectonically complex region where northeast-trending structural elements associated with the Aleutian Arc overlap and merge with arcuate structures of south-central Alaska. The belt of seismic activity and the major zones of tectonic deformation associated with the 1964 earthquake lie largely between and parallel to the chain of active volcanoes and the oceanic trench that constitute the arc and are presumably related genetically to it (fig. 1). Consequently, it is pertinent to review briefly the data and current hypotheses relevant to tectonic processes within the arc.

MAJOR FEATURES

The Aleutian Arc, which sweeps 1,800 miles (2,800 km) across the North Pacific Ocean from Kamchatka to southern Alaska, exhibits all of the striking features characteristic of the festoons of arcs that ring the Pacific basin. These are (1) an arcuate deep oceanic trench — the Aleutian Trench—which is convex toward the ocean basin except near its eastern end, (2) a subparallel volcanic chain on the concave, or inner, side, away from the ocean basin—the Aleutian volcanic arc, (3) an associated belt of active seismicity between the trench and volcanic chain in which the lower limit of hypocenters tends to deepen from the vicinity of the trench toward the arc, and (4) parallel zones of isostatic gravity lows over the trench bottom and gravity highs between the trench bottom and volcanic arc.

The Aleutian Arc differs from most other arcs in that it partly traverses and partly follows along, the margin of the oceanic basin (fig. 1). In its western part the arc consists of the Aleutian Ridge, which is surmounted by a chain of volcanoes that comprise the Aleutian Islands, and the Aleutian Trench, which lies at a distance of less than 155 miles (250 km) on the convex (south) side of the ridge. As it approaches south-central Alaska, the distance between the volcanic arc and the trench gradually increases to more than 250 miles (400 km), the belt of seismic activity becomes broader and less well defined, and the trench gradually shallows to become indistinguishable from the floor of the North Pacific Ocean. It is in this eastern portion of the arc that the line of volcanoes and the belt of seismicity overlap and in part merge with preexisting structural elements that roughly parallel the margin of the Gulf of Alaska (fig. 29). The group of volcanoes in the Wrangell Mountains, which are separated from Mount Spurr, the most northerly volcano of the Aleutian Range, by a gap of 420 miles, are similar in age and composition to the Aleutian Arc volcanoes. These similarities, and their position near the eastern end of the Aleutian Trench and its associated belt of intermediate-depth earthquakes, suggest that the Wrangell Mountains volcanoes may also be genetically related to the arc.

The geology of the Aleutian Islands segment of the arc was summarized by Coats (1962), that of the Aleutian Range and nearby areas by Burk (1965). Both of these writers presented thorough reviews of available bathymetric, seismologic, and marine geophysical data as well as hypotheses concerning origin of the arc. Results of detailed marine geophysical surveys along the eastern end of the Aleutian Arc, carried out during 1964 and 1965 by the Scripps Institute of Oceanography and the U.S. Coast and Geodetic Survey, had not been published at the time this report was completed.

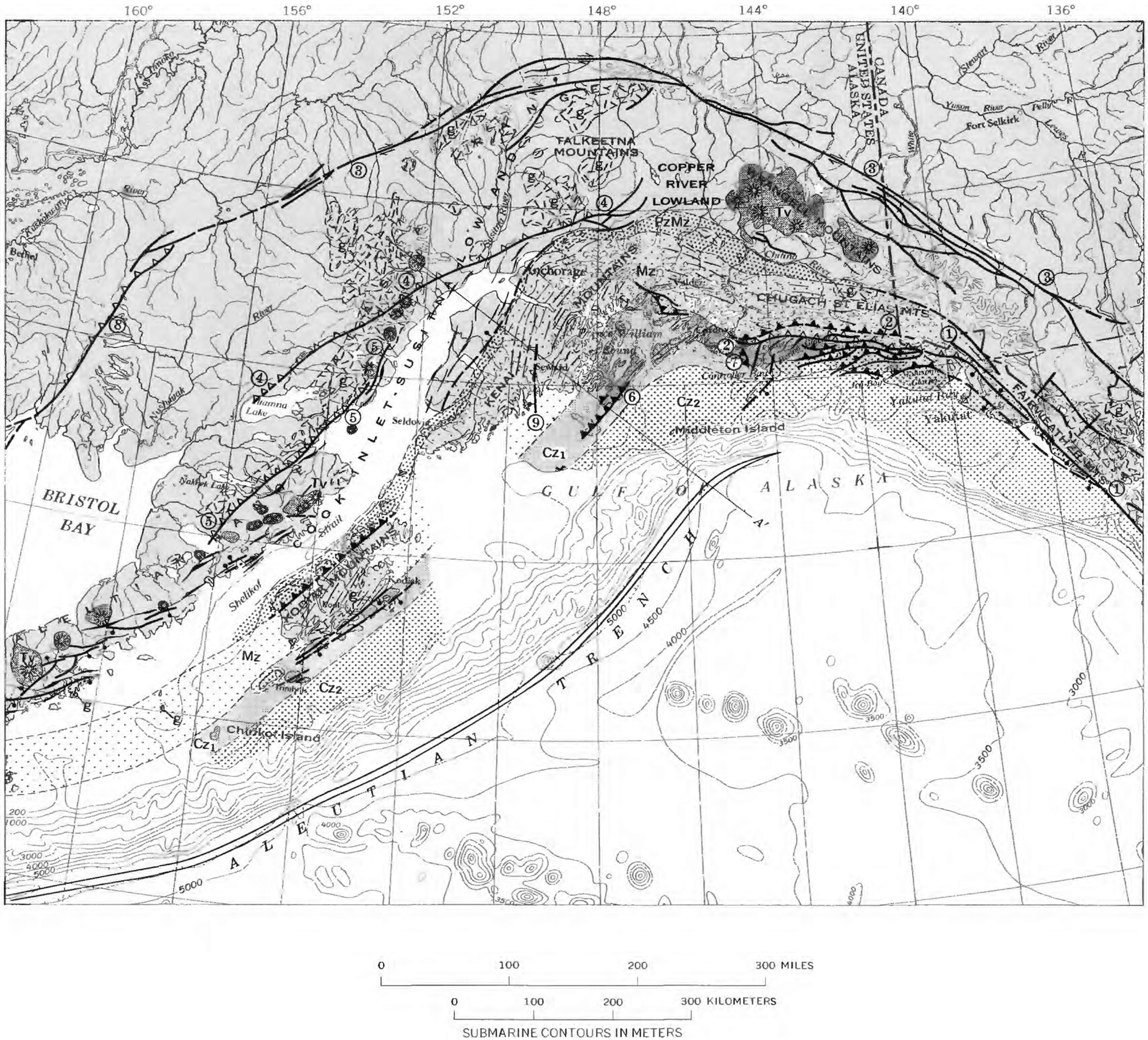
SEISMICITY

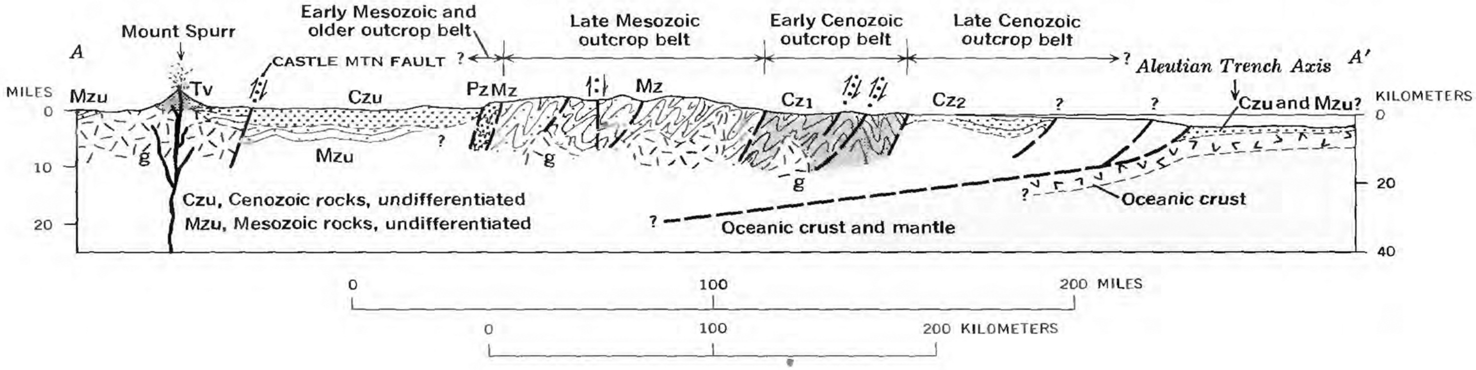
Figures 1 and 30 indicate that, in plan, most of the belt of concentrated seismic activity is between the trench and the volcanic chain along the length of the arc; a small number of earthquakes occurred on the south wall of the Aleutian Trench and north of the volcanic chain. Shallow-depth (>70 km) earthquakes have occurred throughout the area included between the Aleutian Trench and the Aleutian volcanic arc, but most of the intermediate-depth earthquakes (70–170 km) were located in the northern part of this area or north of the volcanic chain. The most easterly intermediate-depth earthquakes have been along the coast of the North Pacific Ocean at long 145° W. in the vicinity of the Wrangell Mountains. In south-central Alaska the belt of seismicity associated with the Aleutian Arc bifurcates into a broad zone of shallow and intermediate depth shocks that sweeps northward into central Alaska and a belt of shallow-focus earthquakes that extends eastward along the continental margin. Comparison of figures 2 and 31 shows that the spatial distribution of the seismicity associated with the 1964 Alaska earthquake closely follows the pattern of previously recorded earthquakes along the arc and thereby implies a genetic relationship between them, as suggested by Arthur Grantz, shortly after the event (Grantz and others, 1964, p. 2).

From studies of earthquake distribution along the Aleutian Arc, Benioff (1954, p. 391) concluded that the hypocenters lie in a broad zone that dips northward from the vicinity of the Aleutian Trench at an average angle of 28°. Saint-Amand (1957, p. 1348) interprets the same data as indicative of a zone dipping at an angle of about 45° between long 172° and 179° W. and at a "somewhat lower angle" in the eastern part of the arc. Detailed studies of seismicity associated with the 1964 earthquake suggest that some planes within the earthquake focal region dip beneath the arc at angles of less than 15°.

Benioff postulated that the dipping seismic zone marked the location of a complex "reverse" fault (termed a "megathrust" by Coats, 1962, p. 103) along which the arc relatively overrides the ocean basin. According to the sea floor spreading hypothesis advanced by Hess (1962, p. 617) and Dietz (1961), arc structures are sites of down-welling mantle-convection currents, and the planar seismic zones dipping beneath them mark the zone of shearing produced by downward-moving material thrust against a less mobile block of the crust and upper mantle. This hypothesis is in general consistent with other data suggestive of active spreading of the sea floor in a northwest-southeast direction away from the East Pacific Rise (Vine, 1966, p. 1412), and with active regional shoreline submergence suggestive of a downward-directed component of deformation near the eastern end of the arc (p. I 60).

Because of the scarcity of standard seismograph stations in Alaska as well as incomplete travel-time data for Alaska earthquakes prior to 1964, the horizontal and vertical distribution of earthquakes could not be defined precisely enough to resolve details of structure within the broad seismic zone either at depth or laterally along it. As a result, large uncertainties exist regarding the


160°
156°
152°
148°
144°
140°
136°
62°
60°
58°
56°
54°
CANADA
UNITED STATES
ALASKA
Yukon River
Fort Selkirk
TALKEETNA MOUNTAINS
COPPER RIVER LOWLAND
PzMz
Tv
Mz
Anchorage
CHUGACH-ST ELIAS MTS
Prince William Sound
Seward
KENAI MOUNTAINS
Cz2
Cz1
Middleton Island
Controller Bay
Icy Bay
Yakutat Bay
Yakutat
FAIRWEATHER
Bethel
Kuskokwim River
Nushagak
Iliamna Lake
Naknek Lake
Seldovia
COOK INLET-SUSITNA LOWLAND
BRISTOL BAY
Shelikof Strait
KODIAK MOUNTAINS
Kodiak
Trinity Is.
Chirikof Island
G U L F O F A L A S K A
A'
A L E U T I A N T R E N C H
5000
4500
4000
3500
3000
0 100 200 300 MILES
0 100 200 300 KILOMETERS
SUBMARINE CONTOURS IN METERS



EXPLANATION

Tv

Andesitic extrusive rocks of active or dormant volcanoes

Cz2

Late Cenozoic bedded rocks

Lighter pattern where projected offshore

Cz1

Early Cenozoic bedded rocks

Lighter pattern where projected offshore

Mz

Late Mesozoic bedded rocks

Lighter pattern where projected offshore

PzMz

Paleozoic and early Mesozoic bedded rocks

Lighter pattern where projected offshore

g

Granitic plutonic rocks

Undifferentiated rocks

Approximate contact

Includes possible fault contacts, Dashed where inferred or concealed

Thrust or reverse fault

Dashed where inferred. Sawteeth on upper plate. Open teeth indicate major fault

Steeply dipping fault

Dashed where inferred. Arrows indicate relative lateral displacement; bar and ball on relatively downthrown side

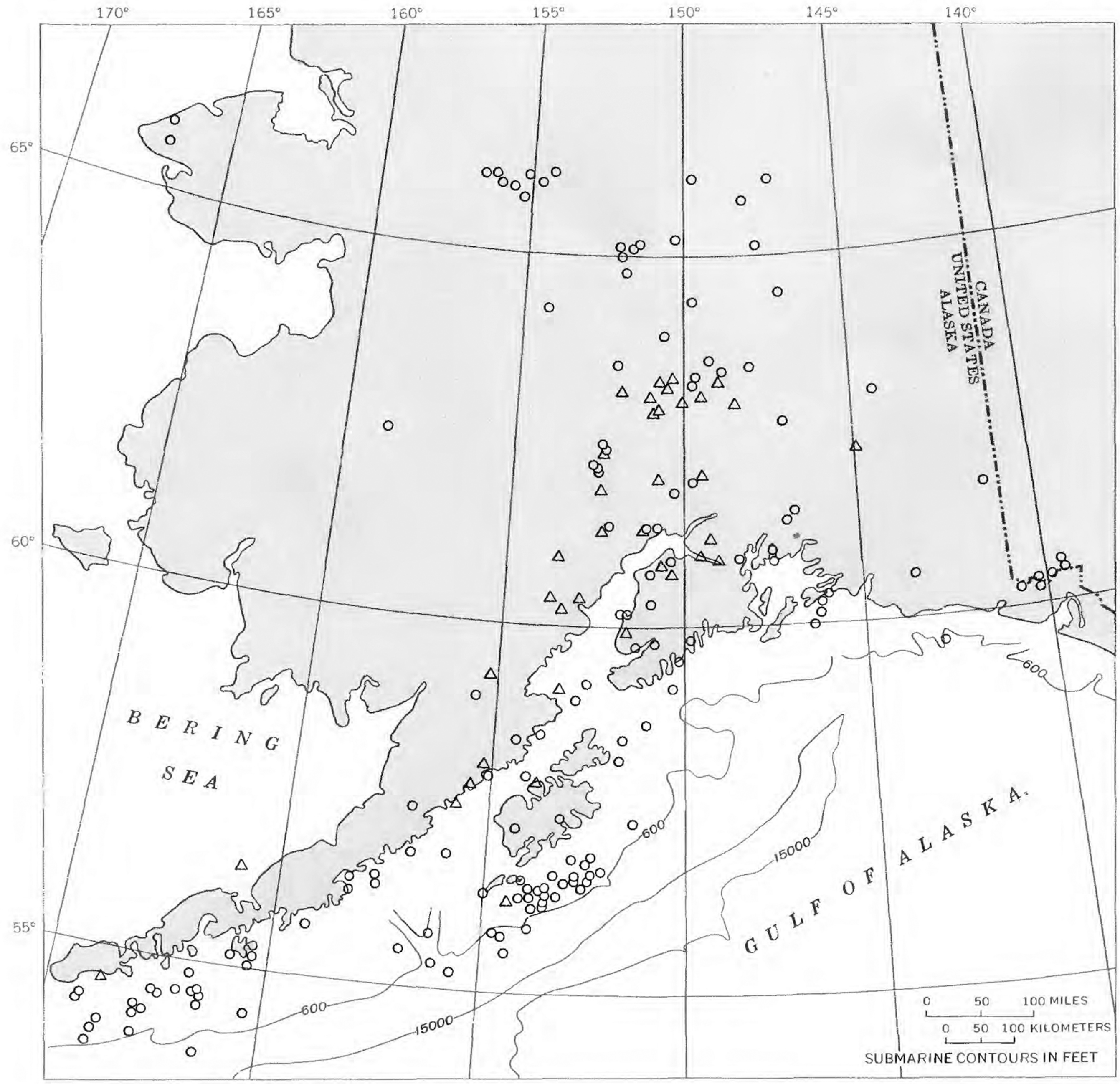
Trend lines showing strike of bedding, schistosity, and folds

Major faults and faults with known Holocene movement

Asterisk indicates known Holocene movement; double asterisk indicates historic movement

No	Fault	Data Source
1**	Fairweather	Tocher (1960); Tarr and Martin (1912); Plafker (1967)
2.	Chugach—St Elias (probable Holocene movement)	Miller and others (1959, p. 42); Plafker (1967)
3*	Denali	St. Amand (1957); Hamilton and Myers(1966); Grantz (1966)
4*	Castle Mtn-Lake Clark	Martin and Katz (1912, p. 72–75); Kelly (1963, p. 289); Grantz (1965, sheet 3)
5.	Bruin Bay	Burke (1966, p. 139); R. L. Detterman, (oral commun., 1967)
6**	Patton Bay and Hanning Bay	Plafker (1968)
7*	Ragged Mtn	Miller (1961)
8*	Holitna-Togiak	Hoare (1961, p. 608–610)
9.	Kenai lineament (possible 1964 movement)	This paper

29.—Generalized tectonic map and idealized vertical section showing selected rock units and structural features of south-central Alaska. Indicated displacement direction on faults is the net late Cenozoic movement only. Geology modified from a manuscript tectonic map of Alaska by P. B. King and from unpublished U.S. Geological Survey data; the thickness of crustal layers and the structure shown in the section are largely hypothetical.



30.—Epicenters of earthquakes ($M \geq 4$) in central Alaska during the period January 1954 to March 1963. Shallow depth ($\leq$ 70 km) earthquakes indicated by circles; intermediate depth ($\geq$ 70 km) indicated by triangles. Data after Tobin and Sykes (1966).

precise orientation and lower depth limits of planes within the zone along which earthquakes originate. Nor is it known to what extent movement on subordinate faults within the upper plate contributes to the recorded seismicity.

Focal mechanism studies of arc earthquakes suggest that the pattern of faulting is extremely complex and that it changes significantly laterally along the arc. The small number of fault-plane solutions of arc earthquake sequences that are based on both P and S waves tend to group in pairs of dip-slip and strike-slip (Bollinger and Stauder, 1965; Stauder, 1967. p. 218–219); those based solely on P waves are almost entirely strike-slip (Hodgson, 1957, p. 641). The dip-slip solutions may be interpreted as representing northwestward-dipping planes that parallel the trend of the eastern part of the arc and are oblique to it in the central part. Coats (1962, p. 95) observed that solutions yielding predominantly strike-slip movement on steeply dipping planes may originate on transverse tear or wrench faults oriented at a large oblique angle to the arc trend. Coats further suggested that the positions of individual volcanoes may be localized by transverse fractures and that their regular position relative to the trench is determined by the distance at which tensional fractures can penetrate through the upper plate to tap eruptible material.

The predominance of transverse faults and other linear elements on land and offshore from the Aleutian Islands (Coats, 1962, fig. 2; Gates and Gibson, 1956, fig. 12) and relatively abrupt transverse boundaries for aftershock sequences of the February 1964 Rat Island earthquake (Jordan and others, 1965), the 1957 earthquake sequence in the Andreanof-Fox Island region (Brazee, 1965), and the 1964 earthquake (p. I 15), all lend support to the inference that transverse faults segment the plate above the megathrust into large tectonic blocks. Conversely, there is no seismological or geological evidence of systematic dextral strike-slip displacement parallel to the arc in support of the speculation that it marks the trace of a major fault along which the North Pacific Ocean basin is rotating counterclockwise relative to North America, as was suggested by Saint-Amand (1957, p. 1367).

INFERRED REGIONAL STRESS PATTERN

According to the sea-floor spreading hypothesis, compressive forces should be directed roughly normal to the arcs and the deep-sea trenches would be interpreted as due to dragging down of the oceanic crust above a descending mantle convection current (Hess, 1955, 1957, 1962). In the Aleutian Arc, transverse compression landward from the trench axis is suggested by (1) the orientation of major stress axes derived from mechanism studies of most, but not all, previous arc earthquakes (Balakina and others, 1961; Lensen, 1961, fig. 4); (2) the orientation of the axes of folded late Cenozoic rocks in the Aleutians, on the Alaska Peninsula and around the Gulf of Alaska roughly parallel to the arc trend (Coats, 1962, p. 93; Burk, 1965, p. 3; fig. 29); and (3) the distribution and nature of the seismicity, surface faulting, and warping associated with the 1964 earthquake. Both the pattern of residual displacements in 1964 (fig. 3), and the late Cenozoic fold-fault pattern (fig. 29) near the eastern end of the arc suggest that regional compressional deformation extends across it, from at least the outer edge of the Continental Shelf to the vicinity of the volcanic arc.

On the other hand, reconnaissance marine-geophysical studies across the Aleutian Trench and along the north wall of the trench have revealed an apparent lack of features indicating transverse compression or the existence of a megathrust intersecting the sea floor. From interpretations of gravity and magnetic data over the Aleutian Trench and outer margin of the arc southwest of Kodiak Island, Peter, Elvers, and Yellin (1965, p. 366) concluded that there probably exists "crustal tension, rather than compression, underthrust, or downbuckle." Marine reflection-refraction data suggest that the trench, with its relatively undeformed veneer of acoustically transparent sediments, reflects an origin by tension or vertical movements of the crust (Shor, 1966, p. 221; Ewing and others, 1965, p. 4599; Von Huene and others, 1966, p. 176).

It should be emphasized, however, that the geophysical investigations, particularly along the north wall of the trench, have provided direct information only on the approximate geometry and properties of the crust; they do not permit unambiguous conclusions regarding the deformations or the nature of the causative forces. More data, including a knowledge of the lithology and internal structure of the crustal rock in these areas, are required to determine whether features that seem to indicate absence of compression actually reflect the regional stress pattern. The seismic and geologic evidence in the eastern part of the Aleutian Arc suggests to me that any regional stress, other than transverse compression, must be largely limited to the part of the arc that lies seaward from the edge of the Continental Shelf.

SUMMARY OF THE PRE-HOLOCENE (RECENT) TECTONIC HISTORY OF SOUTH-CENTRAL ALASKA

The tectonic history along the eastern end of the Aleutian Arc and adjacent parts of south-central Alaska may be interpreted from the geologic record of folding and faulting. This record suggests that the present orogenic cycle, which probably began in Pliocene time, has resulted in regional compressive deformation in a general northwest-southeast to north-south direction around the margin of the Gulf of Alaska. It further indicates that at least three previous major orogenies culminated in south-central Alaska during (1) late Eocene to early Oligocene time, (2) Late Cretaceous to earliest Tertiary time, and (3) Middle(?) Jurassic to Late Jurassic or Early Cretaceous time. These earlier orogenies are indicated by major unconformities that divide the exposed section into four laterally continuous time-stratigraphic units that constitute gross subdivisions of the stratigraphic record. For brevity they are referred to in the following discussion as the late Cenozoic, early Cenozoic, late Mesozoic, and early Mesozoic and older sequences. They are described in the order named because it is the youngest deformation that is of primary concern here and because each successive orogeny tends to mask earlier events and thereby obscure to a large extent the earlier tectonic history.

Each of the four orogenies involved significant structural shortening of thick eugeosynclinal or geosynclinal sequences along the continental margin through folding and imbricate faulting. Deformation of coeval rocks inland from the continental margin was significantly less severe. This relationship may reflect a progressive reduction in horizontal compressive stress away from the continental margin, an increase in crustal competence, or a combination of both factors. In the mountains bordering the Gulf of Alaska and on the continental margin, the four major sequences form laterally continuous outcrop bands, commonly bounded by faults, which progressively increase in age from south to north (fig. 29). The distribution of coeval sequences to the north of the coastal mountains is not shown in figure 29 because the sequences exhibit complex overlapping relationships and are obscured to a large extent by plutonic intrusions, young volcanic rocks, water, or a veneer of unconsolidated deposits. The larger plutons and active or dormant volcanoes in south-central Alaska are delineated on figure 29.

A concise summary of the physiography and general geology of the area affected by tectonic movements during the earthquake, based on the compilation by Wahrhaftig (1966), has been given by Hansen and Eckel (1967) in the introductory volume of this report series and need not be repeated here. The broad tectonic setting and structural history of Alaska has been presented by Payne (1955) and by Gates and Gryc (1963). The stratigraphy and general geology of sedimentary basins in south-central Alaska were discussed by Miller, Payne, and Gryc (1959). Burk (1966) synthesized the geology of the western Alaska Peninsula and analyzed the relationship of the peninsula to the Aleutian Arc and continental margin. Grantz (1966) summarized data on strike-slip faults in Alaska and evaluated their role in the tectonic evolution of Alaska. All these writers included exhaustive bibliographies in their publications. Recent geologic maps along the Gulf of Alaska margin include a 1:500,000-scale compilation of the area between Prince William Sound and southeastern Alaska (Plafker, 1967a) and a 1:250,000-scale compilation of most of the Kodiak group of islands (Moore, 1967). The following section is based largely on the publications cited above, but includes some modifications of the Cenozoic history based on previously unpublished U.S. Geological Survey data.

POST-MIOCENE DEFORMATION

The orogeny, which began in Pliocene time and continued to the present, resulted in differential uplift and faulting throughout southern Alaska (Miller and others, 1959, p. 17). Folding was severe along the margin of the Gulf of Alaska, but gentle elsewhere except in the immediate vicinity of major faults. During this orogeny, the coastal mountains were uplifted and in places, especially to the west of Yakutat Bay, were thrust relatively southward against the Pacific basin along a system of high-angle faults (fig. 29). Major faults and folds in the late Cenozoic sequence tend to parallel the trends of the older structures in the Pacific Border Ranges; there is a general increase in the intensity of folding and magnitude of fault displacements from south to north across the deformed belt.

Transverse trends occur in the structurally complex Controller Bay area east of Prince William Sound where folds involving Oligocene and Miocene strata are typically of small amplitude, tightly compressed, and asymmetric or overturned; axial planes are inclined to the west or north.

Horizontal shortening of upper Oligocene and lower Miocene strata along one 5-mile north-south structure section in the Controller Bay area averages close to 25 percent (Don J. Miller, unpub. data).

To the east of the Controller Bay area the upper Cenozoic rocks form broad synclines and tightly appressed asymmetrical anticlines cut by north-dipping overthrust faults that strike roughly parallel to the coast. At least some of these faults probably represent the leading edges of thrust sheets developed by gravitational gliding off the uplifted Chugach-Saint Elias Mountains. On the south limbs of many of these anticlines, strata as young as late Pliocene are steeply dipping or overturned towards the north. At least two major intraformational unconformities in the upper part of the sequence record pronounced folding and uplift during late Pliocene time. Minor unconformities and thick beds of coarse conglomeratic sediments in the sequence reflect intervals of local uplift and erosion beginning in early Miocene time.

Marine sedimentary rocks of early Pleistocene or younger age on Middleton Island near the edge of the Continental Shelf have been titled northwestward at an average angle of 28°, truncated, raised above sea level, and displaced by active minor faults. A pervasive conjugate system of shear joints cutting the sequence reflects compressive stress directed northwest-southeast, normal to the strike of the beds.

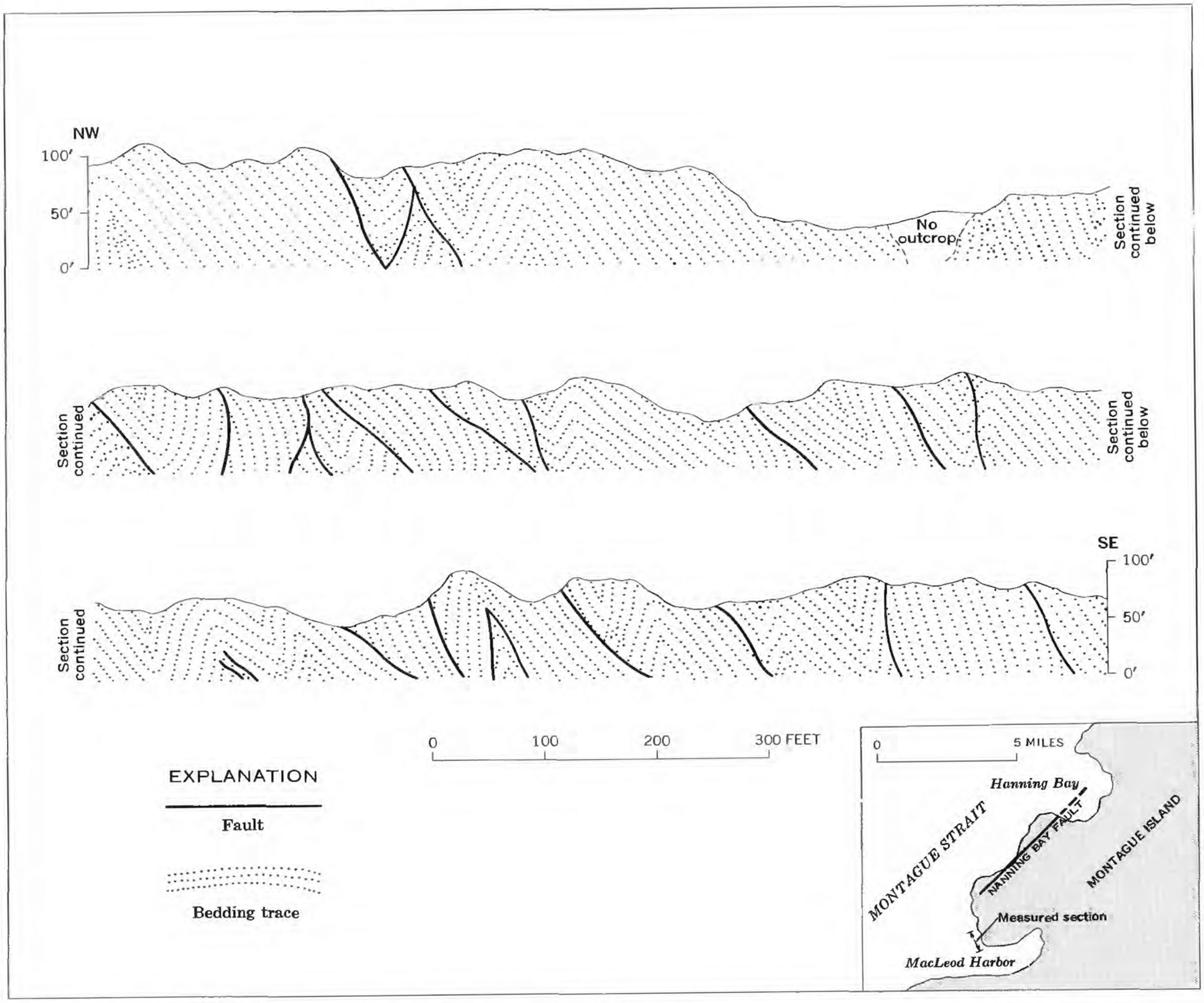
Bedded rocks of Oligocene through middle Miocene age exposed in narrow belts along the southeast coast of the Kodiak Islands, and Oligocene rocks on Chirikof Island, are tightly folded about northeast-trending axes and are locally overturned (Moore, 1967). Relatively undeformed Tertiary rocks are exposed along the margin of the Gulf of Alaska only in the Trinity Islands and on Chirikof Island. On Chirikof, late Pliocene strata are exposed in homoclines with dips of less than 10°.

Late Cenozoic deposits along the Aleutian Volcanic Arc, in the Cook Inlet-Susitna lowland and around the Copper River lowland, were uplifted and slightly deformed during Pliocene and Quaternary time into generally open folds, most of which trend nearly parallel with the main arc or at a slightly oblique angle (Payne, 1955; Coats, 1962, p. 93; Burk, 1965, p. 121; Wolfe and others, 1966, p. A13). Most of the Eocene or younger Tertiary sedimentary rocks around the margin of the Cook Inlet lowland are nearly flat lying, gently tilted, or folded into broad, open structures with flank dips generally less than 10°. Thinning of individual stratigraphic units over anticlinal crests in the subsurface and the occurrence of topographic anomalies in areas mantled with Quaternary deposits suggest continuous active growth of some of these structures during late Cenozoic time (Kelly, 1963, p. 289, 296).

LATE EOCENE TO EARLY OLIGOCENE DEFORMATION

A major episode of deformation and plutonic activity began in late Eocene time and probably culminated in early Oligocene time. This episode resulted in complex folding, faulting, and mild metamorphism of early Tertiary rocks in the coastal belt (Plafker and MacNeil, 1966, p. 68) and probable minor warping and faulting of the age-equivalent strata exposed in the northern part of the Cook Inlet-Susitna lowland (Miller and others, 1959, p. 18). Folds in the southern belt commonly are of short wavelength, tightly appressed with flank dips in excess of 50° and locally overturned both toward the north and south. Individual folds are of small amplitude and lateral extent and are complicated by intricate drag folding and minor overthrust faults. Horizontal shortening across a typical folded section, such as that shown in figure 31, commonly exceeds 45 percent, exclusive of imbrication on faults. Net shortening across the outcrop belt of lower Tertiary rocks is indeterminate because the structure has not been mapped in detail and traceable key horizons are generally absent. Conceivably, it could average as much as 45 percent or more across the entire outcrop belt; a net surface shortening of at least 30 miles in the Prince William Sound area. In a gross sense the strikes of bedding planes and fold axes tend to parallel the structural trends of the older rocks. They are notably divergent and complex in northeastern Prince William Sound and in the area immediately east of the sound (fig. 29), possibly because these areas are situated close to the axis of Carey's postulated Alaska orocline (1958, p. 209–212). Postorogenic potash-rich granite plutons of probable early Oligocene age (Lanphere, 1966, p. D197) with pronounced thermal aureoles are intrusive into the early Tertiary and older rocks in and near the Prince William Sound region.

Deformation of the early Tertiary rocks in the Cook Inlet-Susitna lowland during the early Cenozoic and later orogenies is characterized by broad open folding, flank dips commonly being less than 50° except in the immediate vicinity of faults. The strike of fold axes and of the steeply dipping faults of moderate dis-

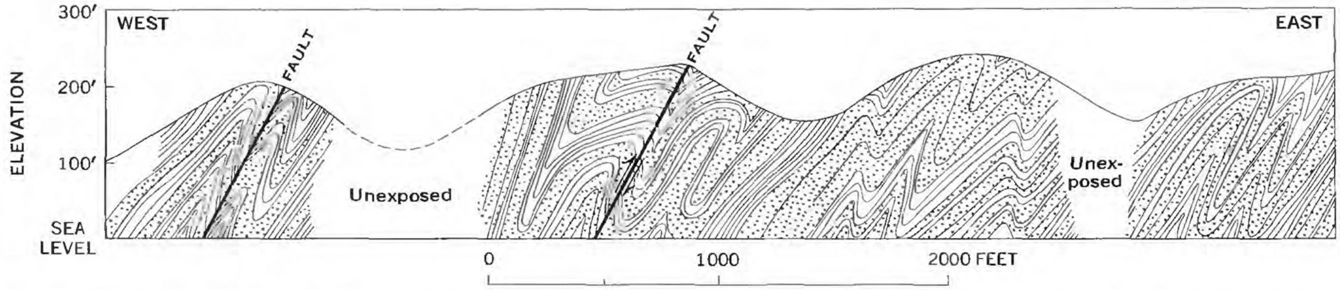


31.—Structure of the early Tertiary Orca Group exposed in a sea cliff along the north shore of MacLeod Harbor, Montague Island, Prince William Sound.

placement that offset the sequence is roughly parallel or at a slightly oblique angle to the trend of the adjacent mountain ranges. Horizontal shortening of Paleocene or Eocene strata through folding and faulting along a 5-mile-long northwest-southeast section across the Matanuska Valley (Barnes, 1962, section *A–A′*) is about 25 percent. This figure undoubtedly approaches the approximate maximum amount of shortening in the sequence, because the section is located in a narrow structurally complex trough bounded on its north side by the active Castle Mountain fault.

LATE CRETACEOUS TO EARLY TERTIARY DEFORMATION

A major episode of diastrophism, corresponding to an early phase of the Laramide orogeny in the time interval from Late Cretaceous to early Tertiary, resulted in regional deformation and widespread intrusive activity in the Kodiak-Kenai-Chugach Mountains (Chugach Mountains geosyncline of Payne, 1955) and in the Alaska Range (Alaska Range geosyncline of Payne, 1955), with relatively slight deformation of rocks in the intervening area (Matanuska geosyncline of Payne, 1955). Figure 32 shows the style of deformation in one particularly well exposed outcrop area of probable late Mesozoic rocks along the south coast of the Kenai Peninsula immediately west of Prince William Sound. Structural shortening by folding across the section is about 55 percent, and there is an unknown amount of additional



32.—Structure of probable late Mesozoic rocks along the southeast coast of the Kenai Peninsula immediately west of Prince William Sound (Martin and others, 1915, p. 216).

shortening by imbrication on overthrust faults. A comparable amount of deformation across the entire late Mesozoic outcrop belt would have resulted in a net shortening of the original sequence by 40–60 miles. Deformation of the sequence was accompanied by emplacement of stocks and small batholiths, predominantly of quartz diorite and related plutonic rocks, and by metamorphism and penetrative deformation of rocks in the cores of the eastern Chugach Mountains, the Saint Elias Mountains, and the Fairweather Range.

During the Late Cretaceous to early Tertiary orogeny, deformation inland from the coastal mountains was mainly by uplift, broad open folding, and displacement on steeply dipping faults as indicated diagrammatically on section *A–A′*, figure 29. Published structure sections across the sequence around the margins of the Copper River lowland (Grantz, 1965, sheet 3; Miller and MacColl, 1964) and on the Alaska Peninsula (Burk, 1965, figs. 21, 22) indicate structural shortening of Late Cretaceous strata that ranges from 5 to 20 percent and probably averages no more than 10 percent.

MIDDLE (?) JURASSIC TO EARLY CRETACEOUS DEFORMATION

All the older rocks in south-central Alaska were strongly affected by the major orogenic episode, roughly corresponding to the Nevadan orogeny, that began between latest Early Jurassic and earliest Middle Jurassic time and may have continued intermittently into earliest Cretaceous time. Rocks in the Seldovia geanticline were folded, complexly faulted, and regionally metamorphosed; deformation in the Talkeetna geanticline was considerably more variable and depended largely upon proximity to major batholiths. The orogeny was accompanied or immediately followed by emplacement of (1) numerous plutonic masses that range in composition from peridotite to granodiorite in the coastal mountains belt and (2) plutons of felsic composition and of batholithic size on the Alaska Peninsula and in the Talkeetna Mountains, with resultant contact metamorphism of the enclosing bedded rocks. The prevailing schistosity and slaty cleavage in the rocks of the coastal mountains are nearly vertical and have a variable strike that generally parallel roughly, or is at a small oblique angle to, the trend of the mountains.

A slight unconformity between rocks of Permian and Middle Triassic age in the Talkeetna Mountains (MacKevett and others, 1964) and the apparent absence of bedded rocks of Pennsylvanian, part of Permian, and Early Triassic age throughout the region suggest that these times were intervals of uplift and erosion.

MAJOR SURFACE FAULTS IN SOUTH-CENTRAL ALASKA

The longer coastal faults, or systems of faults, in south-central Alaska tend to follow the arcuate grain of the mountain ranges and the margin of the Gulf of Alaska; larger faults inland from the coast trend obliquely across some of the mountain ranges (fig. 29). In detail the faults do not form simple arcs but instead consist of linear segments of variable length, the included angles between adjacent segments being as little as 150°. Two minor north-south trending faults with Holocene movement intersect the regional grain at large oblique angles in the Controller Bay area (Ragged Mountain fault) and in the central Kenai Mountains (Kenai lineament).

Of the major faults delineated on figure 29, all but the Bruin Bay fault and the reverse fault along the northwest side of the Kodiak Island group exhibit evidence of post-Miocene movement; several have been active during post-Pleistocene (Holocene) time. Surface deformation was recorded in Alaska in conjunction with only three previous historic earthquakes along the Gulf of Alaska, at Chirikof Island in 1880 (George W. Moore, 1962, unpubl. data), at Yakutat Bay in 1899 (Tarr and Martin, 1912), and along the Fairweather fault in 1958 (Tocher,

TABLE 5.—*Radiocarbon-dated samples used in determining the relative ages of shoreline features*

[Locations of sample sites shown on fig. 33]

Locality No.	Lab. No.	Location	Age (B.P.)	Submergence (−) or emergence (+) relative to preearthquake datum (feet)	Data source	Comments
1	P-1039	Kodiak Island	3,279	[1]−3.3	Clark (1966, p. 370)	Archaeologic site in submerged beach deposits.
2	P-1034	Sitkalidak Isalnd	5,519±78	Slight emergence	do	Archaeologic site in elevated beach ridge.
3	P-1036	do	3,945±65	do	do	Do.
4		Kukak Bay, Shelikof Strait	1,460±95	No apparent change	D. E. Dumond (written commun., March 28, 1966).	Archaeologic site in present beach ridge.
5		do	1,450±130	do	do	Do.
6		do	1,075±100	do	do	Do.
7		do	775±110	do	do	Do.
8	P-138	Yukon Island, Kachemak Bay	1,369±102	−4 to −10	Rainey and Ralph (1959)	Antler bone from archaeological site in submerged beach deposits.
9	P-139	do	2,706±118	[1]−16	do	Do.
10	W-299	Girdwood, Turnagain Arm	2,800±180	[1]−15	Karlstrom (1964, p. 48)	Peat in tidal silts and bog deposits. Possible significant surficial subsidence.
11	W-175	do	700±250	−2½	do	Do.
12	W-1720	Seward	≈200	[1]−7	Plafker and Rubin (1967).	Rooted stump on delta. Indicated submergence may be due to significant surficial slumping of unconsolidated deltaic deposits.
13	W-1589	Perry Island, Prince William Sound.	3,680±300	−11.2	do	Rooted stump on submerged beach.
14	W-1588	Point Nowell, Prince William Sound.	930±200	−17	do	Wood fragment from interbedded peat and gravel on submerged beach.
15	W-1592	Columbia Bay, Prince William Sound.	1,140±250	−6	do	Rooted stump on submerged beach.
16	W-1591	Latouche Island, Prince William Sound.	230±200	−8.5	do	Do.
17	W-1590	MacLeod Harbor, Prince William Sound.	560±200	−9.7	do	Do.
18	W-1764	do	380±200	[1]−8	M. J. Kirkby (written commun., 1967).	Rooted stump in submerged river delta.
19	W-1766	Patton Bay, Montague Island	600±200	[1]−12	do	Wood fragment from submerged peat and gravel deposits.
20	W-1770	do	2,070±200	[1]+20	do	Peat on elevated beach deposits.
21	LJ-943	Copper River Delta area	1,360±50	−16.8	Reimnitz (1966, p. 85)	
22	LJ-945	do	380±20	−13.9	do	Rooted stumps on deltaic and marsh deposits. Submergence may be exaggerated owing to surficial subsidenc.
23	LJ-0032	do	700±50	−16.8	do	
24	LJ-939	do	725±35	−15.8	do	
25	LJ-938	do	700±30	−16.8	do	
26	I-6	Katalla area	7,650±330	[1]+180	Heusser (1960, p. 94)	Peat from elevated terrace.
27	I-7	do	3,770±200	[1]+40	do	Do.
28	W-376	Cape Suckling	390±160	[1]−14	Rubin and Alexander (1958).	Rooted stump on submerged rock platform.
29	W-1729	do	710±200	−14	Plafker and Rubin (1967).	Do.
30	W-462	do	5,120±220	[1]+20	Rubin and Alexander (1958).	Shells on elevated terrace.
31	W-1724	Middleton Island	1,350±200	+22	do	Driftwood on elevated terrace.
32	W-1404	do	2,390±200	+58	Levin and others (1965, p. 392).	Do.
33	W-1405	do	4,470±250	+137	Levin and others (1965, p. 392–393).	Do.
34	W-1722	Stockdale Harbor, Prince William Sound.	>200	−8	Plafker and Rubin (1967).	Rooted stump on submerged beach.
35	LJ-0033	Copper River Delta area	860±50	−15.9	Reimnitz (1966, p. 85)	Rooted stump on deltaic and marsh deposits.
36	LJ-0034	do	1,700±100	−22.0	do	Do.

[1] Approximate.

1960). Although surface breakage may have occurred along faults during other earthquakes in Alaska, such features could easily have gone undetected had they occurred in the vast uninhabited parts of the State.

The sense of late Cenozoic displacement on the faults shown in figure 29 is given where known. In general, faults in south-central Alaska that trend northeast or east are predominantly overthrusts or oblique overthrusts that dip northward at moderate to steep angles, with north sides relatively upthrown. Horizontal movements on major longitudinal faults are largely restricted to the northwest-trending Fairweather fault and to the northwest and east-west trending parts of the Denali fault system, both of which are predominantly right-lateral. Geologic relationships across many of the faults suggest that they have undergone recurrent movement during Cenozoic and much of Mesozoic time and that the sense of displacement along some of them has changed with time. The orientation and the Quaternary displacement on these faults reflects tangential compression oriented north-south to northwest-southeast along the margin of the Gulf of Alaska, perhaps rotating to a more nearly east-west direction in central Alaska. The broad fault pattern implies that the Pacific Ocean basin is moving north to northwest relative to the mainland. Thus, the ocean basin appears to be shearing relatively past the mainland of British Columbia and southeastern Alaska along the system of northwest-trending right-lateral strike-slip faults, whereas it shears relatively beneath the continental margin and Aleutian Arc in central and western Alaska along a system of imbricate thrust (underthrust?) faults.

THE PREEARTHQUAKE HOLOCENE (RECENT) RECORD OF VERTICAL SHORELINE MOVEMENTS

Numerous records of preearthquake differential shoreline movements relative to sea level provide data on the history of vertical tectonic movements during Holocene time. Reconnaissance studies of the displaced shorelines have brought out (1) a general similarity between the pattern of earthquake displacements and the long-term trend of Holocene coastal emergence or submergence, and (2) a remarkable recent widespread submergence over much of the zone that was uplifted during the 1964 earthquake and over at least part of the zone that subsided. In addition, radiocarbon dating of material from coastal sites has provided quantitative data on the duration and rates of these Holocene movements.

Data and interpretations presented in this section are largely taken from a preliminary paper by Plafker and Rubin (1967). All radiocarbon-dated samples referred to, sources, and other pertinent data are listed in table 5; sample locations are shown on figure 33.

As used herein, the terms "uplift" and "subsidence" describe a tectonic rise or fall of the land; "emergence" and "submergence" indicate relative movements that may be the sum of both tectonic movements and eustatic sea-level changes. Tectonic movements include those that result from both diastrophic and isostatic processes. "Long-term" refers to shoreline changes relative to sea level having durations measured in thousands of years; "short-term" refers to the general submergence that occurred during the 1,350 years or less prior to the earthquake.

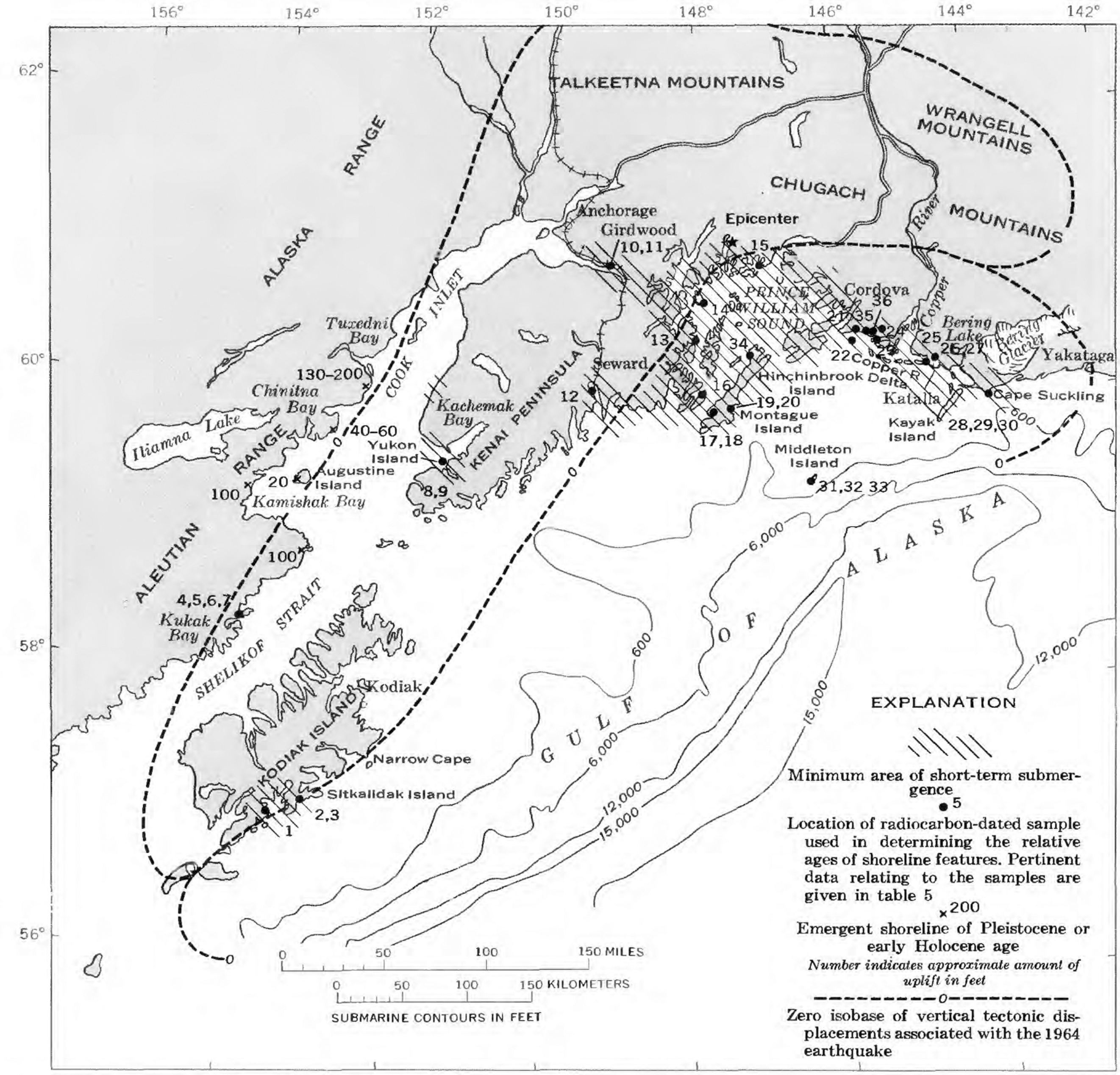
Some of the observed recent submergence in areas shown on figure 33 is undoubtedly exaggerated by the local vibration-induced compaction or slumping of unconsolidated deposits during the earthquakes that frequently rock this seismically active region. As was demonstrated during the 1964 earthquake, surficial submergence may be substantial in areas of thick unconsolidated saturated deposits—especially in those deposits that are not constrained on one or more sides, such as deltas, spits, and barrier beaches. Consequently, rooted tree stumps, on bedrock or on thin beach deposits overlying bedrock, were used to determine amounts and rates of pre-1964 vertical displacement wherever possible.

In the absence of a reliable local eustatic sea-level record for southern Alaska, it is assumed that Holocene sea levels were probably comparable to those in more tectonically stable parts of the world where the sea-level record has been worked out in some detail. Three recent interpretations of eustatic sea levels are shown on figure 34. Most of these studies suggest either (1) a rather rapid rise in sea level at an average rate of about 0.08 inch per year until it reached approximately its present level between 2,000 and 6,000 years ago (Coleman and Smith, 1964), or (2) that sea level has been rising slowly and continuously from about −33 feet to its present level during the past 7,000 years with a generally rising, but fluctuating, sea level between about 15,000 and 7,000 years B.P. (Curray, 1961). Some authorities believe that sea level reached its present position from 3,000 to 5,000 years ago and has been fluctuating above and below its present position by about 3–6 feet ever since (Fairbridge, 1961). For present purposes, it is significant to note that none of these studies suggest that eustatic sea-level fluctuations during the last 10,000 years were large enough alone to explain the relative shoreline displacements found along the coast of south-central Alaska.

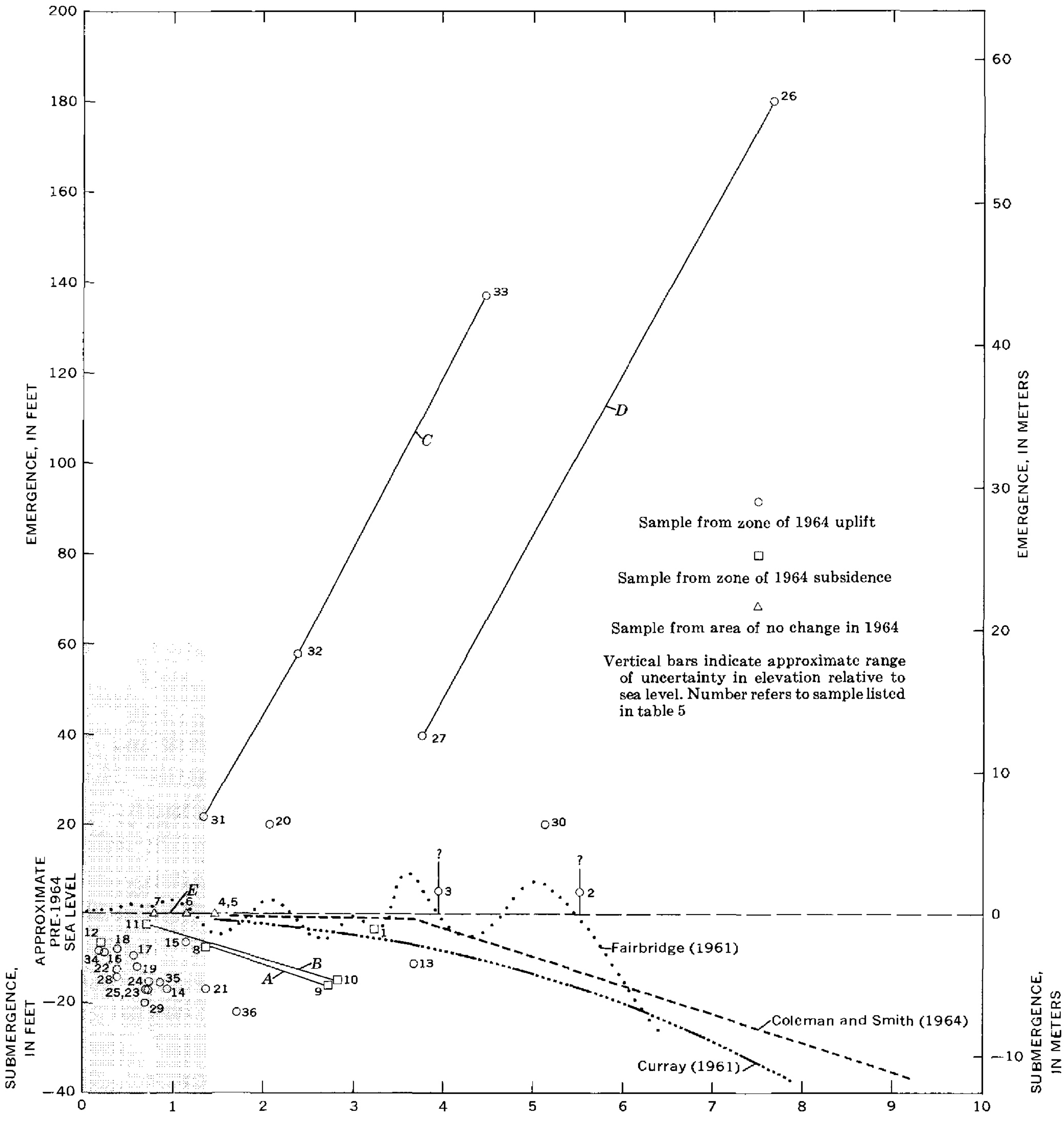
LONG-TERM HOLOCENE EMERGENCE AND SUBMERGENCE

The record of Holocene displacements in the area affected by the earthquake, as deduced from shoreline morphology and from radiocarbon dates, is outlined in the following section. Places referred to and the spatial distribution of radiocarbon-dated shoreline samples in relation to the 1964 deformation are shown in figure 33. The age and pre-1964 position of these samples relative to sea level is shown in figure 34.

Much of the shore in the mountainous southern part of the area that subsided during the earthquake exhibits the characteristic features of a deeply drowned coast. Submerged shorelines occur along most of Kodiak and the adjacent islands, the southern Kenai Peninsula, the south shore of Kachemak Bay, and part of Turnagain Arm. The northern and eastern parts of the Prince William Sound region also appear to be submergent, although the evidence for relative shoreline movements is somewhat obscured by recent glaciation in most of the fiords. In these submerged mountainous areas, former rocky ridges are now peninsulas or islands, and



33.—Areal distribution of vertical displacements associated with the 1964 earthquake, areas of preearthquake submergence, and locations of radiocarbon-dated samples used to determine relative ages of shoreline features.



34.—Pre-1964 positions of radiocarbon-dated samples relative to the pre-1964 sea level and proposed eustatic sea levels of Curray (1961), Fairbridge (1961), and Coleman and Smith (1964). The duration of the inferred period of short-term subsidence is indicated by the stippled area. Sample locations at Yukon Island (*A*), Girdwood (*B*), Middleton Island (*C*), Katalla (*D*), and Kukak Bay (*E*) are connected by solid lines.

35.—Deeply submerged coast along the southern Kenai Peninsula. Embayments, which give the shore a distinctive scalloped appearance, are cirques whose floors are drowned to depths of as much as 300 feet below sea level. Shoreline shown was submerged an additional 2–5 feet in 1964. Photograph by T. L. Péwé.

drowned river valleys or glacial cirques have become embayments (fig. 35). The general scarcity of well-developed sea cliffs, beaches, and similar shore features attests to the recency of the submergence. Elsewhere in the subsided zone, coastal bogs of terrestrial peat and some aboriginal dwelling sites that are now inundated by high tides also indicate long-continued submergence relative to sea level.

The most pronounced submergence appears to be in the vicinity of the axis of maximum subsidence during the 1964 earthquake—a region which roughly coincides with the crest of the Kenai and Kodiak Mountains (pl. 1). Although the absolute maximum amount of postglacial submergence cannot be determined, an indication of it is provided by differences in the altitudes of cirque floors of probable Wisconsin age along this part of the coast which presumably were formed at a fairly uniform level. The lowest cirque floors along the outer coast of Prince William Sound and in most of the Kodiak Island group—areas away from the region of maximum submergence—lie at altitudes ranging from 800 to 1,000 feet above sea level, but cirque floors along the south coast of the Kenai Peninsula range in altitude from 300 feet below sea level to 800 feet above sea level. This difference in cirque levels suggests at least 300, and perhaps as much as several hundred feet, of submergence in the Kenai Peninsula area.

In contrast to the zone of subsidence, the coast in those areas where the land has risen relative to the sea is generally smoother in outline and commonly exhibits, among other features, one or more wave-cut terraces or uplifted beaches rising to elevations of at least 200 feet (figs. 36, 37). In the major zone of uplift such features are characteristic of the points and capes on the seaward side of Kodiak Island, the islands of the southern and eastern Prince William Sound region, much of the mainland coast east of Cape Suckling, and Middleton Island on the Continental Shelf. Comparable emergent shores with postglacial terraces as much as 1,700 feet high occur all along the mainland coast to the east of the area that was affected by the 1964 earthquake.

Relatively stable or emergent shores occur along parts of the Cook Inlet and Shelikof Strait coasts in areas that either subsided slightly, remained unchanged, or were slightly uplifted during the 1964 earthquake (fig. 33). Many of these emergent shorelines are probably pre-Holocene features related to high eustatic sea levels rather than to tectonic movements (Hopkins and others, 1965, p. 1113; Karlstrom, 1964, p. 34–37).

The record of long-term Holocene deformation within the major zone of uplift in the area between the Copper River Delta and Cape Suckling is seemingly anomalous in that uplifted surfaces described as marine terraces, and drowned forests or sphagnum-peat horizons, occur in close association with one another. Dated marine terraces at Katalla (nos. 26, 27, fig. 34) and Cape Suckling (no. 30, fig. 34) record net Holocene emergence; recent net submergence is indicated by (1) a dated wood sample from the Copper River Delta (no. 36, fig. 34) that was submerged 22 feet in the 1,700 years prior to the earthquake but was uplifted only about 6 feet at the time of the earthquake, (2) a widespread horizon of terrestrial peat of unknown age that was penetrated in borings at depths as much as 40 feet below sea level in the lower Copper River Delta and to a depth of 30 feet at Bering Lake (Tarr and Martin, 1914, p. 462–463), and (3) forest horizons in the Katalla and Cape Suckling areas submerged prior to the earthquake by amounts that were considerably larger than the earthquake-related uplift at these same

36.—Muskeg-covered preearthquake marine terrace on Middleton Island at an altitude of 110–125 feet. It is one of five uplifted terraces on the island, and surf-cut rock platform exposed between the base of the sea cliff and the new high-tide level is a sixth terrace formed by uplift of about 11 feet in 1964. White specks are seagulls. Photograph taken near 7-foot tide stage, April 4, 1964.

37.—The linear tree-covered beach ridge in this view is one of nine elevated beach ridges near Katalla, east of the Copper River Delta. Uplift of about 9 feet in 1964 shifted the shoreline several hundred feet seaward where another beach ridge is in process of formation. Photograph taken near zero tide stage, July 28, 1964.

localities. The relative ages of dated samples in these areas suggest that the long-term displacement may have reversed direction from uplift to submergence during the time interval between about 3,770 and 1,700 years B.P.

Radiocarbon dating of organic material from terraces, peat bogs, and archaeological sites in coastal areas affected by the earthquake has provided some preliminary data on average long-term displacement trends relative to sea level at a number of localities. Displacement-time curves at the five localities (*A–E*) in the area for which multiple samples are available are shown in figure 34. Points on the graph are the preearthquake position relative to mean lower low water plotted against age in millenia of radiocarbon-dated material from these five sites. The available dated samples were collected by several different workers over a period of 30 years and were analyzed in three different laboratories. In spite of the small number of samples at each site, uncertainties in their exact positions relative to sea level, and the ever-present problem of analytical or sampling errors, it is noteworthy that the results appear to be remarkably consistent with one another and with the displacements that occurred in 1964.

These data indicate that the two localities at Yukon Island (*A*) and Girdwood (*B*) have subsided relative to present sea level at an average rate of about 0.7–1.0 foot per century during the time interval from 2,800 to 700 years B.P., whereas those at Middleton Island (*C*) and Katalla (*D*) in the uplifted area have risen at the much greater average rate of at least 3.3 feet per century between 7,650 and 1,350 years B.P. The rate of uplift would be increased somewhat if any of the deduced eustatic sea levels are used in place of the 1964 sea level. The shore at Kukak Bay (*E*), which was not affected by the earthquake, has apparently undergone no detectable net change in its present position relative to sea level since at least 1,450 B.P.

Detailed study of the emergent shorelines shows that the long-term vertical movements occurred, at least in part, as a series of upward pulses separated by intervals of stability or even gradual submergence. Evidence for periodic uplift is exceptionally well displayed on Middleton Island where the surface consists of a flight of five gently sloping and well-developed marine terraces separated by wave-cut cliffs or rises ranging from 20 to 30 feet each (Miller, 1953; Plafker, unpub. data, 1963, 1964, 1965). The highest terrace, forming the flat central part of the island, has a general altitude of 130–165 feet. Sudden uplift of 10–13 feet during the 1964 earthquake, which exposed a sixth surf-cut terrace (shown on fig. 36), suggests that perhaps the earlier terraces may also have been expcsed by a series of upward pulses during previous earthquakes. Radiocarbon-dated driftwood from the highest and lowest of these preearthquake platforms on Middleton Island dates initial uplift of the island at 4,470±250 years ago and uplift of the lowest preearthquake terrace at approximately 1,350±200 years B.P. (nos. 33, 31, table 5). Thus, the extrapolated average time interval between successive uplifts of the three intervening terraces would be close to 800 years.

Because most of the dated samples were taken from deposits laid down when the postglacial eustatic sea level was rising gradually or was at about its present stand, the emergent shorelines probably result almost entirely from tectonic movements. Although some slightly submerged shorelines may be attributed to the effects of a late Holocene rise of sea level and to local surficial subsidence of unconsolidated deposits, in most instances the submergence is so large when compared to deduced eustatic curves that it must also be at least partly due to tectonic subsidence.

Regional submergence of the Kenai and Kodiak Mountains is clearly anomalous in that the movements are in a direction that tends to reduce the elevation of these youthful mountain ranges. Furthermore, because the net load of glacial ice on the mountains has been steadily diminished since the Pleistocene, isostatic adjustments, if any, should be upward. Thus, the submergence is a counter-isostatic process that can only be attributed to regional diastrophic movements.

Less certain is the extent to which the regional postglacial uplift reflects isostatic compensation resulting from unloading of ice and the extent to which it is caused by diastrophic movements. Both processes are undoubtedly involved. However, the known seismic activity of the region, the evidence for pulsating rather than continuous emergence, and the apparent late Holocene reversals in the sense of the net displacements over part of the region suggest that the movements were also, to a large extent, diastrophic in nature.

SHORT-TERM TECTONIC SUBMERGENCE

Much of the coast in the region affected by tectonic movements during the 1964 earthquake had experienced a pronounced submergence for several centuries prior to that event. This phenomenon was probably first recorded by the great English navigator, George Vancouver, who explored Prince William Sound between May 25 and June 17, 1794 (Vancouver, 1801, v. 5, p. 335–336). He noted that on northern Montague Island "The shores are in general low, and as had been already observed, very swampy in many places, on which the sea appears to be making more rapid encroachments than I ever before saw or heard of * * * [trees along the shore] were reduced to naked, dead white stumps, by the encroachment of the sea water to their roots; and some stumps of trees, with their roots still fast in the ground, were also found in no very advanced state of decay nearly as low down as the low water of spring tides." The fact that low water of spring tides is about 13 feet below the normal lower growth limit of trees in this area indicates almost that amount of submergence prior to 1794. The characteristic appearance of these drowned forests in Prince William Sound and along the coast at Cape Suckling east of the Sound is illustrated by figures 38 and 39.

Evidence for active submergence along the coast of the type first described by Vancouver was subsequently noted by geologists, archaeologists, and botanists in the following places: (1) the Controller Bay area east of Prince William Sound (Tarr and Martin, 1914), (2) the Copper River region (Schrader, 1900, p. 404; Reimnitz, 1966, p. 112–125), (3) around Prince William Sound (Grant and Higgins, 1910; Moffitt, 1954; Dachnowski-Stokes, 1941; De Laguna, 1956, (4) around Kachemak Bay on the southern Kenai Peninsula (De Laguna, 1934), (5) in the upper Cook Inlet region (Karlstrom, 1964, p. 48), and (6) in at least two localities on Kodiak Island (Clark, 1966).

Reconnaissance studies by the Geological Survey of all these shorelines after the 1964 earthquake suggest that the submergence observed is a regional phenomenon that has affected much, if not all, of the Prince William Sound region, the mainland coast and islands east of Prince William Sound, the south coast of the Kenai Peninsula at least as far as Kachemak Bay, parts of Turnagain Arm, and segments of the southeastern coast of the Kodiak group of islands. Undoubtedly, many more such localities could be found by detailed examination of the shorelines. Areas within

38.—Bleached trunks of spruce trees on Latouche Island, Prince William Sound, killed by salt-water immersion and partially buried in beach gravel as a result of about 8 feet of submergence below preearthquake extreme high-tide level. The locality was exposed by 8 feet of uplift in 1964. Photograph taken May 28, 1964.

39.—Spruce tree stumps (foreground) rooted in a thin layer of peat on a surf-cut bedrock surface about 14 feet below preearthquake extreme high water (indicated by the top of the line of driftwood below the present forest edge in the background). Radiocarbon age of a stump near the base of the stadia rod was 710±200 years (no. 29, fig. 34). These stumps were exposed by about 16 feet of uplift in 1964. Photograph taken July 24, 1964.

which there is evidence of preearthquake submergence are indicated on figure 33 and the ages and positions relative to sea level of the dated samples are shown on figure 34. In a few scattered localities within Prince William Sound, evidence of stable shorelines or of possible recent slight uplift (no more than 4 ft) was found. This evidence suggests that small areas may have acted as independent tectonic units that did not take part in the general submergence.

The available data is insufficient for determining whether the submergence, which affected the offshore islands from Kayak Island to the Copper River Delta, extended as far out on the Continental Shelf as Middleton Island. Preearthquake sea levels that reached to the base of the prominent sea cliff encircling much of the island indicate either a long period of relative stability and erosion or some submergence since the last uplift of the island roughly 1,350 years ago.

The record of preearthquake submergence in the zone that was lowered in relation to sea level is much less complete than that for the zone that was raised, mainly because much of the evidence is now below lowest low tide. Available data suggest that submergence occurred along parts of the coast of the Kenai Peninsula and the Kodiak group of islands but that submergence probably was much less than in the area from Prince William Sound to Cape Suckling. No change in mean sea level attributable to tectonic movements was detected in a 21-year tidal record at Seward and a 15-year record at Kodiak. An apparently large preearthquake submergence on the delta of Spruce Creek near Seward of about 6.9 feet in less than 200 years, as indicated by drowned rooted tree stumps (no. 12, fig. 34), may be partly due to slumping and (or) compaction of deltaic deposits.

Radiocarbon dates from drowned terrestrial plants that were probably killed by sea-water encroachment provide data that permit a crude estimate of the duration and average rate of the short-term preearthquake submergence. Ages in radiocarbon years versus amount of submergence of samples taken from the most deeply submerged terrestrial vegetation at each site are plotted in the stippled area on figure 34. The fairly regular linear distribution of 14 out of 16 samples, all from the area between Seward and Cape Suckling, suggests a general submergence at a rate that averaged roughly 1.7 feet per century for about 930 years. On the other hand, positions of two samples (nos. 15, 21, fig. 34) from this same general area, both of which are older than 930 years, reflect lower rates of submergence, as do three samples from the Cook Inlet region (nos. 8, 11, 15, fig. 34). The maximum indicated average rate of submergence is 8.6 feet in 230 years or about 3.7 feet per century on Latouche Island in Prince William Sound.

Absence of historic records of sudden earthquake-related relative sea-level changes and of geomorphic evidence for such movements suggest that the submergence was probably gradual or that it occurred in numerous small increments over a long period of time. This time interval, as inferred from dated submerged shorelines, was at least 930 years in the Prince William Sound region. The upper limit for the duration of the submergence is far less certain; if it corresponds to the oldest dated submerged-forest sample along the coast (no. 21, fig. 34) and the time of uplift of the lowest preearthquake marine terrace on Middleton Island (no. 31, fig. 34), it could be as much as 1,350 years.

Short-term submergence occurred at a time when sea level was at or near its present stand and in a region where overall isostatic displacements in response to glacial unloading should have been upward. Isostatic adjustments between major earthquakes such as the 1964 event could conceivably be responsible for some of the submergence in areas of tectonic uplift. However, such adjustments are inadequate to explain either short-term submergence in the Copper River Delta-Cape Suckling area that exceeds earthquake-related uplift or submergence in areas of the Kenai Peninsula and Kodiak Island that subsided both before and during the earthquake. By implication, this fact suggests that the submergence was caused at least partly by diastrophism that involved a significant downward-directed component of regional strain of variable amount and rate over much of the coastal belt affected by vertical displacements during the 1964 earthquake.

TECTONIC IMPLICATIONS OF THE RECORD OF HOLOCENE VERTICAL MOVEMENTS

The history of Holocene vertical movements in the region affected by the 1964 earthquake, and in nearby areas, is a fragmentary one based largely on a rapid reconnaissance after the earthquake and on incidental investigations by others. Additional work—involving far more radiocarbon dating—is required to obtain detailed data on vertical movements at selected localities in areas of net Holocene emergence and submergence. However, the following four tentative conclusions regarding the prehistoric tectonic movements seem to be indicated by the available data:

1. Areas of net Holocene emergence or submergence broadly correspond with those areas in which significant amounts of uplift and subsidence occurred during the 1964 earthquake. Thus, the tectonic movements that accompanied the earthquake were apparently but one pulse in a long-continuing trend of deformation. This trend has resulted in regional emergence of parts of the continental margin, simultaneous submergence of the Kenai-Kodiak Mountains belt, and either relative stability or emergence along the shores of Cook Inlet and parts of Shelikof Strait.
2. The amounts of net long-term Holocene emergence and submergence of the coast, which are locally considerably larger than the postulated eustatic sea-level changes for the same time interval, indicate that differential displacement of the shoreline results largely from tectonic movements. Progressive submergence of youthful mountains recently unloaded of ice and pulsating emergence of the Continental Shelf and Gulf of Alaska coast are suggestive of a dominant regional diastrophic deformation.
3. Steplike flights of Holocene marine surf-cut terraces at a number of localities along the coast suggest that the long-term vertical movements occurred as a series of upward pulses that were separated by intervals of stability or even gradual subsidence that locally average 800 years or more in duration. These upward pulses probably represent earthquake-related movements comparable in origin to those which affected parts of the same coast in 1964.

4. Gradual tectonic submergence prevailed during at least the past 900 years, and perhaps as long as 1,360 years, over much of the zone that was uplifted and over at least part of the zone that subsided during the 1964 earthquake. This widespread submergence is tentatively interpreted as direct evidence for a significant downward-directed component of regional strain preceding the earthquake and its duration as the approximate time interval since the last major tectonic earthquake in this same region.

MECHANISM OF THE EARTHQUAKE

GENERAL CONSIDERATIONS

According to the classic elastic rebound theory of earthquake generation (Reid, 1911), which is generally accepted by western geologists and geophysicists, shallow earthquakes are generated by sudden fracture or faulting following a period of slow deformation during which energy is stored in the form of elastic strain within rock adjacent to the fault. When the strength of the rock is exceeded, failure in the form of faulting occurs, the material on opposite sides of the fault tends to rebound into a configuration of elastic equilibrium, and elastic strain potential is released in the form of heat, crushing of rock, and seismic-wave radiation. The drop in elastic strain potential is possibly augmented or partially absorbed by net changes in gravitational potential associated with vertical tectonic displacements.

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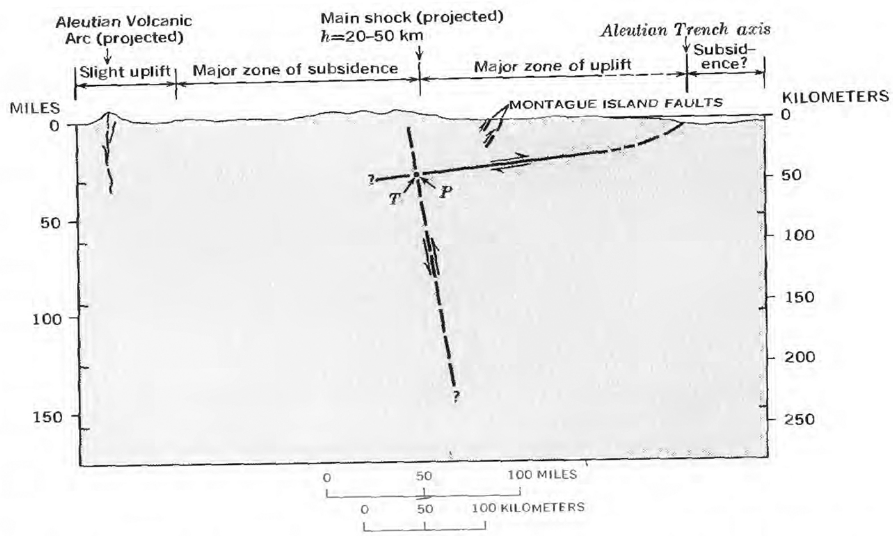


40.—Diagrammatic section showing two alternative interpretations of the 1964 Alaska earthquake based on focal mechanism studies (S. T. Algermissen, written commun., March 1965) and analyses of residual vertical displacements by dislocation theory (Savage and Hastie, 1966). Section is oriented through the focus of the main shock and roughly normal to the regional structural trend. P and T axes are the approximate orientations of the principal pressure and tension axes, respectively, at the hypocenter of the main shock indicated by the focal mechanism studies.

Field investigations demonstrate that there is little likelihood that the primary fault along which the 1964 earthquake occurred is exposed at the surface on land, nor is there evidence for movement on any of the known major continental faults. If the earthquake originated by rupture along one or more faults, the two most plausible models for the orientation and sense of movement on the primary fault consistent with the available fault-plane solutions and dislocation-theory analyses of the residual vertical displacements are (1) relative seaward thrusting along a fault that dips northwestward beneath the continental margin at a low angle, and (2) dip-slip movement on a near-vertical fault, the ocean side being relatively upthrown, that strikes approximately along the zero isobase between the major zones of uplift and subsidence (fig. 40). These two models, which are discussed in the following sections, are referred to as the "thrust-fault model" and the "steep-fault model."

Whether the postulated shearing resulted from the overcoming of frictional resistance to sliding, as set forth in the elastic rebound theory of Reid (1911), or from some other process such as brittle fracture, creep instability, or propagation of flaws cannot be ascertained from available data. Several writers (Orowan, 1960; Griggs and Handin, 1960; Evison, 1963, p. 863–884) have pointed out deficiencies in the elastic rebound mechanism for earthquakes at depths of more than a few miles, where frictional stress on dry fault planes must vastly exceed the rock strength. However, such considerations generally do not take into account the effect of pore fluids which must exist in the real earth down to the depths of the deepest hydrated mineral phases. Recent laboratory studies (Raleigh and Patterson, 1965, p. 3977–3978; Griggs and Blacic, 1965; Raleigh, 1967) suggest the possibility that the Reid mechanism may extend to depths at least as great as the lower crust and upper mantle in regions where frictional resistance to sliding on faults may be reduced by local anomalously high pore pressures or where the strength of the rock is lowered sufficiently in the presence of pore fluids to permit brittle fracture.

An alternative mechanism to be considered is one that explains tectonic earthquakes as originating from sudden expansion and (or) contraction of large volumes of rock due to rapid phase changes (Evison, 1963). According to this concept, faulting is the result rather than the cause of earthquakes. No evidence has yet been found from either natural or experimental petrologic systems to support the idea that reconstructive solid-solid or solid-liquid reactions can occur fast enough throughout sufficiently large volumes of rock to generate major earthquakes or to produce regional vertical displacements (Ghent, 1965). Aside from the theoretical objections, it is difficult to conceive of a reasonable combination of equidimensional volume changes that could cause the observed pattern of vertical and horizontal displacements that accompanied this earthquake. For these reasons, there is little likelihood that phase changes were a primary mechanism for the 1964 earthquake, and this possibility will not be pursued further.

THRUST-FAULT MODEL

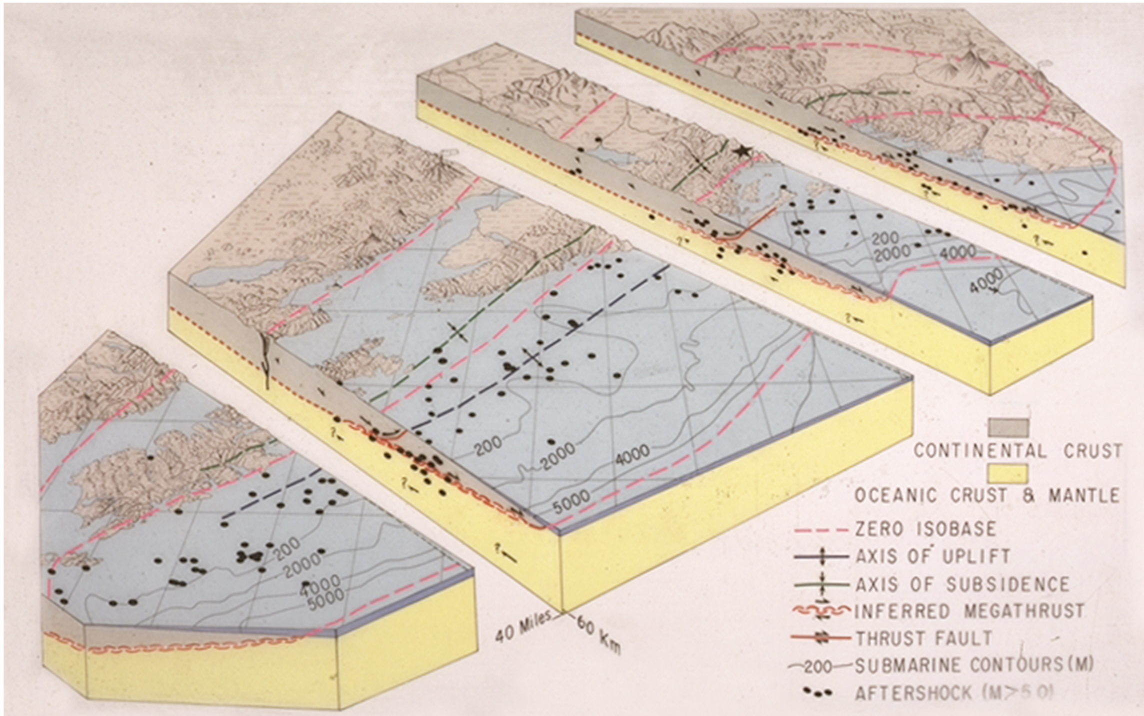
Most of the preliminary data available by the end of the 1964 field season suggested that the earthquake and the associated tectonic deformation resulted from a relative seaward thrusting along the continental margin that was accompanied by elastic horizontal extension behind the thrust block (Plafker, 1965, p. 1686). The thrust-fault model has been strongly reinforced by data subsequently obtained from (1) more detailed fieldwork on the surface displacements and preearthquake movements by the Geological Survey during 1965, (2) comparison of pre- and post-earthquake triangulation surveys for a large segment of the deformed region (Parkin, 1966, fig. 4), and (3) detailed focal-mechanism studies of the main shock and larger aftershocks (Stauder and Bollinger, 1966). The essential features of the suggested model, which are illustrated diagrammatically by figure 41, are presented below.

OUTLINE OF THE MODEL

According to the thrust-fault model, the earthquake resulted from shear failure along an inferred major zone of movement, or megathrust, which dips northwestward beneath the continental margin from the vicinity of the Aleutian Trench, and on subsidiary faults within the overthrust plate. The zone within which movement occurred, as delineated by the belt of major aftershocks, parallels the Aleutian Trench on the northwest for a distance of some 600 miles and is from 110 to 180 miles wide (fig. 2). It lies almost entirely within the major zone of uplift and the portion of the adjacent zone of subsidence that is on the seaward side of the axis of subsidence.

The pattern of earthquake-related horizontal displacements (fig. 16) and the evidence for regional preearthquake shoreline submergence along the continental margin (fig. 33) suggest that shear failure followed a long period of elastic strain accumulation (930–1,360 years) during which the upper plate was horizontally compressed roughly normal to the trend of the Aleutian Arc and probably simultaneously depressed relative to sea level. A hypothetical deformation cycle is illustrated diagrammatically in figure 42. The data suggest that the dynamic drive was probably provided by underthrust of the lower plate, with a downward-directed component of movement. R. K. Hose has pointed out that distortional drag due to an underthrust would have generated a shear couple in which the relative strains and orientation of potential shear planes coincide reasonably well with those deduced from fault-plane solutions for the main shock and many or the larger aftershocks (oral commun., Nov. 1965). Alternative mechanisms such as movement of the continental margin over the oceanic basin, or almost horizontal movements, could not be responsible for the combination of both horizontal shortening and regional submergence indicated by the available data.

Previous seismicity in the region during the strain-accumulation phase of the 1964 earthquake may have resulted from adjustments within the strained volume of rock involving predominantly lateral offsets or relatively local dip-slip displacement. This possibility is suggested by the fact that, although prior historic earthquakes with magnitudes of 7 or more have been recorded



Fault, showing sense and relative amount of displacement
Queried where inferred

Inferred megathrust zone

M-discontinuity (after Shor, 1962, fig. 4)

Relative direction and amount of tectonic displacement
Queried where inferred

Epicenter of main shock

Epicenter of larger aftershock (M>5.0) and projected hypocenter of aftershock within 25 miles of vertical section

Active or dormant volcano

41.—Schematic block diagram showing the postulated relationship of the seismic activity and tectonic displacement associated with the 1964 Alaska earthquake to major structural elements of south-central Alaska. Horizontal and vertical scales equal below sea level; vertical scale above sea level exaggerated × 4. Drawn by H. T. Koelbl.

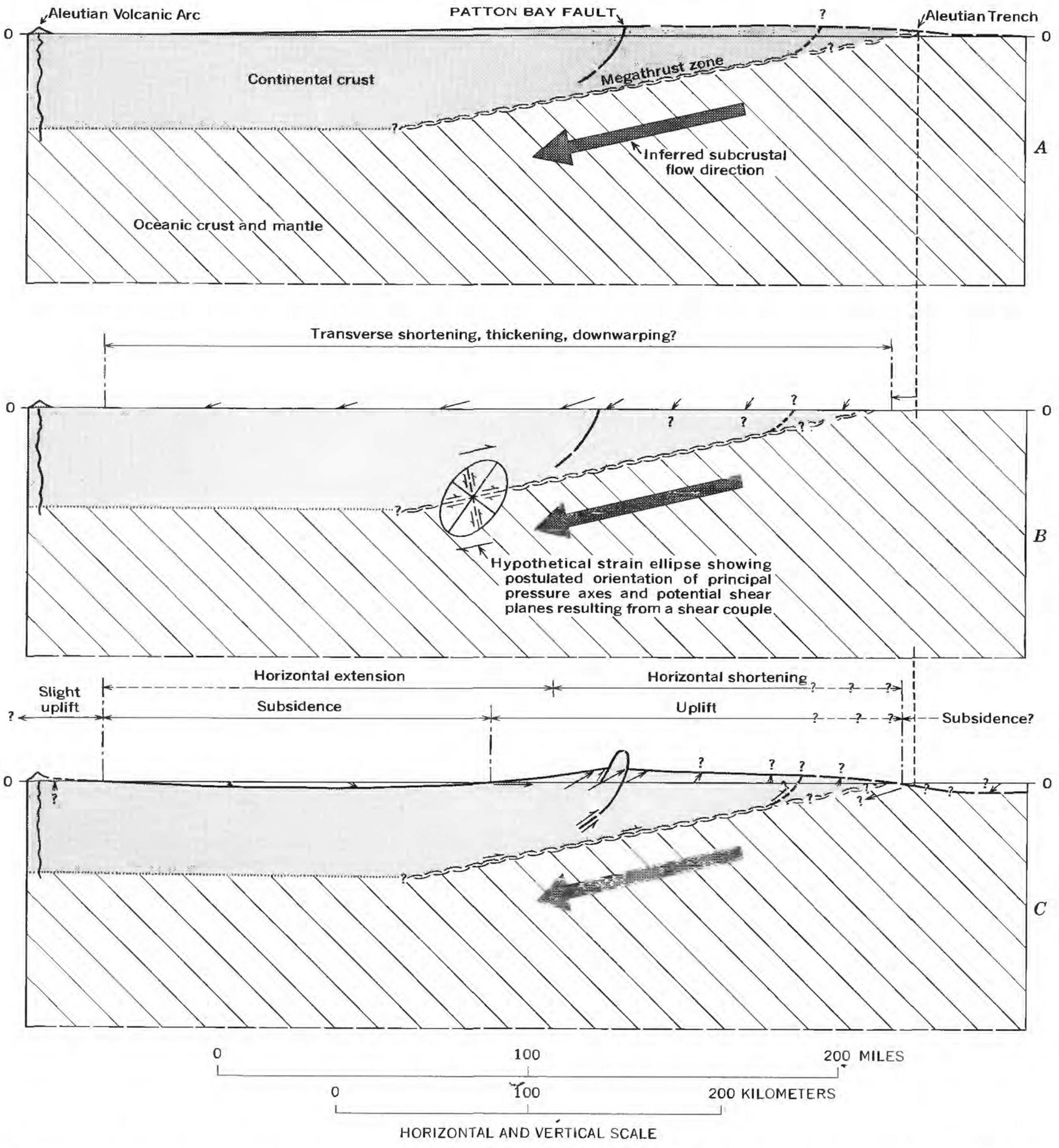
(Wood, 1966, p. 24), none of these were accompanied by known regional tectonic deformation—and certainly none with deformation on the scale of the 1964 earthquake.

Faulting at the time of the earthquake presumably was initiated at the hypocenter of the main shock in northern Prince William Sound, from which point it propagated simultaneously updip towards the Aleutian Trench and along strike both towards the southwest and east within the area encompassed by the aftershocks. Shear failure was accompanied by elastic rebound in the upper plate above the thrust, which resulted in (1) relative seaward displacement and uplift of a part of the continental margin by movement along the inferred megathrust and the subsidiary reverse faults that break through the upper plate to the surface, and (2) simultaneous elastic horizontal extension and vertical attenuation (subsidence) of the crustal slab behind the upper plate. These movements, possibly in combination with unidentified submarine faulting and (or) underthrusting of the lower plate in the opposite direction, resulted in the observed and inferred tectonic displacements at the surface.

Indicated stress drops at the surface across the zone of subsidence (on the order of a few hundred bars) are comparable in magnitude to those reported for other tectonic earthquakes. For the idealized case of homogeneous strain and purely elastic distortion of the crust, the stress drop (P) is a function of Young's modulus (E) of the slab and the reduction of horizontal strain (ϵ) across it:

$$\triangle P = E\epsilon$$

For an average E of 3×10^5 bars



42.—Diagrammatic time-sequential cross sections through the crust and upper mantle in the northern part of the region affected by the 1964 earthquake. *A*, Relatively unstrained condition after the last major earthquake. *B*, Strain buildup stage during which the continental margin is shortened and downwarped. *C*, Observed and inferred displacements at time of the earthquake during which a segment of the continental margin is thrust seaward relative to the continent. Datum is the upper surface of the crust beneath the cover of water and low-velocity sediments. Vertical displacements at the surface, which are indicated by the profiles and by arrows showing sense and relative amount of movement, are about × 1,000 scale of the figure.

(7×10^{11} dynes cm^{-2}), and average and maximum strain across the subsided zone of 1.1×10^{-4} and 3×10^{-4}, respectively, the average stress drop was only about 77 bars (77×10^{6} dynes cm^{-2}). The maximum drop, which occurs near the axis of subsidence, was about 210 bars (210×10^{6} dynes cm^{-2}).

The megathrust in the earthquake focal region dips north to northwest at an angle of 9°–15°, according to fault-plane solutions of the main shock and several aftershocks, but its precise depth and configuration are uncertain. The limited available data on focal depths of the earthquake (p. I 4) and on deep crustal structure (Shor, 1962, fig. 4; Hales and Asada, 1966, p. 420) suggest that the megathrust may coincide with the unstable interface at the base of the continental crust beneath the continental slope and shelf. It probably extends into the upper mantle toward the north to some unknown depth (no deeper than the deepest hypocenters in the region) where stress relaxation is sufficiently rapid to absorb the applied force by plastic, rather than elastic, deformation.

MAJOR UNRESOLVED PROBLEMS

The cause of the slight uplift that occurred to the north of the major zone of subsidence is uncertain. Presumably, this uplift involves an elastic distortion related to strain changes resulting from seaward thrusting along the continental margin at the time of the earthquake. It could represent an increase in vertical strain (Poisson bulge) resulting from a sudden increment in the regional horizontal compressive strain oriented roughly normal to the zone (R. O. Burford, written commun., 1966). If this explanation for the uplift is correct, retriangulation within the areas of slight uplift should show transverse horizontal shortening. A test of the suggestion could be made if the postearthquake triangulation net were extended northward into this zone.

A second major problem is that there exists an apparent asymmetry in the volumes of uplift (89 cu mi, 372 cu km) and subsidence (29 cu mi, 122 cu km) in the two major zones that implies a substantial increase in net gravitational potential. By making the assumptions that the displacements decrease linearly from a maximum at the surface to zero at the base of the continental crust, that there were no density changes within the affected volume of crust, and that the crust has an average density of 2.7, the approximate upper limit for this change may be calculated. An average lowering of 2½ feet (75 cm) in a crustal block having the dimensions of the zone of subsidence would result in a potential energy loss of 2.7×10^{19} ft-lbs (3.6×10^{26} ergs). The increase in gravitational potential in the uplifted area would be about 4.1×10^{19} ft-lbs (5.6×10^{26} ergs), assuming an average uplift against gravity of roughly 6 feet (1.8 m) for a wedge-shaped block with a length of 475 miles (100 km), a width of 115 miles (185 km), and an average thickness of about 11 miles (18 km).

These data suggest an apparent net increase in gravitational potential of about 1.5×10^{19} ft-lbs (2×10^{26} ergs). Even with the assumption that vertical displacements extend only halfway through the slab, the indicated gravitational potential increase is still roughly two orders of magnitude larger than the total released seismic-wave energy of about $1\text{–}2 \times 10^{24}$ ergs (calculated from the empirically derived Gutenberg-Richter relationship between energy and earthquake magnitude $\log E = 11.4 + 1.5\,M$; Richter, 1958, p. 366).

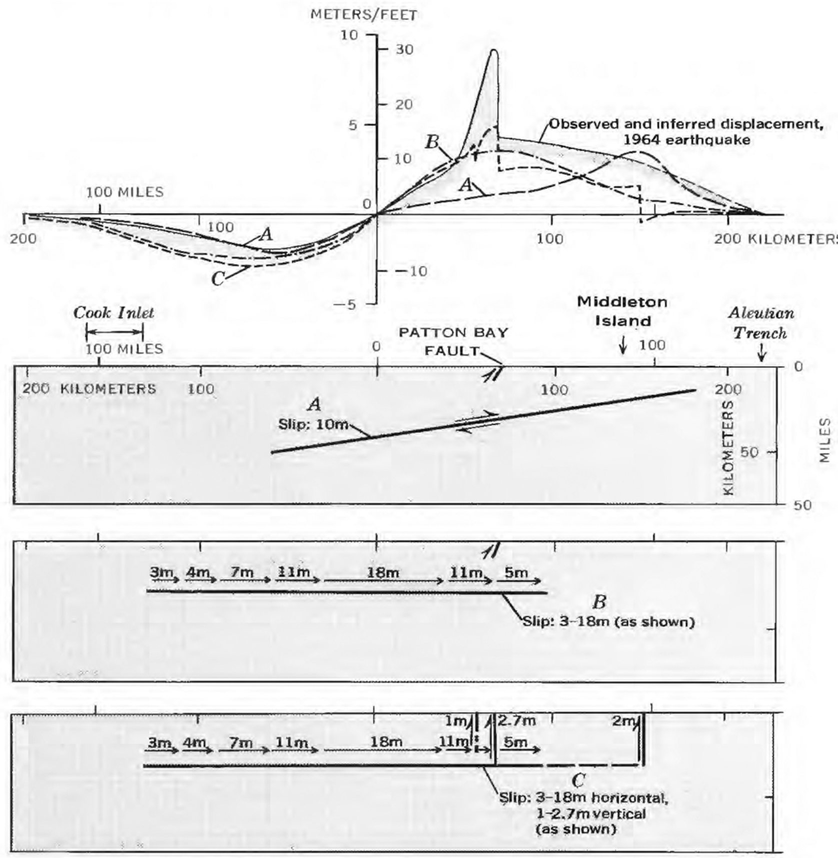
In fact, however, the net change in gravitational potential due to elastic-rebound mass redistribution in the crust must be considered as indeterminate because data are unavailable on (1) the seaward and continentward limits of the displacement field, (2) possible elastic density changes attendant upon stress drops within and behind the upper plate, and (3) the extent to which the movements may have been compensated by elastic depression of the denser mantle beneath zones of uplift and a corresponding rise beneath the zone of subsidence.

REPRESENTATION BY DISLOCATION THEORY

Savage and Hastie (1966) and Stauder and Bollinger (1966) have used dislocation theory to compare theoretical profiles of vertical surface displacement for various gently dipping and horizontal overthrust models to the observed and inferred profile. Their assumed models, with the corresponding profiles along a vertical plane oriented normal to the fault strike, are plotted to a common scale on figure 43. These models have, of necessity, been highly generalized to permit a mathematical analysis by dislocation theory, and, in part, the studies are based upon preliminary and incomplete observational data.

The basic assumption in this procedure is that the observed displacements can be modeled by the corresponding fields of a planar dislocation sheet in a homogeneous, isotropic, elastic half-space with displacement discontinuity matching the observed or inferred fault slip. Inasmuch as the depth, configuration, and net slip on the postulated faults are to a large degree speculative, considerable latitude exists in the parameters used for those calculations. Furthermore, because faults that break to the surface are not dislocations in a semi-infinite medium, their contribution to the deformation can only be examined qualitatively.

Figure 43 shows that, although each of the assumed models can approximately account for the observed subsidence, none of them gives a close fit between the theoretical and actual profiles in the uplifted zone. Stauder and Bollinger (1966, p. 5293) suggest that the observed tectonic surface displacements can be approximated to any desired degree by assuming combined differential-slip motion and a shallowly dipping thrust plane. Such a model would be more nearly in accord with the data which suggest a dipping master fault having the approximate configuration shown in figure 43*A* with differential slip as indicated in figure 43*B*. To be realistic, however, it would also have to include the effects of (1) imbrication along known and suspected thrust or reverse faults that break to the surface, (2) possible breaking to the surface along the shallowing leading edge of the megathrust, and, perhaps (3) possible simultaneous underthrusting of the lower plate along an inclined fault plane.



43.—Comparison of observed and inferred profile of surface displacement along line *B–B'* of plate 1 (shaded line) with computed profiles for movement on faults of slip and cross-sectional configuration shown in diagrams below the profiles. Model *A* from Savage and Hastie (1966, fig. 3); models *B* and *C* from Stauder and Bollinger (1966, fig. 7).

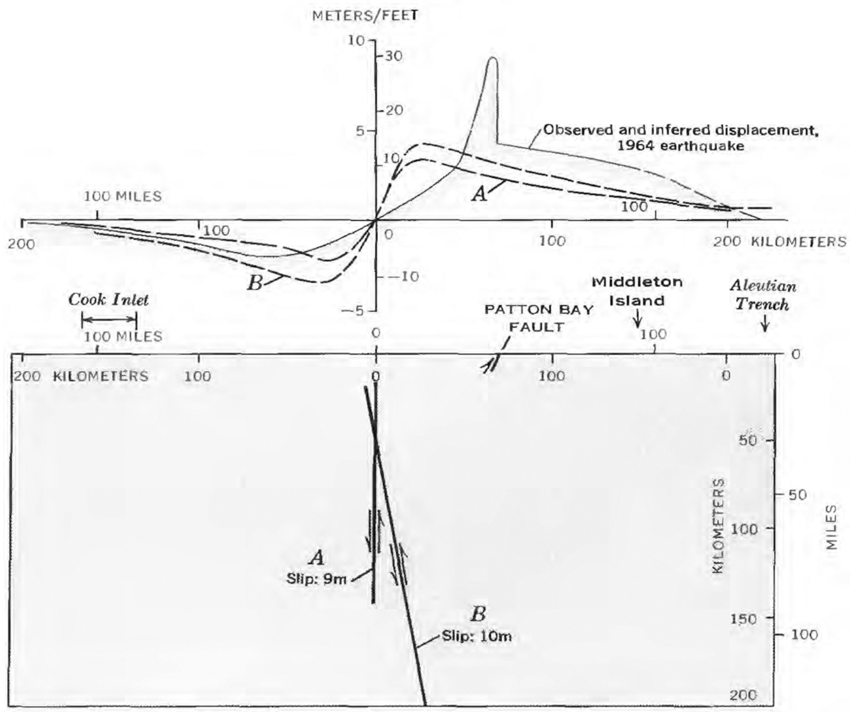
STEEP-FAULT MODEL

OUTLINE OF THE MODEL

According to the steep-fault model, the earthquake resulted from elastic rebound on a near-vertical fault that strikes approximately along the line of zero change in land level, with the southeast block up and the northwest block down relative to sea level. A steeply dipping fault was suggested by a preliminary fault-plane solution based on surface waves (Press and Jackson, 1965, p. 867; Press, 1965, p. 2404). However, Savage and Hastie (1966, p. 4899–4900) have pointed out that a unique surface-wave solution for the fault orientation cannot be inferred from the surface waves, inasmuch as the direction of rupture propagation is along the null axis and therefore the radiation of seismic waves will be essentially the same for either of the two possible fault planes.

The main appeal of the steep-fault model is that it can readily account for the gross distribution of uplift and subsidence in two major zones, as well as the occurrence of the earthquake epicenter close to the zero isobase between these zones.

However, as noted elsewhere (Plafker, 1965, p. 1686), there are no compelling geologic, seismologic, or geodetic data in support of the steep-fault model. Aftershocks are not grouped along its postulated trace, but instead lie mainly in a broad belt along the continental margin mainly to the south of the zero isobase (fig. 2). Furthermore no field evidence exists for new surface breakage in the vicinity of the zero isobase—despite the fact that it intersects the coast in more than 15 localities (fig. 3). And finally there is no evidence that the line corresponds to a major geologic boundary with the seaward side relatively upthrown, as might be expected if it marked the trace of a major fault



44.—Comparison of observed and inferred profile of surface displacement along line *B–B'* of plate 1 (shaded line) with computed profiles for movement on steeply dipping faults of slip and cross-sectional configuration shown in the diagram below the profiles. Model *A* from Press and Jackson (1965, fig. 1); model *B* from Savage and Hastie (1966, fig. 3).

along which vertical movement has occurred in the past. On the contrary the overwhelming majority of surface faults that parallel the coast in this part of Alaska have exactly the opposite sense of displacement (fig. 29).

The apparent absence of a surface dislocation along the zero isobase between the major zones prompted the suggestion that the displacement represents flexure above a near-vertical fault at depth (Press and Jackson, 1965, p. 868; Press, 1965, p. 2405). Such a fault would have to extend to the considerable depth of 62–124 miles (100–200 km) below the free surface to account for the areal distribution of residual vertical displacements normal to its inferred strike.

REPRESENTATION BY DISLOCATION THEORY

Observed residual tectonic displacements have been compared to the theoretical surface displacements that would occur on faults of varying inclination, slip, and dimension by application of dislocation theory (Press and Jackson, 1965; Press, 1965; Savage and Hastie, 1966). These analyses are subject to the same basic assumptions as were previously outlined. Assumed models, with corresponding profiles along a vertical plane oriented normal to the fault strike, are plotted to a common scale on figure 44. The resultant profiles show the same general spatial distribution of uplift and subsidence as the profile of observed and inferred vertical displacements along a northwest-southeast line through the southwest tip of Montague Island. They differ fundamentally, however, in that the axes of uplift and subsidence are too close together by at least 50 miles (80 km), or a factor of one-half, and the indicated changes within the two major zones are notably more equal in amplitude than in the observed profile. The distance between the two axes would tend to decrease even further if the faulting is shallow—as is suggested by the spatial distribution of the aftershocks. Even if the vertical displacements could be explained by these models, no reasonable combination of fault dimensions, slip, and dip could also duplicate the systematic horizontal displacements observed in the field (fig. 18).

SUMMARY AND CONCLUSIONS

The earthquake of March 27, 1964, was accompanied by regional vertical and horizontal displacements over an area probably in excess of 110,000 square miles in south-central Alaska. Major deformation and related seismic activity were largely limited to an elongate segment of the continental margin lying between the Aleutian Trench and the chain of late Cenozoic volcanoes that comprise the Aleutian Volcanic Arc.

Geologic evidence in the earthquake-affected region suggests that the earthquake was but the most recent pulse in an episode of deformation that probably began in late Pliocene time and has continued intermittently to the present. The net effect of these movements has been uplift along the Gulf of Alaska coast and continental margin by warping and imbricate faulting; subsidence or relative stability characterized most of the adjacent landward zone extending inland approximately to the Aleutian Volcanic Arc. The length of time since the last major tectonic earthquake that involved regional warping, as inferred from radiocarbon-dated displaced shorelines in the epicentral region, appears to have been about 930–1,360 years.

Because the primary fault or zone of faulting along which the earthquake is presumed to have occurred is not exposed at the surface on land, a unique solution for its orientation and sense of slip cannot be made. Nevertheless, the vertical displacements, when considered in relationship to the focal-mechanism studies and spatial distribution of seismicity, strongly favor the hypothesis that the primary fault was a major thrust fault, or megathrust. The data suggest that the segment of the megathrust along which slippage occurred was 550–600 miles long and 110–180 miles wide and that it dips from the vicinity of the Aleutian Trench at a gentle angle beneath the continental margin.

According to the thrust-fault hypothesis, the observed regional uplift and transverse shortening along the continental margin during the earthquake resulted from (1) relative seaward displacement of the upper plate along the dipping primary thrust, (2) imbrication on the known subsidiary reverse faults that break through the upper plate to the surface, and (3) crustal warping. Movement on other subsidiary submarine reverse faults, as well as rebound of the lower plate toward the continent, may have contributed to the uplift. Simultaneous subsidence in the zone behind the fault block presumably reflects an elastic horizontal extension and vertical attenuation of previously compressed crustal material. Stored elastic-strain energy within the thrust block and the segment of crust behind the block that was affected by subsidence was the primary source of energy dissipated during the earthquake in the form of seismic waves, heat, crushing of rock, and, perhaps, net changes in gravitational potential. Major unresolved problems are the cause of the slight uplift in the area north of the two major zones of deformation and the apparent large increase in potential energy over the displacement field.

The implication of a genetic relationship between the 1964 Alaska earthquake and the Aleutian Arc appears inescapable in view of the nature of the surface deformation and seismic activity associated with the earthquake. Available geologic, geodetic, and geophysical data from the region affected by tectonic deformation during the earthquake are compatible with the concept that arc structures are sites of down-welling mantle convection currents and that planar seismic zones dipping beneath them mark the zone of shearing produced by downward-moving material thrust against a less mobile block of the crust and upper mantle. Conversely, the data provide severe constraints for alternative hypotheses that relate deformation in the eastern Aleutian Arc, at least, primarily to transverse regional extension or to movement on major longitudinal strike-slip faults or shallow steep dip-slip faults.

Preearthquake strain directed at a gentle angle downward beneath the arc in the epicentral region is suggested by geodetic evidence for horizontal shortening of as much as 64 feet roughly normal to the Gulf of Alaska coast and by geologic evidence for progressive coastal submergence in the same region in excess of 16 feet. If the duration of preearthquake strain accumulation, as inferred from radiocarbon-dated downed shorelines, was about 930–1,360 years, average annual horizontal displacement was about 0.59–0.83 inch per year (1.5 to 2.1 cm per year). The indicated shortening, which is probably a minimum value inasmuch as there probably also was some permanent shortening, is reasonably compatible with sea-floor spreading rates of about 2.9 cm per year in a northwesterly direction from the Juan de Fuca ridge. Such rates and spreading directions have been deduced from paleomagnetic studies of the northeastern Pacific ocean floor (Vine, 1966, p. 1407). Alternative driving mechanisms, in which the upper plate overrides the ocean basin or in which the regional strain is directed horizontally, do not readily account either for the widespread preearthquake subsidence relative to sea level that accompanied horizontal shortening in coastal areas affected by earthquake-related tectonic deformation or for the orientation of principal stress axes at the hypocenter of the main shock and many of its aftershocks.

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